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The Composition and Petrogenesis of the Lower Crust: A Xenolith Study

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Granulite facies lower crustal xenoliths from a single basaltic vent (Hill 32) in the McBride volcanic province, north Queensland, Australia, illustrate the extreme lithologic diversity of the deep crust. These xenoliths are dominantly mafic, but intermediate and felsic granulites represent a significant proportion (~20%) of the xenolith population. All xenoliths have high-grade mineralogies and generally well-equilibrated textures, as well as decompression features which are indicative of derivation from deep crustal levels (0.7-1.0 GPa). Major and trace element chemistry for 12 xenoliths, chosen to span the observed lithologic range, are used to constrain the petrogenesis of these rocks. In the smaller, layered samples, metamorphic differentiation may lead to nonrepresentative sampling. However, in most instances such secondary processes can be identified and the original chemical characteristics delineated. The mafic xenoliths formed through a variety of processes, including crystallization of mafic magma, crystal accumulation from mafic and felsic magmas, and partial melting of intermediate compositions to yield a mafic residuum. The two analyzed intermediate xenoliths are metasediments, based upon high alumina to alkali ratios and rare earth element patterns, whereas the two felsic xenoliths have compositions similar to igneous rocks. Comparisons of noncumulate and/or restite xenoliths with unmetamorphosed rock types show that K and Rb are variably depleted whereas Th and U are strongly depleted, in all rock types. The large ion lithophile element depletion patterns for xenoliths are similar to those of rocks from granulite facies terrains, suggesting similar processes were operative. Suggestions of an anomalously Ba- and Sr-rich lower crust are not supported by the data. By using the observed lithologic proportions of xenoliths at Hill 32, and the chemical analyses presented here, a weighted mean composition of the lower crust can be obtained and compared with recently proposed lower crustal compositions. The weighted mean composition is mafic, and formed through a combination of basaltic underplating, crystal accumulation, tectonic underplating, and partial melting. No simple model of lower crust formation through basaltic or andesitic underplating nor intracrustal melting is sufficient to explain lower crust formation in Phanerozoic continental margin settings.

INTRODUCTION

The lower continental crust is a relatively inaccessible region of the earth, making estimates of its bulk composition difficult. This, in turn, places large uncertainties on estimates of bulk crust composition and, consequently, volumetric estimates of depleted mantle, the presumed depleted complement to the enriched crust.

Unlike the upper crust, where sediments provide a good means whereby average composition can be estimated, lower crustal samples are limited to two types: (1) granulite facies terrains and (2) xenoliths from rapidly erupted basalts or kimberlites.

Granulite terrains provide good exposures of rocks which have resided for some length of time at high pressures and temperatures. Study of these terrains allows delineation of contact relationships, volumetric lithological proportions, and the effects of high-grade metamorphism on a variety of rock types. However, if granulite terrains are the direct consequence of double thickening of the crust, caused by Himalayan-type continent-continent collisions [Newton and Perkins, 1982], then they cannot be regarded as representative of the lower crust. They would instead represent the midcrustal region

of a very thick crust. Recent descriptions of pressure-temperature paths for granulites suggest several terrains have undergone substantial isobaric cooling prior to uplift (Enderby Land granulites [Ellis, 1980; Harley, 1985] and Adirondacks [Bohlen *et al.*, 1985]). Such paths, if they can be shown to occur over a long time period (i.e., >100 m.y.), indicate that these granulites may not have formed through double thickening. These terrains may thus be more representative of the lower crust, but they still require a major tectonic event to expose them.

Samples of present-day lower crust are restricted to xenoliths. Decompression features and high-pressure mineralogies indicate these rocks are samples of lower crust which were rapidly transported to the earth's surface. Consequently, studies of xenoliths, though of limited use in defining contact relationships, and perhaps absolute lithologic proportions, are very important for determining the bulk composition of the lower crust and the processes responsible for its formation and modification. However, despite the need for good geochemical data on these rock types, very few exist (published major and trace element data are currently available from only eight localities: Massif Central, France [Leyreloup *et al.*, 1977], Kilbourne Hole, New Mexico [Padovani and Carter, 1977a; Wandless and Padovani, 1985], Lesotho, southern Africa [Griffin *et al.*, 1979; Rogers and Hawkesworth, 1982], Eifel, West Germany [Stosch *et al.*, 1986], Hoggar, Algeria [Leyreloup *et al.*, 1982], Boomi Creek, New South Wales, Australia [Wilkinson, 1975; Wilkinson and Taylor, 1980], Calcutteroo, South Australia [Arculus *et al.*, 1987] and Chudleigh province, north Queensland, Australia [Kay, 1983; Rudnick *et al.*, 1986].

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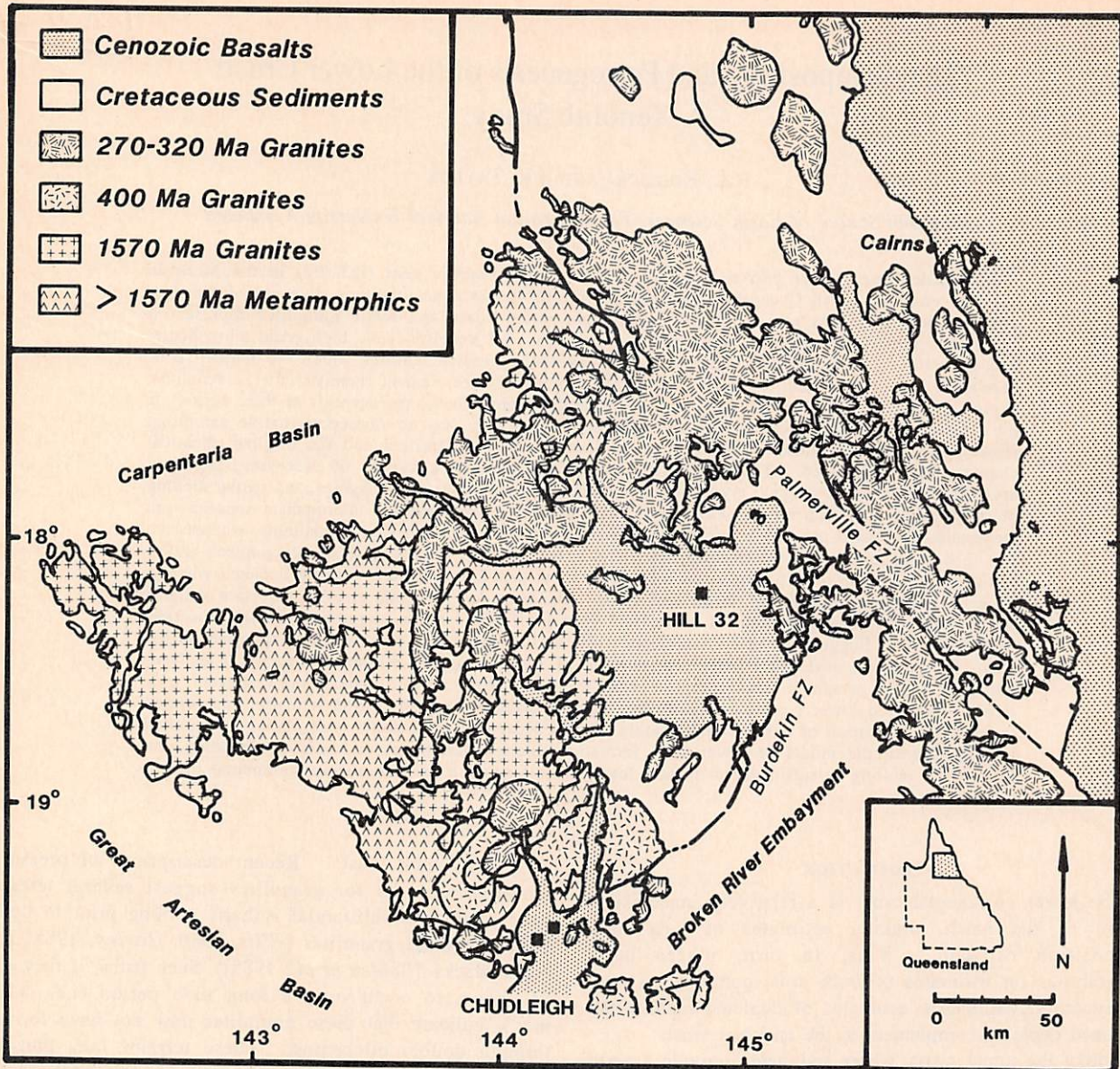


Fig. 1. Location of northern Queensland volcanic provinces and xenolith-bearing cinder cones (squares). Hill 32 occurs near the center of the McBride province, which is situated within a Proterozoic crustal block, amid extensive late Paleozoic felsic volcanics and high-level intrusions [after White, 1965].

This study presents detailed major and trace element analyses for 12 lower crustal xenoliths from a young basaltic cinder cone, called Hill 32, in the McBride volcanic province of north Queensland, as well as mineralogical and mineral chemical data. The aim of this paper is to define the processes giving rise to the lower crust and compare the data presented here with lower crust composition estimates in order to refine these estimates. A separate paper presents U-Pb zircon ages determined by ion microprobe for seven of these xenoliths [Rudnick and Williams, 1987].

GEOLOGICAL SETTINGS AND PREVIOUS WORK

The xenoliths come from a single, young (<3 m.y. old) basaltic cinder cone, called Hill 32, in the McBride volcanic province of north Queensland. The McBride volcanics are one of six major late Cenozoic basaltic provinces which occur in north Queensland; lower crustal and upper mantle xenoliths are found in nearly all

provinces [Stephenson and Griffin, 1976]. Lower crustal xenoliths from the more southerly Chudleigh volcanic province were the subject of an earlier geochemical study [Rudnick *et al.*, 1986]. Hill 32 erupts through and onto Proterozoic crust of the Georgetown Inlier, which is composed of greenschist to amphibolite facies metamorphics intruded by 1570 Ma granites (Figure 1). Extensive Devonian and late Paleozoic calc-alkaline granites and ash flow tuffs occur throughout this region of north Queensland, and may have formed above a westward dipping subduction zone [Sheraton and Labonne, 1978]. An early seismic refraction survey suggests a crustal thickness between 40- and 45-km depth in this region [Finlayson [1968]; see Dooley [1980] for a review of available geophysical data].

Mineralogy and mineral chemistry of a separate suite of crustal xenoliths from Hill 32 were described by Kay and Kay [1983]. They collected a total of 19 xenoliths and

provide major element compositions for four of them. All xenoliths they analyzed are mafic quartz tholeiites and are interpreted as basaltic intrusions which crystallized at depth during the late Paleozoic Tasman orogeny. Major and trace element analyses of the host alkali basalt are given by Irving and Frey [1984].

ANALYTICAL TECHNIQUES

Mineral analyses reported in Tables 3-6 were obtained by energy-dispersive X-ray analysis using a Technisch Physische Dienst (TPD) electron probe at the Research School of Earth Sciences, Australian National University, with an accelerating voltage of 15 kV and a beam current of 3 nA. Data reduction was performed using peak integration with background subtraction and corrections as outlined by Ware [1981]. All mineral analyses represent rim compositions. Cr and Ni analyses of ferromagnesian phases in the intermediate to felsic xenoliths were performed on a Cameca WDS (Wavelength Dispersive Spectroscopy) electron probe using an accelerating voltage of 20 kV and a beam current of ~107 nA. Fe³⁺ in pyroxenes was calculated using the charge balance method of Papike et al. [1974].

Sample preparation is identical to that described by Rudnick et al. [1986]. Weathered surfaces were removed by sawing. The rocks were next crushed in a steel jaw crusher, split, and ~30 g was crushed in an agate ring mill. Major element analyses were determined by wide-beam EDS (Energy Dispersive Spectroscopy) microprobe analyses on glasses created in a positive Ar-pressure molybdenum strip heater. Each fused sample was analyzed a minimum of 10 times and the results averaged. Two samples (85-101 and 85-108) were fused a second time and give results indistinguishable from the first. P₂O₅ was measured for the same glasses using the Cameca, WDS, microprobe. TiO₂ and MnO were measured by inductively coupled plasma spectroscopy (ICP). Analytical uncertainty for all the major elements is less than 5%.

Trace elements were determined through a variety of techniques: Sc, V, Cr, Co, Ni, Cu, Sr, and Zr were determined by ICP (estimated uncertainty ≤ 10%) on all samples except 83-157, 83-160, and 83-162, for which these elements and Rb were measured by XRF [Norris and Chappell, 1977]. The remaining trace elements were determined through spark-source mass spectrometry (SSMS) [Taylor and Gorton, 1977]. Analytical uncertainty for SSMS analyses is as follows: ≤ 5% for rare earth element (REE); 5-10% for Ba, Y, Sn, W, Hf, Th and U; and 10-15% for Rb, Pb, Cs, Nb, and Mo.

THE XENOLITH SUITE

The Hill 32 xenoliths described here cover a marked mineralogical and chemical range. In total, 36 crustal xenoliths, of variable sizes, were collected at Hill 32. These represent all crustal xenoliths which were encountered during two separate collecting trips to this locality. From thin section examination it is possible to derive the relative lithologic proportions of these crustal xenoliths, which are given in Table 1, column one. Column two includes the samples described by Kay and Kay [1983], and column three shows relative lithologic proportions for a suite of 47 crustal xenoliths collected by

TABLE 1. Lithologic Proportions for Hill 32 Crustal Xenoliths

	1	2	3
Mafic granulites	23 (64%)	42 (76%)	38 (80%)
Intermediate granulites	7 (19%)	7 (13%)	9 (20%)
Felsic granulites	4 (11%)	4 (7%)	
Granites (upper crustal)	2 (6%)	2 (4%)	
Total	36	55	47

1, all samples from this study; 2, this study plus samples from Kay and Kay [1983]; 3, from Stolz [1987, and personal communication, 1987].

A. J. Stolz (personal communication, 1986). Keeping in mind the uncertainties regarding the randomness of sampling by the basalt and the thoroughness of collection, these proportions may be regarded as rough estimates of the lower crustal composition. Clearly, the majority of xenoliths are mafic, but intermediate to felsic granulites comprise a significant proportion of the lower crust here (~20%).

Mineralogy and Petrography

Mineralogy of the Hill 32 xenoliths varies considerably, reflecting the variable bulk-rock composition and PT equilibration conditions. Table 2 lists the mineralogies of the 35 lower crustal xenoliths collected, along with textures, grain size, and P and T estimated from various thermobarometers. Petrographic descriptions of the analyzed xenoliths are given in the appendix. In general, Hill 32 xenoliths possess well-equilibrated textures (Table 2 and Kay and Kay, [1983]); bona fide relict igneous textures are confined to one 2-pyroxene granulite (85-125). Coronas were observed in several xenoliths: one metasedimentary xenolith (85-101) contains very thin rims of orthopyroxene on garnet, and four mafic xenoliths have garnet rimming spinel (85-125), clinopyroxene (85-106), and Fe-oxides (85-107, 85-122). The coronas in the metasediment reflect garnet breakdown due to decreasing pressure or increasing temperature, whereas the coronas in the mafic xenoliths reflect garnet formation through isobaric cooling reactions [Lovering and White, 1969; Ellis and Green, 1985].

Decompression features are present in nearly all Hill 32 xenoliths. Garnet in xenoliths of all compositions is variably replaced by dark kelyphite, which may contain quenchlike crystals up to 0.25 mm long. Figure 2a shows garnet idiomorphs which are replaced by skeletal anorthite, pyroxene, and spinel. Kelyphite in these and other xenoliths is interpreted to form by decompression melting brought on by the rapid eruption of the host basalt from great depths [Padovani and Carter, 1977b; Kay and Kay, 1983; Garvey and Robinson, 1984]. Similarly, scapolite is rimmed by anorthite aggregates, which are also interpreted to be due to rapid decompression [Lovering and White, 1964]. The intermediate to felsic xenoliths contain colorless to brown glass along grain boundaries, particularly near biotite (Figure 2b), which is similar to that described in xenoliths from Kilbourne Hole, New Mexico [Padovani and Carter, 1977b].

Hydrous phases are relatively common in Hill 32

TABLE 2. Mineralogies and Textures of Hill 32 Lower Crustal Xenoliths

Sample	Size, g	Major Minerals	Accessory Minerals	Texture/ Grain Size	Layering (Thickness)	T, °C P, GPa
<i>Mafic Xenoliths: Two-Pyroxene Granulites</i>						
85-100	300	Opx*-Pc-Cpx	Rut-Bio-Il-Ap-Zir	Poly, 1 mm	3.5 cm	960-1000 ⁰¹
85-102		Opx-Pc-Cpx	Am-Rut-Ap-Zir	Poly, 1mm	yes	
85-120	340	Pc-Opx-Cpx	Il-Am-Rut-Zir	Poly, 1 mm	yes	860-910 ⁰¹
85-123		Pc-Cpx-Opx	Sp-Gar-Am	Coronal, <1 mm	no	
85-125		Pc-Opx-Cpx	Rut	Poly, 1 mm	no	
85-126		Pc-Cpx-Opx	Am-Rut	Poly, 2 mm	no	
85-130		Opx-Pc-Cpx	Il-Ap-Zir	Poly, 1 mm	yes	
85-133		Pc-Cpx-Opx	Ox-Ap	Cata, Poly, 2 mm	no	
<i>Mafic Xenoliths: Garnet-Clinopyroxene Granulites</i>						
83-159	650	Pc-Gar-Cpx [§]	Am-Il-Mt-Rut-Ap-Zir	Poly, 2 mm	no	840-920 ⁰²
85-106	330	Cpx [†] -Gar*-Am-Sc	Il-Pc-Mt	Coronal, 4 mm	no	890-1070 ⁰²
85-107	220	Gar-Pc-Q	Cpx-Il-Zir-Ap	Poly, 1 mm	no	680-760 ⁰²
85-114	410	Gar-Cpx-Pc	Rut	Poly, 2 mm	no	830-910 ⁰²
85-119		Cpx-Pc-Gar	Q-Rut	Poly, 1 mm	no	
85-121		Cpx-Gar*-Pc	Il-Rut-Am	Poly, 3 mm	no	
85-122		Cpx-Gar-Pc-Am	Il	Coronal, Poly, 3 mm	no	
<i>Mafic Xenoliths: Two Pyroxene Garnet Granulites</i>						
83-158	(300)	Cpx [†] -Gar*-Pc	Am-Il-Opx-Zir	Poly, 3 mm	no	940-1000 ⁰² , 1.1-1.5 ³
85-108	270	Pc-Cpx-Gar*Q	Rut-Opx-Bio-Ap	Poly, 1 mm	no	810-860 ⁰² , 0.85 ⁴
85-110		Cpx-Gar-Pc	Opx-Ox-Am	Poly, 3 mm	no	
85-124		Pc-Gar-Opx	Cpx-Rut	Poly, 1 mm	no	
85-127		Pc-Gar-Opx-Cpx	Il-Rut	Poly, 1 mm	no	
85-129		Cpx-Gar-Pc	Il-Opx	Poly, 3 mm	no	
<i>Mafic Xenoliths: Others</i>						
85-117		Gar*-Pc-Q	Rut	Poly, 1-2 mm	yes	
85-118		Pc-Cpx	Il-Ap-Am	Poly, 3 mm	no	
<i>Intermediate Xenoliths</i>						
83-157	280	Pc*-Gar*-Q*-Opx*	Bio-Cpx-Rut-Ap-Zir	Poly, 2 mm	1 cm	630 ⁰² , 800 ⁰⁵ C 0.8-1.0 ⁴
85-99		Gar-Pc-Q	Sil-Rut-Il-Ap	Poly, 3 mm	yes	
85-101	170	Gar-Pc-Q*	Rut-Il-Opx-Sil-Zir	Poly, Coronal, 3 mm	1 cm	1.1-1.3 ⁴ 0.83-1.2 ⁶
85-103		Gar-Pc-Q	Rut	Poly, 1 mm	yes	
85-111		Pc-Q-Opx-Cpx	Gar-Ap-Ox	Poly, 1 mm	no	
85-128		Gar-Pc-Q	Rut-Il	Porph, Poly, 2 mm	no	
85-131		Gar-Pc-Sil	Rut-Zir	Poly, 2 mm	no	
<i>Felsic Xenoliths</i>						
83-160	290	Pc-Q*-Opx-Gar-Kf	Cpx-Il-Bio-Mt-Ap Rut-Zir	Poly, 1 mm	no	740-900 ⁰¹ , 0.9-1.2 ⁴
83-162	320	Q*-Kf*-Gar	Ap-Zir-Rut-Mt-Il	Poly, 1 mm	no	
85-112		Q-Pc-Gar	Bio-Zir	Poly, 1 mm	no	
85-132		Pc-Q-Gar	Rut-Ap	Poly, 1 mm	no	

Poly, Polygonal; cata, cataclastic; porph, porphyroblastic

Minerals are listed in order of estimated relative abundance. Am, amphibole; Ap, apatite; Bio, biotite; Cpx, clinopyroxene; Gar, garnet; Il, ilmenite; Kf, alkali feldspar; Mt, magnetite; Opx, orthopyroxene;

Ox, Fe-oxide; Pc, plagioclase; Q, quartz; Rut, rutile; Sc, scapolite; Sil, sillimanite; Zir, zircon.

*Contains acicular rutile inclusions.

†Exsolution lamellae throughout mineral.

§Exsolution lamellae restricted to cores of mineral.

¹Wells [1977]; ²Ellis and Green [1979]; ³Harley and Green [1982]; ⁴Perkins and Newton [1981];

⁵Ferry and Spear [1978], and ⁶Newton and Haselton [1981].

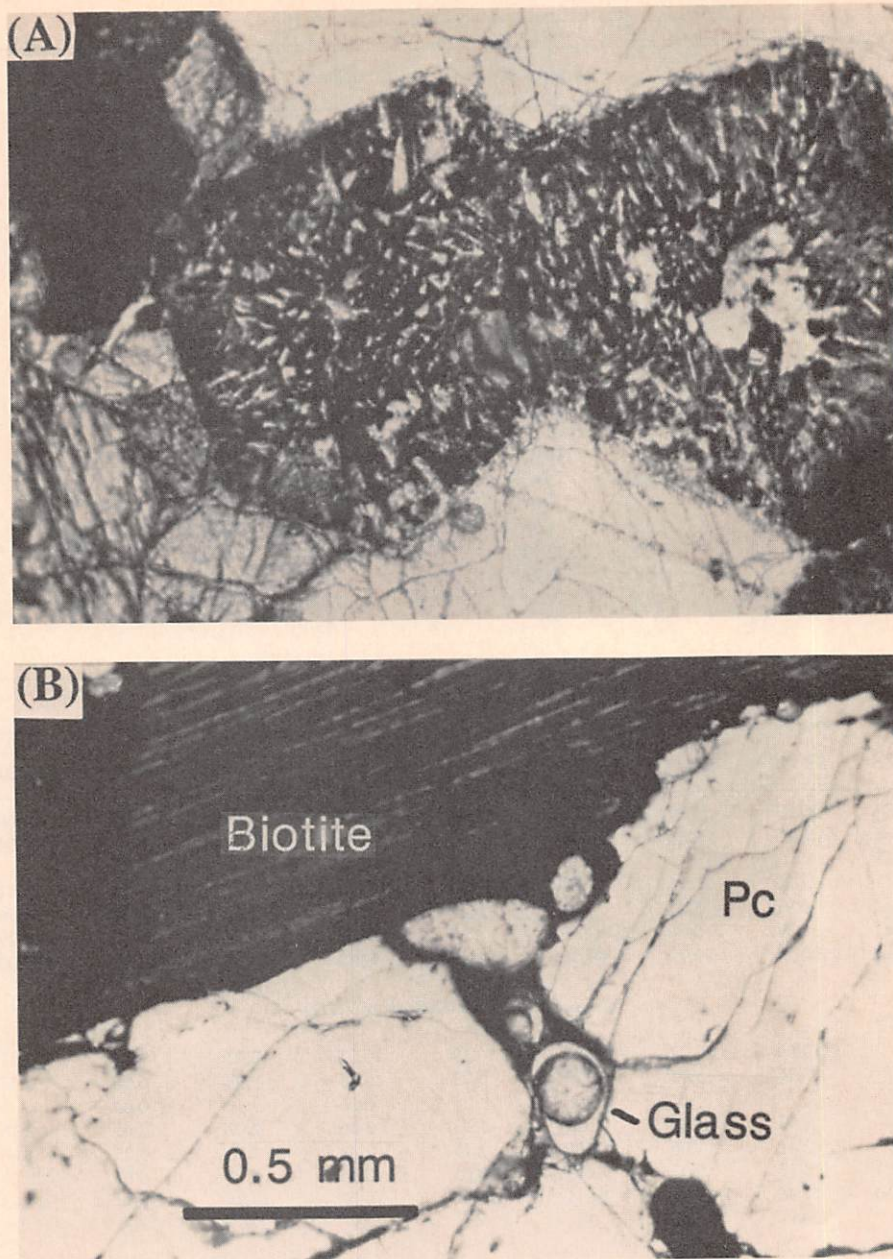


Fig. 2. Photomicrographs showing decompression features in McBride xenoliths. (a) Skeletal quench crystals of plagioclase (white) spinel and pyroxene (dark) in melted garnet in 85-127. White area at core of crystal on right is relict garnet. (b) Glass along crystal boundaries in intermediate xenolith 83-157. Scale bar applies to both photos.

xenoliths, as noted by *Kay and Kay* [1983]. In the samples investigated here, amphibole generally makes up less than 5% of the rock, and it can sometimes be seen rimming pyroxene. Sulfur-rich meionitic scapolite, which is shown to form in granulite facies under high activity of CO_2 [Goldsmith, 1976], is present in only one mafic xenolith; however, CO_2 -rich fluid inclusions have been observed in nearly all xenoliths. Rutile, apatite, and ilmenite are the most common accessory phases. Zircon is present in several xenoliths and has been used to date the xenoliths [Rudnick and Williams, 1987].

Mineral Chemistry

Representative mineral analyses for four of the McBride xenoliths have been previously reported [Rudnick and

Taylor, 1987]. In addition, *Kay and Kay* [1983] give representative mineral analyses for several Hill 32 xenoliths they studied. Representative microprobe analyses of feldspars, clinopyroxenes, orthopyroxenes, amphiboles, and garnets are given in Tables 3-6 for the xenoliths which have been analyzed for major and trace elements. The mafic xenoliths range from two-pyroxene granulites to garnet-clinopyroxene granulites; intermediate to felsic granulites all contain garnet, and several contain either pyroxenes or sillimanite. Minerals are generally moderately zoned; the mineral compositions reflect both whole-rock chemistry and deep crustal equilibration.

Feldspars. Feldspar compositions are shown in Figure 3 and representative analyses are given in Table 2. Plagioclases in the intermediate to felsic xenoliths

TABLE 3. Representative Analyses of Feldspars in Hill 32 Xenoliths

Intermediate and Felsic Xenoliths								
	83-157	85-101*	85-101†	83-160	83-160‡	83-162		
SiO ₂	56.34	57.29	60.10	56.95	64.65	65.41		
Al ₂ O ₃	27.76	26.60	24.48	27.23	20.17	19.06		
FeO		0.24	0.60		0.30	0.14		
CaO	9.99	8.06	5.72	8.83	0.93	0.09		
Na ₂ O	5.45	6.04	7.13	6.12	4.66	2.25		
K ₂ O	0.45	0.89	1.34	0.67	9.11	12.93		
Total	100.00	99.12	99.36	99.79	99.81	99.88		
<i>Number of Ions on the Basis of 32 (O)</i>								
Si	10.13	10.37	10.81	10.25	11.70	11.99		
Al	5.88	5.67	5.19	5.77	4.30	4.10		
Fe		0.04	0.09		0.04	0.02		
Ca	1.92	1.56	1.10	1.70	0.18	0.02		
Na	1.90	2.12	2.49	2.13	1.64	0.80		
K	0.10	0.20	0.31	0.15	2.10	3.01		
Total	19.93	19.96	19.98	20.01	19.97	19.90		
An	49	40	28	43				
Mafic Xenoliths								
	83-158	83-159	85-100	85-106	85-107	85-108	85-114	85-120
SiO ₂	59.83	55.94	53.94	56.77	58.76	55.97	52.35	54.60
Al ₂ O ₃	25.55	28.06	28.76	27.37	25.75	28.32	30.95	29.04
FeO			0.39		0.41		0.26	
CaO	7.20	10.11	11.69	9.24	7.62	10.50	13.76	11.31
Na ₂ O	6.97	5.61	4.28	6.13	6.70	5.05	3.58	4.76
K ₂ O	0.74	0.06	0.38	0.24	0.58	0.71	0.12	0.29
Total	100.28	99.79	99.44	99.87	99.82	100.55	101.00	100.00
<i>Number of Ions on the Basis of 32 (O)</i>								
Si	10.65	10.07	9.76	10.19	10.52	10.03	9.42	9.85
Al	5.36	5.95	6.13	5.79	5.43	5.98	6.56	6.17
Fe			0.06		0.06		0.04	
Ca	1.37	1.95	2.27	1.78	1.46	2.02	2.65	2.18
Na	2.41	1.96	1.50	2.14	2.32	1.75	1.25	1.66
K	0.17	0.01	0.09	0.05	0.13	0.71	0.03	0.07
Total	19.96	19.94	19.81	19.96	19.92	19.94	19.94	19.93
An	35	50	59	45	37	45	67	56

* Matrix plagioclase

† Plagioclase rimming orthopyroxene on garnet

‡ K-feldspar part of perthite.

from An₅₀ to An₃₅. Plagioclase rims associated with orthopyroxene coronas in 85-101 are distinctly more sodic than the matrix plagioclases. Alkali feldspar is present in two of the felsic xenoliths and is characterized by perthitic texture. Minor compositional zoning is present in plagioclases from most samples, with Na increasing toward the rims; however, plagioclases in 83-157 have slightly more calcic rims. Plagioclases in the mafic xenoliths exhibit a wide compositional range from An₇₅ to An₃₆; more than 4 times the range found by *Kay and Kay* [1983] for Hill 32 xenoliths. Generally, only minor zoning is present in these plagioclases (all but one sample (85-100) have Na increasing toward the rims). However, plagioclase in 85-114 is strongly zoned, with Na-rich rims (rims, An₅₁; cores, An₇₂).

Garnet. Garnets are particularly susceptible to alteration through decompression melting; all garnets in Hill 32 xenoliths have been melted to varying degrees, resulting in varying thicknesses of kelyphitic rims. In some instances, analysis of this kelyphite yields a garnetlike composition and fits a garnet formula. In other cases, Na₂O and/or K₂O are present, indicating incorporation of components external to the garnet. The analyses presented in Table 4 and plotted in Figure 4 are from pristine garnets, unless otherwise indicated.

Garnets in the intermediate and felsic xenoliths are generally less calcic than those from the mafic xenoliths (Figure 4). These garnets contain between 7 and 18% grossular, and all fall within *Coleman et al.*'s [1965] group B classification, garnets from crustal eclogites. Garnets in

TABLE 4. Representative Analyses of Garnets in Hill 32 Xenoliths

Intermediate and Felsic Xenoliths						
	83-157	85-101	83-160*	83-162		
SiO ₂	39.92	39.48	38.45	39.37		
Al ₂ O ₃	22.47	22.89	22.35	21.89		
FeO	20.93	20.38	23.92	24.18		
MnO	0.29	0.82	0.59	1.24		
MgO	11.65	2.75	8.80	10.07		
CaO	4.44	3.54	6.52	3.39		
Total	99.77	99.37	100.63	100.14		
<i>Number of Ions on the Basis of 24 (O)</i>						
Si	6.01	5.95	5.87	6.00		
Al	3.99	4.06	4.02	3.94		
Fe	2.64	2.57	3.05	3.08		
Mn	0.04	0.10	0.08	0.16		
Mg	2.61	2.75	2.00	2.29		
Ca	0.72	0.57	1.06	0.55		
Total	16.00	16.02	16.08	16.03		
Mafic Xenoliths						
	83-158	83-159	85-106	85-107	85-108	85-114
SiO ₂	40.33	39.41	40.62	38.06	39.04	40.28
Al ₂ O ₃	22.38	22.06	22.30	21.16	21.94	22.70
FeO	20.34	21.52	19.50	28.27	23.01	19.54
MnO	0.37	0.48	0.54	0.56	0.47	0.36
MgO	11.38	9.84	12.04	4.16	8.52	11.03
CaO	6.01	7.33	5.89	7.80	7.02	6.21
Total	100.80	100.66	100.88	100.00	100.00	100.12
<i>Number of Ions on the Basis of 24 (O)</i>						
Si	6.01	5.96	6.03	6.00	5.98	6.03
Al	3.93	3.93	3.90	3.93	3.96	4.00
Fe	2.54	2.72	2.42	3.73	2.95	2.44
Mn	0.05	0.06	0.07	0.07	0.06	0.04
Mg	2.53	2.22	2.66	0.98	1.94	2.46
Ca	0.96	1.19	0.94	1.32	1.15	0.99
Total	16.02	16.08	16.02	16.03	16.04	15.97

* Kelyphitic.

two samples, 83-157 and 85-101, show a slight increase in Ca content near their rims; the remaining garnets are unzoned. Garnets in the mafic xenoliths have from 15 to 28% grossular, and all but one fall within group B. Garnets in 85-107, an extremely Mg-poor mafic xenolith, fall well into the group C classification (blueschist garnets), reflecting the bulk-rock composition. Although zoning patterns are difficult to delineate because of the kelyphitization, garnets in sample 83-159 have a considerable range in grossular content.

Pyroxenes. Composition of clinopyroxenes in the

McBride xenoliths are shown plotted on the pyroxene quadrilateral in Figure 5 (nonquadrilateral components were subtracted from the analyses as described by *Lindsley* [1983]). Orthopyroxene forms the typical exsolution lamellae, where present, but in one sample (83-158) both garnet and plagioclase occur as exsolution lamellae within the clinopyroxene. Typically, TiO₂, Al₂O₃, and Na₂O contents are quite variable, with most clinopyroxenes having lower concentrations of these oxides on their rims or near exsolution lamellae. FeO, MgO, and CaO zonations are less pronounced, but clinopyroxenes in both

TABLE 5. Representative Analyses of Pyroxenes in Hill 32 Xenoliths

	Intermediate and Felsic Xenoliths							
	Clinopyroxene		Orthopyroxene					
	83-157	83-160	83-157	85-101*	83-160			
SiO ₂	52.65	51.93	53.99	53.23	52.23			
TiO ₂		0.47						
Al ₂ O ₃	3.28	4.10	2.87	2.84	1.76			
FeO*	4.04	8.45	16.07	21.36	24.01			
MnO				0.27	0.14			
MgO	15.00	13.18	26.53	21.11	21.08			
CaO	23.30	21.36	0.39	0.41	0.77			
Na ₂ O	0.24	0.51						
Total	98.51	100.00	99.85	99.21	100.00			
<i>Number of Ions on the Basis of 6 (O)</i>								
Si	1.95	1.92	1.95	1.98	1.96			
Al	0.14	0.18	0.12	0.12	0.08			
Ti		0.01						
Fe	0.12	0.26	0.48	0.66	0.75			
Mn				0.01	0.01			
Mg	0.83	0.73	1.42	0.17	1.18			
Ca	0.92	0.85	0.02	0.02	0.03			
Na	0.02	0.04						
Total	3.99	3.99	3.99	3.96	4.00			
Mafic Xenoliths								
	Clinopyroxenes							
	83-158	83-159	85-100	85-106	85-107	85-108	85-114	85-120
SiO ₂	49.60	50.53	52.25	49.64	51.60	52.55	51.32	51.85
TiO ₂	1.01	0.63	0.43	0.30		0.28	0.56	0.50
Al ₂ O ₃	7.31	5.98	4.04	8.37	2.88	3.15	4.49	4.16
Cr ₂ O ₃			0.22					
FeO*	8.20	7.05	6.67	8.06	12.14	7.17	6.83	7.48
MgO	11.75	12.43	14.49	11.50	11.31	13.68	13.47	13.17
CaO	20.76	22.22	22.07	20.76	21.62	22.86	22.09	22.58
Na ₂ O	1.37	0.82	0.16	1.22	0.43	0.14	0.54	0.27
Total	100.00	99.66	100.34	99.84	99.98	100.00	99.30	100.02
<i>Number of Ions on the Basis of 6 (O)</i>								
Si	1.84	1.88	1.92	1.84	1.95	1.95	1.91	1.92
Al	0.32	0.26	0.18	0.37	0.13	0.14	0.20	0.18
Ti	0.03	0.02	0.01	0.01		0.01	0.02	0.01
Fe	0.25	0.22	0.20	0.25	0.38	0.75	0.21	0.23
Mg	0.65	0.69	0.79	0.64	0.64	0.75	0.75	0.73
Ca	0.83	0.88	0.88	0.82	0.88	0.90	0.88	0.89
Na	0.10	0.06	0.01	0.09	0.03	0.01	0.04	0.20
Total	4.02	4.00	3.99	4.01	4.00	3.98	4.00	3.99

* All Fe as FeO.

† Occurs as thin rims on garnets.

pyroxene granulites have higher Mg numbers on their rims. The Mg numbers of the clinopyroxenes in the mafic xenoliths do not correlate with those of the whole rocks, suggesting that equilibration temperatures (which range between 700 and 1000°C, Table 2) exerted the dominant influence on this parameter. There is a rough negative correlation between Na₂O in the clinopyroxenes and the anorthite content of the coexisting plagioclases for the mafic xenoliths, which is attributed to the breakdown of

calcic plagioclase under high pressures (Figure 6). Such a correlation was noted by *Griffin et al.* [1979] for Lesotho xenoliths and *Kay and Kay* [1983] for the Queensland xenoliths.

Orthopyroxene compositions are given in Table 5 and plotted in Figure 5. The crossing tie lines observed on the pyroxene quadrilateral for the mafic xenoliths are attributed to variable equilibration temperatures and whole-rock compositions between the xenoliths.

Table 5 (continued)

Mafic Xenoliths				
	Orthopyroxenes			
	83-158	85-100	85-108	85-120
SiO ₂	53.66	55.19	53.29	53.10
Al ₂ O ₃	3.09	2.56	2.20	3.04
FeO*	18.48	17.67	20.73	20.85
MnO	0.12	0.36	0.22	0.39
MgO	24.64	24.44	22.80	22.25
CaO	0.56	0.62	0.75	0.37
Total	100.55	100.84	100.00	100.00
<i>Number of Ions on the Basis of 6 (O)</i>				
Si	1.94	1.98	1.96	1.96
Al	0.13	0.11	0.09	0.13
Fe	0.56	0.53	0.64	0.64
Mn	0.003	0.01	0.01	0.01
Mg	1.33	1.31	1.25	1.22
Ca	0.02	0.02	0.03	0.02
Total	3.99	3.96	3.99	3.98

Volatile-bearing phases. Amphibole is present in variable proportions in half of the analyzed mafic xenoliths but is absent in the intermediate to felsic xenoliths (Table 2). Compositions are given in Table 6; all amphiboles are ferroan pargasites to ferroan pargasitic hornblendes (based on Hawthorne's [1981] classification). The amphiboles are compositionally uniform throughout the suite, irrespective of whether they occur as rims on pyroxenes or as discrete grains. Biotite occurs in several

of the intermediate to felsic xenoliths and one of the mafic xenoliths. Analyses are given in Table 6; all are Ti-rich. Scapolite is present in only one of the xenoliths and it is a sulfur-rich meonite (Table 6).

PT conditions. Table 2 lists temperature and pressure estimates calculated from several thermobarometers. In all cases, adjacent mineral rim analyses were used in these calculations in order to give best estimates of last equilibration conditions. Ferric Fe was calculated based

TABLE 6. Representative Analyses of Volatile-Bearing Phases in Hill 32 Xenoliths

	<i>Biotites</i>			<i>Amphiboles</i>				<i>Scapolite</i>
	83-157	83-160	85-100	83-158	83-159	85-106	85-120	85-106
SiO ₂	37.48	37.24	36.93	40.49	41.52	40.18	40.36	45.17
TiO ₂	6.09	6.55	5.29	3.38	3.24	2.74	1.97	
Al ₂ O ₃	14.81	14.14	14.74	14.16	15.35	15.09	13.14	26.14
Cr ₂ O ₃			0.33	0.09	0.18		0.29	
FeO	9.42	14.42	10.06	12.51	11.57	12.31	12.64	0.28
MnO				0.06				
MgO	17.07	13.52	16.56	13.32	12.57	11.61	11.72	0.27
CaO				10.66	11.86	11.25	11.49	16.91
Na ₂ O	0.15	0.33		3.29	2.77	2.55	1.38	3.44
K ₂ O	9.20	8.97	9.60	0.81	0.42	1.10	1.46	0.12
SO ₃								4.19
Total	94.37	95.18	93.51	98.80	99.47	96.82	94.45	96.53
<i>Number of Ions on the Basis of 24 (O)</i>								
Si	6.00	6.05	6.01	6.20	6.25	6.25	6.46	
Al	2.80	2.71	2.83	2.55	2.72	2.77	2.48	
Ti	0.73	0.80	0.65	0.39	0.37	0.32	0.24	
Cr			0.04	0.01	0.02	0.04		
Fe	1.26	1.96	1.37	1.60	1.46	1.60	1.69	
Mn				0.01				
Mg	4.08	3.28	4.02	3.04	2.82	2.69	2.80	
Ca				1.75	1.91	1.88	1.97	
Na	0.05	0.10	0.98	0.81	0.77	0.43		
K	1.88	1.86	1.99	0.16	0.08	0.22	0.30	
Total	16.82	16.77	16.90	16.69	16.46	16.51	16.40	

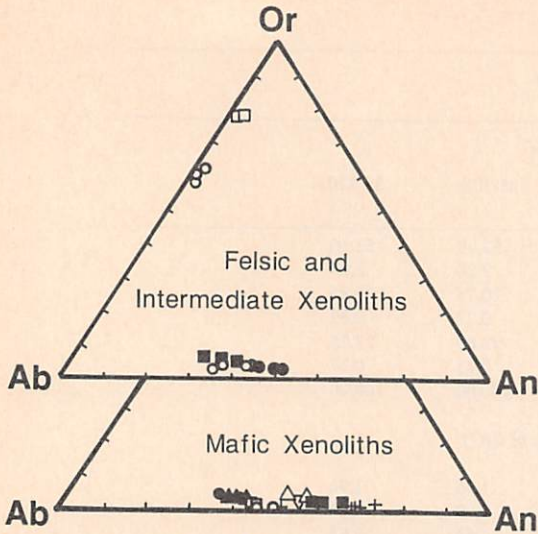


Fig. 3. Composition of feldspars in analyzed Hill 32 xenoliths. Each symbol represents a separate sample and total observed range in compositions are plotted. Key to symbols: UPPER [filled circle] = 83-157, [filled square] = 85-101, [open circle] = 83-160, [open square] = 83-162; LOWER [filled circle] = 83-158, [open circle] = 83-159, [filled square] = 85-100, [open square] = 85-106, [filled triangle] = 85-107, [open triangle] = 85-108, [plus] = 85-114, [inverted open triangle] = 85-120.

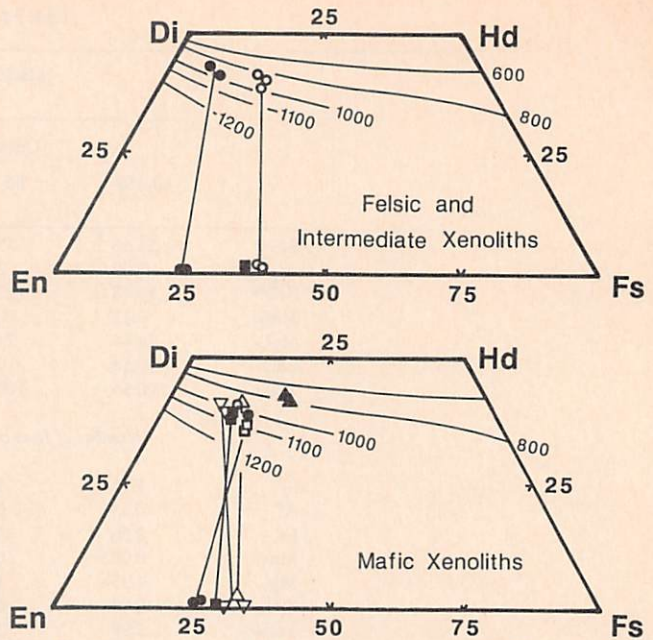


Fig. 5. Compositions of pyroxenes in Hill 32 xenoliths. Projections onto pyroxene quadrilateral follow method of *Lindsley* [1983]; experimentally determined isotherms are given for reference, at 1.0-GPa pressure. Symbols as in Figure 3.

upon charge balance for pyroxene analyses [*Papike et al.*, 1974].

The compositional zonation within the clinopyroxenes and some feldspars may indicate cooling, suggesting that the calculated temperatures represent the blocking temperature for Ca-Mg-Fe exchange in each xenolith. However, the general lack of mineral reactions within the

xenoliths suggests that minimum cooling and/or pressure change has occurred to these rocks since formation of their metamorphic mineral assemblages. Exceptions to this include the orthopyroxene rims on garnet in 85-101 and various coronal textures (discussed above) in the mafic xenoliths. All but the former are interpreted in terms of isobaric cooling reactions. The temperatures and pressures may therefore be regarded as rough estimates of the xenolith's pressure and temperature of last equilibration. Most xenoliths have equilibrated in the deep crust, between 26 and 40 km. However, the relatively low temperature recorded in mafic granulite 85-107 (680-760°C) suggests a shallower derivation, perhaps as shallow as 18 km.

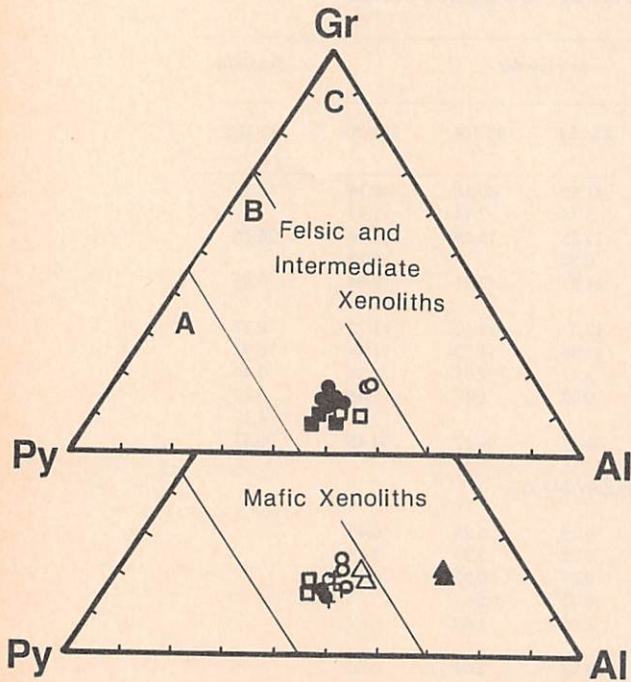


Fig. 4. Composition of garnets in Hill 32 xenoliths. Fields A, B, and C are eclogitic garnet fields for kimberlites, gneisses, and blueschists, respectively [*Coleman et al.*, 1965]. Symbols as in Figure 3.

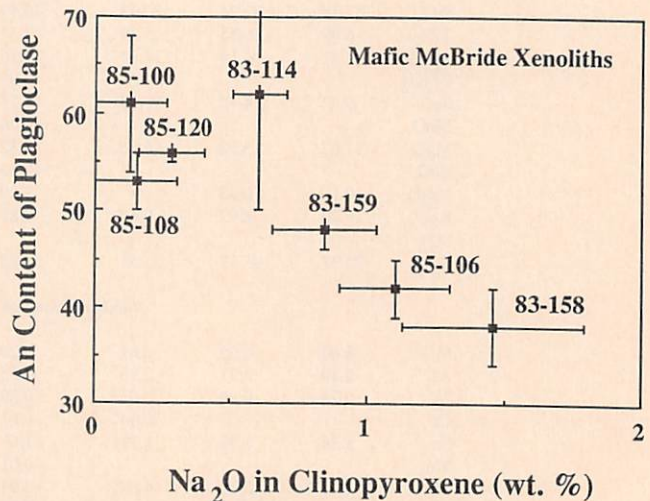


Fig. 6. Na₂O in clinopyroxene versus anorthite content of coexisting plagioclases for McBride mafic xenoliths, excluding 85-107 (which is a cumulate from a felsic magma). Error bars represent total observed compositional variations for the minerals.

TABLE 7. Major Element Compositions and CIPW Normative Mineralogies of Hill 32 Xenoliths

	Metasediments		Felsic Melts		Mafic Melts			Cumulates		Restites/Cumulates		
	83-157	85-101	83-160	83-162	85-100	85-108	85-120	83-158	85-106	83-159	85-107	85-114
SiO ₂	54.61	55.93	63.98	66.94	51.70	48.90	52.39	46.72	46.31	43.41	52.77	41.20
TiO ₂	1.47	0.57	0.91	0.56	0.88	1.09	0.57	1.13	1.41	1.90	1.46	3.26
Al ₂ O ₃	19.79	19.83	15.74	14.30	11.26	15.16	16.50	14.70	15.65	19.72	19.04	19.63
FeO*	9.91	11.69	5.88	6.89	13.30	12.42	9.45	11.73	9.73	15.43	14.46	17.11
MnO	0.15	0.35	0.16	0.28	0.20	0.22	0.15	0.20	0.15	0.27	0.25	0.33
MgO	6.28	5.33	2.98	2.71	14.50	8.48	9.17	10.42	9.20	7.58	2.18	9.18
CaO	4.63	3.95	4.31	1.59	6.82	12.70	9.30	13.60	15.61	9.29	6.50	8.36
Na ₂ O	2.20	1.98	3.34	1.45	1.06	0.89	2.34	1.45	1.76	1.38	2.27	0.45
K ₂ O	0.35	0.39	1.56	4.24	0.16	0.16	0.19	0.10	0.10	0.22	0.28	0.14
P ₂ O ₅	0.23	0.15	0.45	0.21	0.19	0.32	0.28	0.07	0.19	0.58	0.78	0.26
Total	99.62	100.17	99.31	99.17	100.07	100.34	100.34	100.12	100.11	99.78	99.99	99.92
Mg #	53.0	44.8	47.4	41.2	66.0	54.9	63.4	61.3	62.8	46.7	21.2	48.9
<i>C.I.P.W. Norms</i>												
q	16.0	20.8	20.4	30.4	2.0	1.2	1.2	0	0	0	14.6	0
c	7.9	8.4	1.8	4.6	0	0	0	0	0	1.7	5.0	4.2
or	2.1	2.5	9.2	25.0	0.9	0.9	1.1	0.6	0.6	1.3	1.6	0.8
ab	18.6	16.7	28.3	12.3	9.0	7.5	19.8	12.3	9.5	11.7	19.2	3.8
an	21.5	18.0	18.4	7.4	25.5	36.9	34.0	33.3	34.5	42.3	27.2	40.0
ne	0	0	0	0	0	0	0	0	2.9	0	0	0
di	0	0	0	0	5.8	19.8	8.4	27.4	33.9	0	0	0
hy	28.5	30.1	20.1	16.9	52.5	29.2	32.5	4.5	0	22.8	25.0	35.0
ol	0	0	0	0	0	0	0	17.4	13.9	12.3	0	6.4
mt	1.9	2.2	1.1	1.3	2.5	2.4	1.8	2.2	1.9	3.0	2.8	3.3
il	2.8	1.1	1.5	1.1	1.3	1.9	0.8	2.6	2.5	3.7	3.0	6.3
ap	0.5	0.4	1.1	0.2	0.4	0.8	0.7	0.2	0.4	1.4	1.9	0.6

Here q, quartz; c, corundum; or, orthoclase; ab, albite; an, anorthite; ne, nepheline; di, diopside; hy, hypersthene; ol, olivine; mt, magnetite; il, ilmenite; ap, apatite.

Mg # = $100(\text{Mg}/\text{Mg} + \sum\text{Fe})$.

* Total Fe as FeO.

Whole-Rock Composition

Analyzed xenoliths were selected based upon lithology, freshness and size. Most analyzed xenoliths were larger than 200 g (Table 2), and none contain macroscopic or microscopic evidence of invasion of the host basalt. The 12 analyzed xenoliths span the observed lithologic range.

Metamorphic recrystallization may produce layering in previously massive rocks (i.e., metamorphic differentiation). Thus if sample sizes are small in relation to the layering, certain minerals will be preferentially sampled, causing a corresponding change in the whole-rock chemistry. For this reason, the analyzed xenoliths' sizes are given in Table 2, along with the presence or absence of layering and the typical thicknesses of observed layers. Two layered McBride xenoliths (83-162 and 85-101) appear to have experienced metamorphic enrichment of garnet (as seen in enriched heavy rare earth element (HREE) contents and higher FeO and MgO) and one (85-100) has been enriched in orthopyroxene (giving the whole-rock low Al₂O₃ and high Cr and Ni contents). These features are discussed in more detail in the appropriate sections.

Major Elements. Major element compositions, along with CIPW norms for the xenoliths are given in Table 7. The mafic rocks range from silica oversaturated to undersaturated compositions. Mg numbers range from 47 to 66, but one very Mg-poor sample has an Mg number of 21. Only one of the mafic xenoliths (85-120) has a major element composition similar to common mafic igneous rock types; 85-108 has low Na₂O, but otherwise its bulk composition is similar to that of a silica oversaturated tholeiite. The remaining mafic xenoliths have major element compositions which have been modified by secondary processes such as crystal accumulation and partial melt extraction. For example, three of the mafic xenoliths have Al₂O₃ contents higher than 19% (83-159, 85-107, and 85-114), suggesting preferential accumulation of plagioclase. As none of these xenoliths are layered, the plagioclase increase is likely to be due to either cumulate or partial melting processes and not metamorphic differentiation. One mafic xenolith has very low CaO and Al₂O₃ contents (85-100, with 6.8 and 11%, respectively). This xenolith contains a large orthopyroxene-rich layer

TABLE 8. Rare Earth Element and Y Contents of Hill 32 Xenoliths (ppm)

	Metasediments		Felsic Melts		Mafic Melts			Cumulates		Restites/Cumulates		
	83-157	85-101	83-160	83-162	85-100	85-108	85-120	83-158	85-106	83-159	85-107	85-114
La	33.2	10.2	22.9	29.2	15.5	11.3	10.4	2.6	4.0	14.7	13.6	4.4
Ce	66.1	17.9	48.7	49.4	38.3	30.0	25.0	8.2	13.4	37.0	30.8	9.2
Pr	7.25	1.65	5.62	5.28	4.61	4.20	3.56	1.38	2.06	4.72	4.47	1.19
Nd	29.7	7.3	23.4	21.3	21.2	18.1	16.9	8.0	11.1	22.6	21.0	5.7
Sm	6.46	3.32	5.36	6.18	4.44	4.38	4.13	2.90	3.60	4.28	5.36	1.64
Eu	1.61	1.20	1.57	1.25	1.00	1.00	1.23	0.95	1.25	2.27	2.09	1.00
Gd	6.20	5.97	4.59	8.06	3.43	3.70	3.93	3.64	4.02	3.49	5.87	2.82
Tb		1.29	0.81	1.77	0.56	0.58	0.64		0.72	0.59	0.99	0.64
Dy	6.74	8.18	5.26	12.36	3.25	3.43	3.77	4.77	4.20	3.98	5.88	5.38
Ho	1.40	1.95	1.23	2.82	0.62	0.71	0.78	1.07	0.90	0.89	1.17	1.48
Er	3.90	5.61	3.83	8.02	1.69	2.08	2.17	2.96	2.29	2.48	3.40	5.00
Yb	4.08	5.70	4.43	7.96	1.50	1.85	1.81	2.80	1.75	2.84	3.19	5.35
Eu/Eu*	0.77	0.82	0.94	0.54	0.78	0.75	0.93	0.90	1.00	1.72	1.17	1.43
(La/Yb) _N	5.8	1.3	4.2	2.7	5.9	3.8	3.2	0.6	1.1	4.1	2.8	0.8
Y	35	47	33	67	16	18	20	26	22	22	30	34

(Table 2), suggesting its composition has been modified by metamorphic concentration of orthopyroxene. This layering is unlikely to be due to primary igneous crystal accumulation given plagioclase is present, and the rock has a negative Eu anomaly (see next section). Finally, two mafic xenoliths have major element compositions similar to high-pressure pyroxenes (83-158 and 85-106), with low Na₂O and high CaO contents relative to basaltic compositions. Both these analyses fit a pyroxene formula well (cation totals for both analyses cast to six oxygen atoms equal 4.03 and 3.98 for 83-158 and 85-106, respectively).

The two xenoliths with intermediate SiO₂ contents have relatively high Al₂O₃ contents (Table 7) and do not resemble common igneous rock types. Their high Al₂O₃ relative to Na₂O and CaO suggests a metasedimentary origin (see petrogenesis section). Both samples are layered, but only 85-101 appears to have been affected by metamorphic differentiation. Concentration of metamorphic garnet, which constitutes about 50% of the rock, is reflected by the high FeO, MnO, and MgO content of this sample.

The two felsic xenoliths have major element compositions similar to calc-alkaline rocks. Using normative mineralogies to classify these rocks [Streckeisen, 1976], 83-160 is a granodiorite and 83-162 is an adamellite; both are peraluminous and both crystallized in the late Paleozoic [Rudnick and Williams, 1987], a time of extensive calc-alkaline volcanism in this region of north Queensland (Figure 1; Oversby *et al.* [1980]). It is therefore useful to compare the chemistries of these xenoliths with those of the Paleozoic felsic igneous rocks [Sheraton and Labonne, 1978]. Sample 83-160 is characterized by low SiO₂ and K₂O, and high Al₂O₃, MgO, FeO, and CaO relative to the late Paleozoic granites; it falls within, but at the low SiO₂ end of, the major element differentiation trends of the late Paleozoic

calc-alkaline suite. In contrast, 83-162 does not fall along the major element differentiation trends of the late Paleozoic granites. For its SiO₂ content it has higher FeO, MnO, and MgO and lower Na₂O and CaO. These features are indicative of metamorphic concentration of garnet in this sample, which is suggested by elevated HREE (see below). Both felsic xenoliths have chemical affinities with the late Paleozoic granites.

Rare earth elements. Rare earth elements (REE) are immobile on a whole-rock scale during regional metamorphism, at low fluid/rock ratio. Consequently REE patterns in high-pressure granulites, which form under low activity of H₂O, are extremely useful petrogenetic indicators. REE for 12 McBride xenoliths are given in Table 8, and REE patterns are plotted in Figures 7 and 8. The samples are grouped according to similarity in REE patterns.

Figure 7a shows REE patterns for three mafic granulite xenoliths. All exhibit variable enrichments of Eu relative to other REE. As feldspar is the most efficient mineral at fractionating Eu²⁺ from REE³⁺, and because all these samples are plagioclase bearing, it seems likely that these xenoliths experienced plagioclase enrichment in an open system (i.e., either crystal accumulation or melt extraction). This interpretation is supported by the high Al₂O₃ contents of these xenoliths. Light rare earth element (LREE)/HREE ratios for these samples are variable. Samples 83-159 and 85-107 are both LREE enriched, whereas 85-114 has a peculiar pattern with a minimum in the middle REE, near Sm. This unusual HREE enrichment bears discussion. The only common rock-forming minerals capable of producing fractionations within the HREE (i.e., changing Gd/Yb ratios) are orthopyroxene or garnet [Irving, 1978; Fujimaki *et al.*, 1984]. However, the observations that REE concentrations in orthopyroxenes are very low compared with garnets and that garnet is the main ferromagnesian phase in this sample, suggest that it

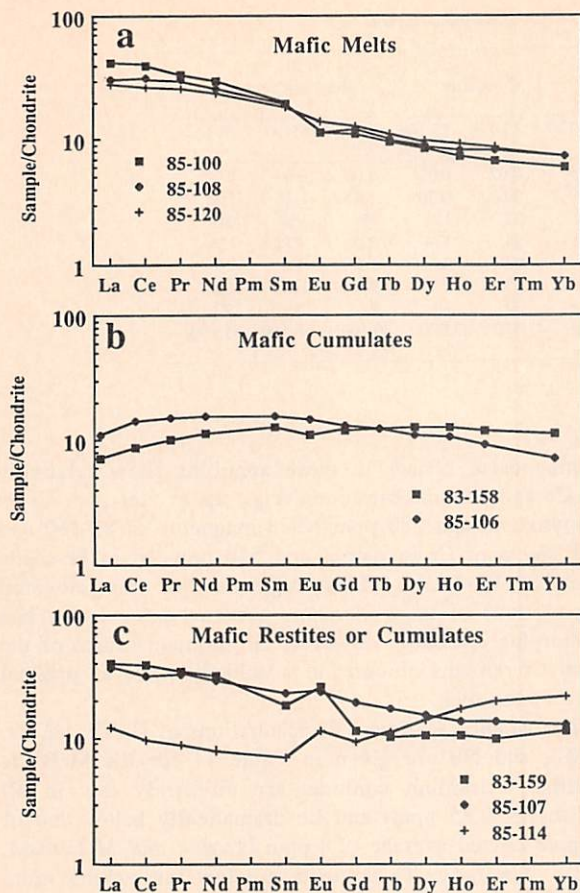


Fig. 7. REE patterns for Hill 32 mafic xenoliths: (a) cumulate or restitic xenoliths, and (b) pyroxene cumulate xenoliths, (C) possible mafic melt xenoliths.

is responsible for the HREE enrichment. Garnets in equilibrium with igneous melts have positive slopes in their HREE, similar to that of 85-114 [Irving and Frey, 1978]. The fractionation within the HREE, plus the xenolith's large size (~500 cm³, of which about one third was used for analysis) and massive texture, suggests the garnet enrichment occurred through crystal-melt interaction rather than metamorphic differentiation.

Figure 7b gives REE patterns of two mafic xenoliths which are LREE depleted; one without, and one with a slight negative Eu anomaly. As mentioned earlier, the major element compositions of these xenoliths both fit a pyroxene formula, and their REE patterns reflect the influence of pyroxene in an open magmatic system. The lack of a positive Eu anomaly suggests that the minor amount of modal plagioclase present either crystallized from an interstitial melt or exsolved from a relatively high-pressure pyroxene.

The REE patterns of the remaining mafic xenoliths are shown in Figure 7c. All display similar patterns: LREE enriched ((La/Yb)_N = 3.9-7.0), with or without negative Eu anomalies ((Eu/Eu*) = 0.75-0.93). (Note that although 85-100 has been enriched in orthopyroxene, it does not affect the whole-rock REE pattern due to low REE concentrations in orthopyroxene [Irving, 1978; Fujimaki et al., 1984].) Such REE patterns are similar to those found in continental basalts [Basaltic Volcanism Study Project (BVSP), 1981].

REE patterns of the two intermediate xenoliths are shown in Figure 8a. Sample 83-157 has an REE pattern similar to the average post-Archean Australian shale (PAAS), shown for reference, whereas that of 85-101 is HREE enriched, with a negative Eu anomaly. Unlike the previously discussed HREE enrichment in the mafic xenolith 85-114, the HREE are not fractionated relative to one another, but are uniformly elevated. This observation, coupled with the small size of the xenolith (Table 2) and the occurrence of garnet-rich layers, suggests that the REE pattern is the result of metamorphic concentration of garnet.

The two felsic xenoliths (83-160 and 83-162) have REE patterns characterized by LREE enrichments ((La/Yb)_N = 2.7-4.2, Figure 8b) and flat HREE. The slight concave upward REE pattern of 83-160 is similar to that of medium-K suite calc-alkaline volcanics [Gill, 1981]. Sample 83-162 has the largest negative Eu anomaly found in the suite (Eu/Eu* = 0.54), although this is in part a function of its elevated HREE, which will cause the calculated Eu/Eu* to be higher. The HREE enrichment is probably due to metamorphic concentration of garnet, as it is small and contains garnet-rich layers (Table 2).

Large cation elements. Table 9 gives concentrations for the elements Cs, Rb, Ba, Sr, and Pb. In general, the mafic xenoliths have low concentrations of Cs and Rb, and high K/Rb (630-2800) and K/Cs ratios (K/Cs = 30,000-83,000). Cs and Rb concentrations are considerably more variable in the intermediate to felsic xenoliths and are lower than those inferred for their respective protolith compositions. The range of K/Rb ratios exhibited by the entire xenolith suite is present in these rocks. K/Rb ratios range from typical upper crustal values (e.g., K/Rb = 142-263) for the felsic xenoliths to high ratios in the intermediate, metasedimentary xenoliths (K/Rb = 970-3,600). K/Cs ratios in the intermediate to felsic xenoliths range from 73,000 to 320,000.

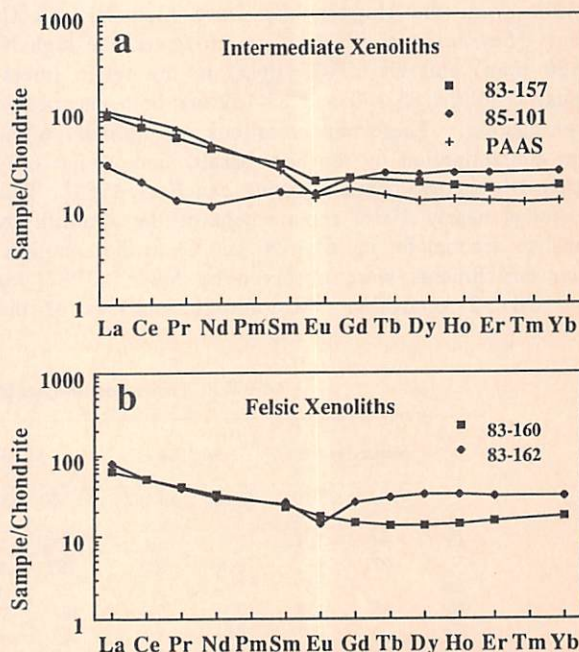


Fig. 8. REE patterns for Hill 32 felsic and intermediate xenoliths: (a) metasedimentary xenoliths and (b) felsic xenoliths.

TABLE 11. High-Valency Cation Contents of Hill 32 Xenoliths (ppm)

	Metasediments		Felsic Melts		Mafic Melts			Cumulates		Restites/Cumulates		
	83-157	85-101	83-160	83-162	85-100	85-108	85-120	83-158	85-106	83-159	85-107	85-114
U	0.37	0.15	0.85	0.37	0.18	0.08	0.04	0.01	0.05	0.09	0.63	0.13
Th	1.14	0.08	3.24	0.98	0.55	0.86	0.09	0.04	0.23	0.15	0.47	0.35
Th/U	3.08	0.53	3.81	2.65	3.06	10.7	2.25	4.00	4.60	1.67	0.75	2.69
Hf	9.09	1.78	5.75	6.54	3.19	1.31	2.15	1.11	2.11	7.21	33.9	2.50
Zr	381	77	199	206	107	44	63	45	65	348	1608	93
Zr/Hf	42	43	35	31	34	34	29	41	31	48	47	37
Sn	1.7	1.0	2.8	2.5	1.6	0.8	2.0	1.5	2.2	2.6	7.6	1.2
Mo	1.5	0.8	1.9	0.9	0.5	0.7	0.4	0.4	0.4	1.0	0.8	1.1
Nb	6.6	20	8.1	16	8.9	4.4	14	1.4	3.4	15	21	35
La/Nb	5.0	0.5	2.8	1.8	1.7	2.6	0.7	1.9	1.2	1.0	0.6	0.1

into three genetic types: (1) xenoliths which originated as nearly "unmodified" basaltic melts, (2) xenoliths with the chemical features of crystal cumulates, and (3) xenoliths with chemical features of mafic residues left after partial melt extraction. Distinguishing between crystal cumulates and residua is often difficult in the absence of primary igneous textures; discussion of possible criteria has been given by Rudnick *et al.* [1986]. One common approach is that originally proposed by Frey and Prinz [1978], where they distinguish cumulates from restites by the variation in compatible and incompatible trace element concentrations in a genetically related suite of rocks. They noted that at variable but low degrees of partial melting, incompatible trace elements will change dramatically in the residue, but compatible trace elements will show little variation. The reverse is true for crystal cumulates. Overall, this approach works well for genetically related rocks, but unfortunately in the McBride xenoliths, as in many other xenolith suites, no genetic link can be presumed and this approach is limited. However, a similar approach based upon phase petrology can be useful. For example, if a xenolith has chemical features pointing to the importance of a given phase during the xenolith's generation (such as igneous garnet or clinopyroxene), then the position of this phase on the liquidus of the presumed melt type can help to determine whether that phase was a cumulate or restite. For this approach to be valid, one must assume, as in the case of trace element variations, that only small degrees of partial melting or crystal accumulation occur. Below we outline the rationale behind our preferred origins for these xenoliths.

Only one mafic xenolith, 85-120, has major and trace element composition similar to a basaltic melt. It has chemical features of an oversaturated, low Ti tholeiite. Rare zircons found within this xenolith, with late Paleozoic ages, may be xenocrystic, or may have crystallized during the metamorphism [Rudnick and Williams, 1987]. Two other mafic xenoliths (85-100 and 85-108) may have originated as mafic melts. Both have REE patterns similar to 85-120 (Figure 7c) and both are silica oversaturated. However, 85-108 has a low Na₂O and a high CaO content for a mafic melt, implying possibly some clinopyroxene accumulation, and 85-100 experienced orthopyroxene enrichment through metamorphic differentiation, as discussed above. Zircons in 85-100 yield a range of concordant ages between 400 and 250 Ma which are

interpreted to reflect intrusion at 400 Ma followed by slow cooling and metamorphism in the deep crust. In contrast, zircons in 85-120 yield concordant ages between 320 and 200 Ma, and thus this mafic melt probably intruded at 320 Ma, at the beginning of the last major granite-forming event [Rudnick and Williams, 1987]. Primitive mantle-normalized trace element abundances for these meltlike xenoliths are shown in Figure 9. Large ion lithophile elements, which may have been depleted by metamorphism (see below), are not plotted. All three samples are depleted in the high field strength cations Zr, Hf, and Ti relative to the REE. However, Nb is only depleted in 85-100 and 85-108; it is slightly enriched in 85-120. Sr is depleted in 85-100 and 85-108, correlating with large negative Eu anomalies in these samples ($Eu/Eu^* = 0.75-0.78$), and suggests plagioclase fractionation from the melt. All of these chemical features (except the lack of Nb depletion in 85-120) are similar to those of either island arc basalts or continental flood basalts.

Two mafic xenoliths, 83-158 and 85-106, have major element compositions similar to high-pressure pyroxenes and REE patterns characterized by LREE depletions. It is likely that these xenoliths originated as high-pressure pyroxene cumulates from a mafic melt. The modal garnet and plagioclase would thus represent later exsolved phases, or possibly phases crystallized from interstitial melt. Garnet and plagioclase exsolution lamellae in the

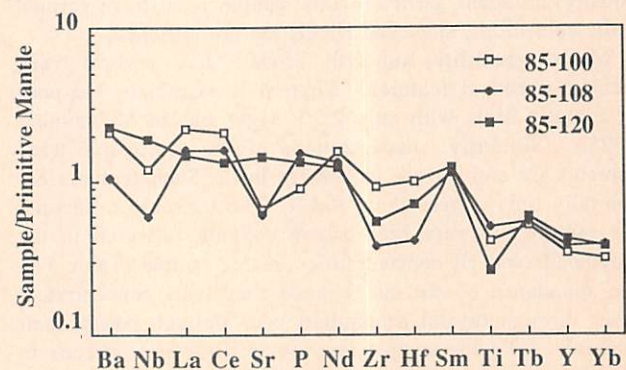


Fig. 9. Primitive mantle-normalized trace element abundances in the three mafic xenoliths with melt-like compositions. Note that Sr depletion in 85-100 and 85-108 correlates with a negative Eu anomaly in these two samples. Primitive mantle normalizing values from McDonough *et al.* [1985].

clinopyroxenes of 83-158 provide textural evidence for the former. Other trace element characteristics such as high Cr, Ni, Sc, and V and low Sr and Ba concentrations are consistent with this interpretation. The abundance of clinopyroxene in these xenoliths suggests they are not restites, but rather are clinopyroxene cumulates, since pyroxene is an early crystallizing phase in many basalt types [BVSP, 1981; Green, 1980].

Finally, three mafic xenoliths, 83-159, 85-107, and 85-114 have chemical characteristics indicative of a cumulate or restite origin. All have high Al_2O_3 contents and positive Eu anomalies.

Sample 85-114 is large, and unlayered, and shows the obvious effects of plagioclase and garnet concentration in its REE pattern (Figure 7a). The lack of mineralogical layering implies that these features are not due to metamorphic differentiation. Plagioclase and garnet only coexist at near-solidus temperatures for basalts and andesites (see review of experimental data by Green [1980]), suggesting this sample is a restite rather than a cumulate. The high Ti and Nb content of this sample also shows that rutile (which is present in the mode), or possibly another Ti-rich phase, was an important residual phase during the partial melting event.

The trace element chemistry of sample 83-159 does not allow direct distinction between a restite or cumulate origin, but suggests that plagioclase was an important phase during the generation of this sample. However, when its chemical characteristics are combined with observations on the ages of zircon and the types of inclusions within them, a restite origin is suggested. Several zircons from 83-159 contain recrystallized melt inclusions of approximate intermediate composition ($\text{SiO}_2 \approx 63\%$), and define a chord between 1580 and 200 Ma on a concordia diagram [Rudnick and Williams, 1987]. The presence of euhedral amphibole inclusions within the zircons further suggests crystallization from an intermediate parent melt containing significant amounts of H_2O (at least 3 wt. %, based on experiments of Naney [1983]). The simplest model to explain these observations is that the zircons crystallized from an intermediate melt during the Proterozoic. This rock was then partially melted during high-grade metamorphism in the late Paleozoic, leaving a mafic residue of plagioclase, ferromagnesian minerals (possibly amphibole or clinopyroxene), and zircon; the modally abundant garnet in this sample must have formed in the subsolidus, since the HREE are not enriched.

Mafic granulite xenolith 85-107 has several very peculiar chemical features. First, it is extremely Mg-poor for a mafic rock, with only 2.2% MgO and an Mg number of 21. Similarly, concentrations of the compatible trace elements Cr and Ni are extremely low. Such features are generally only observed in felsic igneous rocks. Second, this sample has very high zircon content, reflected in the extreme Zr and Hf concentrations in this sample (Table 11). The abundance of zircons suggests they were concentrated, either through crystal accumulation or through partial melt removal. Like the sample previously discussed, zircons in this rock contain inclusions of trapped melt. However, in these zircons the melt has not been recrystallized and yields a homogeneous granitic composition, with a low Mg number (~25) [Rudnick and Williams, 1987]. All of these observations suggest that 85-107 represents a mafic

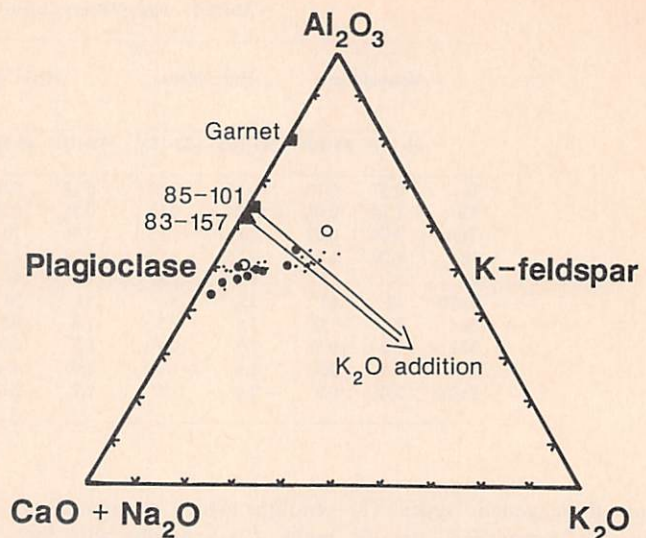


Fig. 10. Chemical index of alteration (CIA) diagram, with oxides plotted as molecular proportions [after Nesbitt and Young, 1982, 1984]. Mineral compositions and field of igneous rocks are shown for reference. Solid circles represent average igneous rock types for andesitic through granitic compositions [Nesbitt and Young, 1984; Taylor and McLennan, 1985]. Dots represent late Paleozoic and Proterozoic north Queensland felsic igneous rocks [Sheraton and Labonne, 1978]. Both 83-157 and 85-101 have significantly higher Al_2O_3 than typical igneous rocks, suggesting both have experienced weathering. Arrows show trajectory upon which rocks may have originally plotted if granulite facies metamorphism has caused depletion in K_2O . However, in order for both rocks to have originally plotted within the igneous field, 3.3% K_2O (for 83-157) or 5.5% K_2O (for 85-101) would have to have been lost. Open circles represent the two felsic Hill 32 xenoliths. Note that metamorphic garnet concentration in 83-162 causes it to plot at slightly higher $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O})$ than typical igneous rocks.

cumulate from a granitic magma. The complete overlap in metamorphic and igneous zircon ages for this rock [Rudnick and Williams, 1987] supports a model whereby crystal accumulation occurred from a granitic melt in a deep crustal environment. The cumulates remained in the deep crust, where they recrystallized as they cooled.

Intermediate xenoliths. The major and trace element chemistry and mineralogy (sillimanite in 85-101) of the intermediate xenoliths indicate a sedimentary origin for these rocks. The high Al_2O_3 contents relative to Na_2O and CaO suggest these samples have been weathered. This is best demonstrated using the CIA (chemical index of alteration) triangular diagram, developed by Nesbitt and Young [1982, 1984] to document the effects of chemical weathering. Both samples fall near the plagioclase-kaolinite join, far above the field of unweathered igneous rock types (Figure 10). Several caveats must be considered when using this diagram for metamorphic rocks. First, CaO in the diagram is meant to be the CaO in the silicate portion of the rock; therefore in sedimentary rocks, CaO in carbonate and apatite are subtracted. Because it is difficult to determine how much, if any, carbonate and apatite may have originally been present in these rocks, all CaO is taken as silicate CaO, making this value a maximum. Second, granulite facies metamorphism may cause depletion in K_2O , which is difficult to effectively constrain (see below and Rudnick et al., [1985]), so that the rocks may have originally plotted along a line toward the K end of the

triangle (Figure 10). However, the amount of K_2O that would have to have been removed from these rocks during metamorphism, if they originally plotted within the igneous rock trend, is unreasonably large (3 wt. % for 83-157 and 5 wt. % for 85-101). Finally, it is apparent from the REE chemistry that 85-101 probably has elevated garnet concentrations due to unrepresentative sampling. Its relatively high Sc content relative to typical post-Archean shales (36 ppm versus 10-20 ppm, respectively [Taylor and McLennan, 1985]) supports this, since Sc partitions into garnet [Irving and Frey, 1978]. Garnet addition must occur at the expense of plagioclase and quartz (the other important modal phases in this rock) so the sample would have originally plotted further away from the Al_2O_3 end of the triangle (Figure 10). However, the whole-rock composition would not plot within the unweathered igneous rock field unless all garnet were removed, and this seems unlikely.

In addition to major elements, the REE pattern of 83-157 is like those of post-Archean shales (Figure 8a). Similar REE patterns have been observed in metasedimentary xenoliths from Kilbourne Hole, New Mexico [Wandless and Padovani, 1985]. Even though the REE pattern of 85-101 shows the effects of metamorphic garnet enrichment, this sample still has a negative Eu anomaly, typical of those in post-Archean shales. We therefore conclude that both of these rocks were originally metapelites.

Felsic Xenoliths. The two felsic xenoliths have major and trace element characteristics similar to calc-alkaline igneous rocks. These features, coupled with their late Paleozoic crystallization ages, suggest they are related to the extensive late Paleozoic calc-alkaline igneous rocks which outcrop near Hill 32 (Figure 1). These samples are also shown on the CIA diagram (Figure 10). Sample 83-160 clearly plots within the field of igneous rocks, but 83-162 has higher $Al_2O_3/(CaO + Na_2O)$ than igneous rocks. Again, preferential sampling of garnet is the likely cause of this. If only 10% of garnet is subtracted from this sample, then 83-162 plots at the high K_2O end of the igneous rock trend. Because both samples have high-grade mineralogies, indicating equilibration at deep crustal levels, it is likely that these rocks formed by partial melting of the lower crust and have not moved far from their source regions.

In summary, the McBride xenoliths represent a diverse lithologic assemblage with no apparent genetic link among any of the xenoliths. Most are mafic, but these mafic xenoliths arose from diverse processes, including intrusion and crystallization of mafic mantle-derived melts into the lower crust, crystal accumulation from mafic and felsic melts, and partial melting of intermediate rock types, leaving a mafic residue. The intermediate xenoliths are supracrustal in origin and the felsic xenoliths represent melts which formed and crystallized in the deep crust.

Metamorphic Effects

High-grade metamorphism has the potential of depleting rocks in large, incompatible elements such as Rb, (Cs), Th, and U [Heier, 1973]. Studies of rocks from granulite facies terrains show that (1) not all granulites are depleted and (2) such depletions, where they do occur, are not coupled with other geochemical indicators of partial melting (e.g., the

presence of positive Eu anomalies), suggesting that the depletions are due to metamorphic mineralogical changes coupled with fluid fluxing [Rudnick et al., 1985].

In order to evaluate the effects of granulite facies metamorphism on depletion of large ion lithophile elements (LILE) in granulite facies xenoliths, the petrogenesis of the rocks must be considered. For example, depletions in LILEs and an increase in K/Rb ratios may arise from igneous processes such as crystal accumulation and partial melt extraction. This is because K is less incompatible than Rb in all igneous minerals excluding mica. In these cases, the degree of resulting depletion depends strongly on the proportion and composition of interstitial melt. In addition, preferential sampling of ferromagnesian phases in layered xenoliths will cause relative depletion of these elements and may change their ratios. Therefore it is difficult to constrain LILE depletions due to metamorphism in samples which have been generated or affected by these processes.

The mafic xenoliths from the McBride province have low overall abundances of LILEs and high K/Rb and K/Cs ratios; however, most of these xenoliths formed as either cumulates or restites. Only three of the mafic xenoliths from the McBride province represent possible crystallized melt compositions, and one of these is enriched in orthopyroxene. The two remaining xenoliths (85-108 and 85-120) have very high K/Rb ratios and low overall K and Rb abundances, features which suggest metamorphic depletions of both K and Rb. Rb/Cs ratios for these and all mafic granulite samples are near 30, similar to Rb/Cs of basaltic rocks (W. F. McDonough et al., Rb and Cs in the earth and moon and the evolution of the earth's mantle, submitted to *Earth and Planetary Science Letters*, 1987). However, the very low Rb and Cs concentrations suggest that both elements were severely depleted. The Th and U contents of these samples are also very low for continental tholeiites [cf. BVSP, 1981], and one sample (85-108) has a very high Th/U ratio, suggesting U depletion relative to Th.

The intermediate and felsic xenoliths have formed through "supracrustal" processes such as sedimentation and intracrustal melting, thus allowing comparison of their LILE concentrations with assumed protolith compositions. The two intermediate xenoliths have low K contents for sedimentary rocks and their K/Rb ratios are very high (Table 9). One of these samples is strongly enriched in metamorphic garnet and thus its LILE concentrations may not reflect the true bulk-rock value if a larger sample had been available. However, these difficulties are overcome with the second sample, and it appears to have been depleted in Rb, and perhaps K, relative to a sedimentary protolith. The Th and U contents of this sample are also very low, but its Th/U ratio is similar to or somewhat lower than that of shales, suggesting proportional depletion of both these elements. In contrast with the intermediate xenoliths, the two felsic xenoliths have relatively high K contents and K/Rb ratios similar to the upper crust. There is no evidence of K or Rb depletion in these samples. It is interesting that even though one sample has elevated garnet (83-162), it has only slightly affected the K and Rb contents, and has apparently not fractionated one from the other. Th and U contents of these samples are up to an order of magnitude lower than in

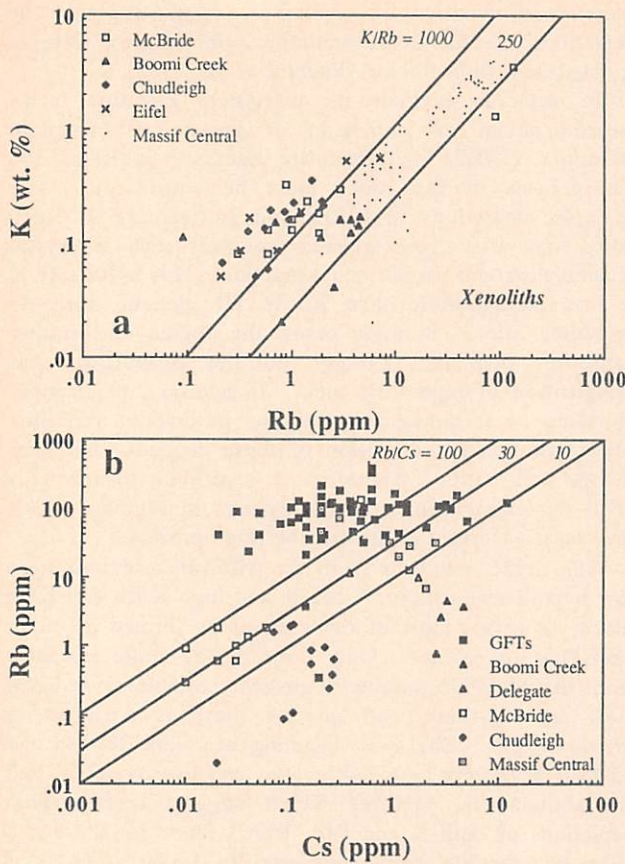


Fig. 11. (a) Rb versus K for lower crustal xenoliths, excluding xenoliths carried in kimberlites. Data sources are McBride, north Queensland (this paper); Boomi Creek, New South Wales [Wilkinson, 1975; Wilkinson and Taylor, 1980]; Chudleigh, north Queensland [Rudnick et al., 1986]; Eifel, west Germany [Stosch et al., 1986]; Massif Central, France [Leyreloup et al., 1977; Dupuy et al., 1979]. (b) Cs versus Rb for lower crustal xenoliths and granulite terrains. Data sources for xenoliths are the same as in Figure 11a; data sources for granulite terrains are Heier and Brunfelt [1970]; Sighinolfi et al. [1981]; Condie et al. [1986]; Taylor et al. [1986]; McCulloch et al. [1987].

typical granitic rocks [cf. Rogers and Adams, 1978], but Th/U ratios are only slightly lower than those of granites (2.6 and 3.6 versus 3.5-6.0 for granitic rocks), which suggests proportionately equal depletion in both these elements. In contrast with the mafic xenoliths, Rb/Cs and K/Cs ratios in the intermediate to felsic xenoliths are quite high, indicative of Cs depletion relative to Rb and K.

Figure 11 shows the Rb-K and Rb-Cs characteristics of the McBride xenoliths relative to data for granulite facies xenoliths and granulite terrains. Xenoliths entrained in kimberlites have been left out of these plots because of the effects of possible contamination from the hosts [Griffin et al., 1979; Rogers and Hawkesworth, 1982]. The McBride xenoliths show the largest range in K and K/Rb ratios yet reported within a single suite of granulite xenoliths, although xenoliths from the Massif Central show nearly as large of range (Figure 11a). The range and anticorrelation of K/Rb ratio with K content in the xenoliths are similar to that observed in rocks from granulite facies terrains [Rudnick et al., 1985]. In contrast, the variation in Rb/Cs ratios of granulites from terrains is quite distinct from that of xenoliths (Figure 11b). Terrain granulites tend to have

higher Rb concentrations and Rb/Cs ratios than granulite xenoliths. This variation probably reflects the differences in bulk-rock composition between these two lower crustal rock types and the importance of K-feldspar in the partitioning of Rb and Cs in granulites. All available Cs analyses for terrain granulites are for intermediate to felsic rocks, most of which contain K-feldspar. In contrast, most granulite xenoliths are mafic and do not contain K-feldspar. The K_D^{Cs} between K-feldspar and saline aqueous fluid is up to an order of magnitude lower than that of Rb [Carron and Lagache, 1980]; thus exchange between K-feldspar and fluid will greatly increase the Rb/Cs ratio of the solid. In contrast, mineral phases in mafic granulites equally exclude Rb and Cs so that exchange between a fluid and a mafic granulite will cause severe depletions of these elements but will not fractionate one from the other. This is supported by the variations observed in Rb/Cs ratios between the mafic and intermediate/felsic lithologies in the McBride xenoliths (Table 9).

Figure 12 shows Th-U variations for granulite xenoliths. Such data are scarce, but the available analyses show similar trends to those of rocks from granulite terrains, albeit at lower concentrations. This again reflects the dominance of mafic lithologies in xenoliths when compared with granulite terrains. Th/U ratios in xenoliths are similar to those of granulite terrain rocks of similar Th and U concentrations. The similarities in LILE contents and LILE ratios between granulite xenoliths and rocks from granulite terrains may reflect similar depletion processes, the nature of which is as yet poorly understood.

A Ba- and Sr-Rich Lower Crust?

Previous studies found high Ba and Sr contents in lower crustal xenoliths [Arculus et al., 1987] and have postulated a Ba- and Sr-enriched lower crust in island arc environments [Arculus and Johnson, 1981]. The McBride province xenoliths are not enriched in Ba, and Ba/La ratios are not extreme (Table 9). In fact, whereas the Sr contents of the mafic xenoliths are similar to those of common basalts, the Ba contents are generally lower than found in such rocks. This is probably due to the effects of partial melt extraction and crystal accumulation, during which Ba is highly depleted and Sr may or may not be depleted,

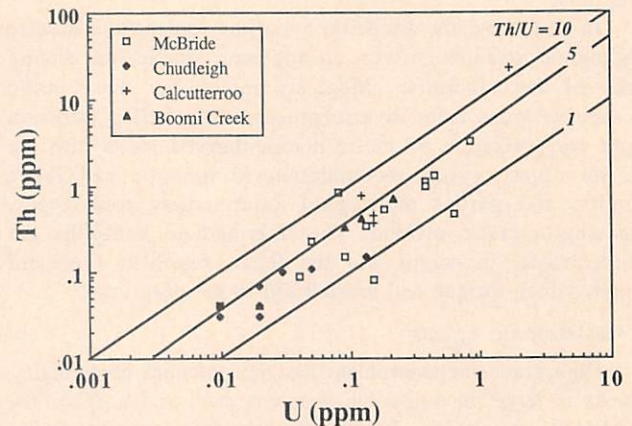


Fig. 12. Th versus U for Hill 32 xenoliths, excluding xenoliths carried in kimberlites. Data sources: McBride, north Queensland (this paper); Chudleigh, north Queensland [Rudnick et al., 1986]; Boomi Creek, New South Wales [Wilkinson and Taylor, 1980].

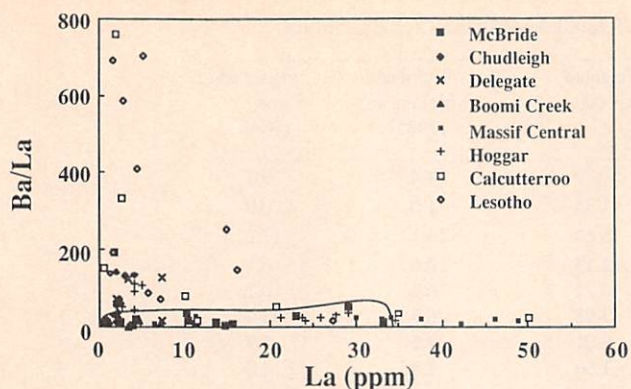


Fig. 13. La versus Ba/La ratio for lower crustal xenoliths. Open symbols are xenoliths contained within kimberlitic hosts; solid symbols are xenoliths hosted by alkalic basalts or nephelinites. Sources of data: McBride, north Queensland (this paper); Chudleigh, north Queensland [Rudnick *et al.*, 1986], Delegate, New South Wales (R. J. Arculus, unpublished data, 1986); Boomi Creek, New South Wales [Wilkinson and Taylor, 1980], Massif Central, France [Dupuy *et al.* 1979], Hoggar, Algeria [Leyreloup *et al.*, 1982], Calcutteroo, South Australia [Arculus *et al.*, 1986], Lesotho, southern Africa [Rogers, 1977; Griffin *et al.*, 1979; Rogers and Hawkesworth, 1982].

depending upon the presence of plagioclase. REE data confirm that both processes have influenced the composition of these xenoliths.

Figure 13 shows the variation of La versus Ba/La for all available data on lower crustal xenoliths. The field of the McBride province xenoliths is outlined. Several important features are apparent from this diagram. First, it is only xenoliths which are carried by kimberlites or other alkaline earth-rich hosts (e.g., Chino Valley latite [Arculus *et al.*, 1987]) that have extreme Ba/La ratios. Second, high Ba/La ratios only occur at low La contents. Both features indicate that the high Ba/La ratios observed in some lower crustal xenoliths may be artifacts of indirect contamination by the host kimberlite. Kimberlites are noted for their high alkali contents, and unlike the REE, many of these elements are readily leached during weathering and groundwater percolation [Dawson, 1980]. Recent analyses of garnets from ultramafic xenoliths carried in kimberlites show them to have a calcium carbonate phase high in Sr, but with insignificant Rb and REE contents (Ba was not measured) [Richardson *et al.*, 1985]. This contaminant was regarded as a precipitate from groundwater. This is in agreement with microprobe data that show that the Ba and Sr in Ba- and Sr-rich lower crustal xenoliths are contained in grain boundary strontio-barytes [Emery *et al.*, 1985]. We conclude that, until further evidence becomes available, the Ba and Sr enrichments seen in some lower crustal xenoliths from kimberlites should be viewed with caution and should not be considered a common feature of lower crustal xenoliths. A similar conclusion has been reached by R. J. Arculus (personal communication, 1986).

Comparison With Other Xenolith Suites

The wide range of lithologic compositions which are present in the McBride xenoliths are unusual in eastern Australia [Griffin and O'Reilly, 1986]. For example, just 150 km to the southwest, in the Chudleigh volcanic province, lower crustal xenoliths are exclusively mafic (only one felsic xenolith was found out of a total of 77 xenoliths collected by us [Rudnick *et al.*, 1986; Kay and

Kay, 1983]. Other eastern Australian localities contain only rare intermediate to felsic lithologies (e.g., charnockite xenoliths at Delegate [Lovering and White, 1969]), but such lithologies are apparently more abundant in the South Australian kimberlitic pipes which erupt through Proterozoic crust (e.g., Calcutteroo pipe [Ferguson *et al.*, 1979]).

Does the scarcity of reported intermediate to felsic granulite xenoliths in eastern Australian volcanics reflect an entirely mafic lower crust, or is it possible that intermediate to felsic lithologies are underrepresented, either due to preferential sampling and preservation by the basaltic host or preferential sampling by investigators? We suggest the latter is true and that both factors may play a role. For example, crustal xenoliths in many of the Victorian Newer basalt vents are restricted to garnet-sillimanite feldspar-quartz compositions, which are so extensively melted that they have a texture and density approaching pumice. (These occur at Mt. Leura, Mt. Noorat, and Mt. Porndon [McDonough *et al.*, 1987b].) Similar rocks have been observed in the Atherton volcanics of north Queensland. Such rocks are inherently difficult to work on and have yet to be described in the literature on eastern Australian xenolith occurrences. In addition, intermediate to felsic lower crust may exist beneath localities at which no such xenoliths are observed. For example, isotopic compositions of the mafic Chudleigh xenoliths require assimilation of a more evolved crustal component by these rocks [Rudnick *et al.*, 1986]. Determining whether other suites of exclusively mafic lithologies show similar crustal signatures or not requires accumulation of more isotopic data.

The above observations suggest that intermediate and felsic lithologies may be more abundant in the eastern Australian lower crust than recent estimates indicate. A compositionally heterogeneous lower crust is a reasonable expectation given the diverse processes which form the lower crust at continental margins (i.e., basaltic underplating, tectonic underplating of supracrustals and partial melting). Therefore the lithologies present in the McBride suite are regarded as representative of at least this portion of the lower crust, and it is illustrative to compare their compositions with recent estimates for the bulk composition of the lower crust.

Lower Crustal Composition

An average composition of the north Queensland lower crust was obtained by weighting the average compositions of intermediate/felsic xenoliths and mafic xenoliths according to their observed lithologic proportions (column 2 of Table 1). Table 12 provides the mean intermediate/felsic and mafic xenolith compositions and the weighted mean for the suite. Also given are recent estimates of lower crustal composition from Weaver and Tarney [1984] and Taylor and McLennan [1985]. In averages of this type, where only a small number of samples are available, a single sample with extreme chemistry can significantly affect the mean. Such is the case with Mg-poor mafic xenolith 85-107, which has extreme Zr and Hf contents due to its high zircon content. However, recent experimental and geochemical studies of granites have suggested that such accessory phase-rich rocks may occur in the lower crust [Watson and Harrison,

TABLE 12. Average Composition of McBride Xenoliths and Lower Crustal Estimates

	Mean Intermediate Felsic	Mean Mafic	Weighted Average	Taylor and McLennan [1985]	Weaver and Tamey [1984]
SiO ₂	60.37	47.93	50.29	54.4	59.2
TiO ₂	0.88	1.46	1.35	1.0	0.9
Al ₂ O ₃	17.42	16.46	16.64	16.1	17.2
FeO	8.59	12.95	12.13	10.6	6.1
MnO	0.24	0.22	0.22	0.2	0.1
MgO	4.33	8.84	7.98	6.3	3.4
CaO	3.62	10.27	9.01	8.5	5.9
Na ₂ O	2.24	1.45	1.60	2.8	4.0
K ₂ O	1.64	0.17	0.45	0.34	2.4
Sc	28	41	33	36	
V	133	237	217	285	
Cr	82	322	276	235	45
Co	26	39	31	35	
Ni	80	156	141	135	27
Cu	53	24	29	90	
Rb	57	1.1	11.8	5.3	66
Sr	213	193	196	230	601
Y	46	24	28	19	15
Zr	216	297(109)	281(127)	70	181
Nb	13	13	13	6	10
Mo	1.3	0.7	0.8	0.8	
Sn	2.0	2.4	2.3	1.5	
Cs	0.25	0.03	0.07	0.1	
Ba	743	87	212	150	605
La	24	9.6	12.3	11	29
Ce	45	24	28	23	61
Pr	4.95	3.3	3.6	2.8	
Nd	20	15	16	12.7	23
Sm	5.33	3.84	4.1	3.17	4.9
Eu	1.41	1.35	1.36	1.17	1.29
Gd	6.2	3.86	4.31	3.13	
Tb	1.29	0.67	0.79	0.59	0.65
Dy	8.14	4.33	5.05	3.60	
Ho	1.85	0.95	1.12	0.77	
Er	5.34	2.76	3.25	2.20	
Yb	5.54	2.64	3.19	2.20	1.50
Hf	5.79	6.69(2.81)	6.51(3.32)	2.10	5.80
W	0.72	0.46	0.51	0.70	
Pb	6.5	2.5	3.3	4.0	12
Th	1.4	0.34	0.54	1.06	6.0
U	0.4	0.15	0.21	0.28	1.25

Blanks indicate value not reported. Zr and Hf numbers in parentheses are means excluding sample 85-107, which has extreme concentrations of these elements.

1983; *Sawka and Chappell*, 1986]. The question therefore is whether the proportion of this rock type observed here (i.e., 1 in 12) is accurate. For this reason, Zr and Hf values have been calculated with and without this sample. A more realistic estimate of average Zr and Hf in the north Queensland lower crust may lie between these two values.

A convenient way in which to compare the lower crust compositional models is by normalizing them to the upper crustal composition of *Taylor and McLennan* [1985]. By plotting the data in this fashion, two features can be readily evaluated: (1) the variations between models and (2) the degree of fractionation between the upper and lower continental crust (i.e., elements with a ratio of < 1.0 are depleted in the lower crust compared with the upper crust).

Considering the uncertainties associated with this xenolith-derived composition, similarities between it and the lower crustal model of *Taylor and McLennan* [1985]

appear remarkable. The largest discrepancy occurs in some of the high-field strength elements and in the alkali elements; Nb, Zr, Hf, and Rb concentrations are higher in the xenolith average than in the proposed lower crust model, and Cs, Th, and U are dramatically lower in the xenoliths than the lower crustal model. In addition, Na and Cu are significantly depleted in the xenoliths with respect to the lower crustal model. Taylor and McLennan's lower crust was modeled as the complement left after 25% of upper crust was subtracted from their total crust composition. The granodioritic composition of the upper crust, coupled with the presence of an average negative Eu anomaly, led them to hypothesize that intracrustal melting is the dominant process by which the crust differentiates. Their model thus predicts that a large proportion of the lower crust is residual material. Interestingly, while such material is present in the McBride suite, it forms only a

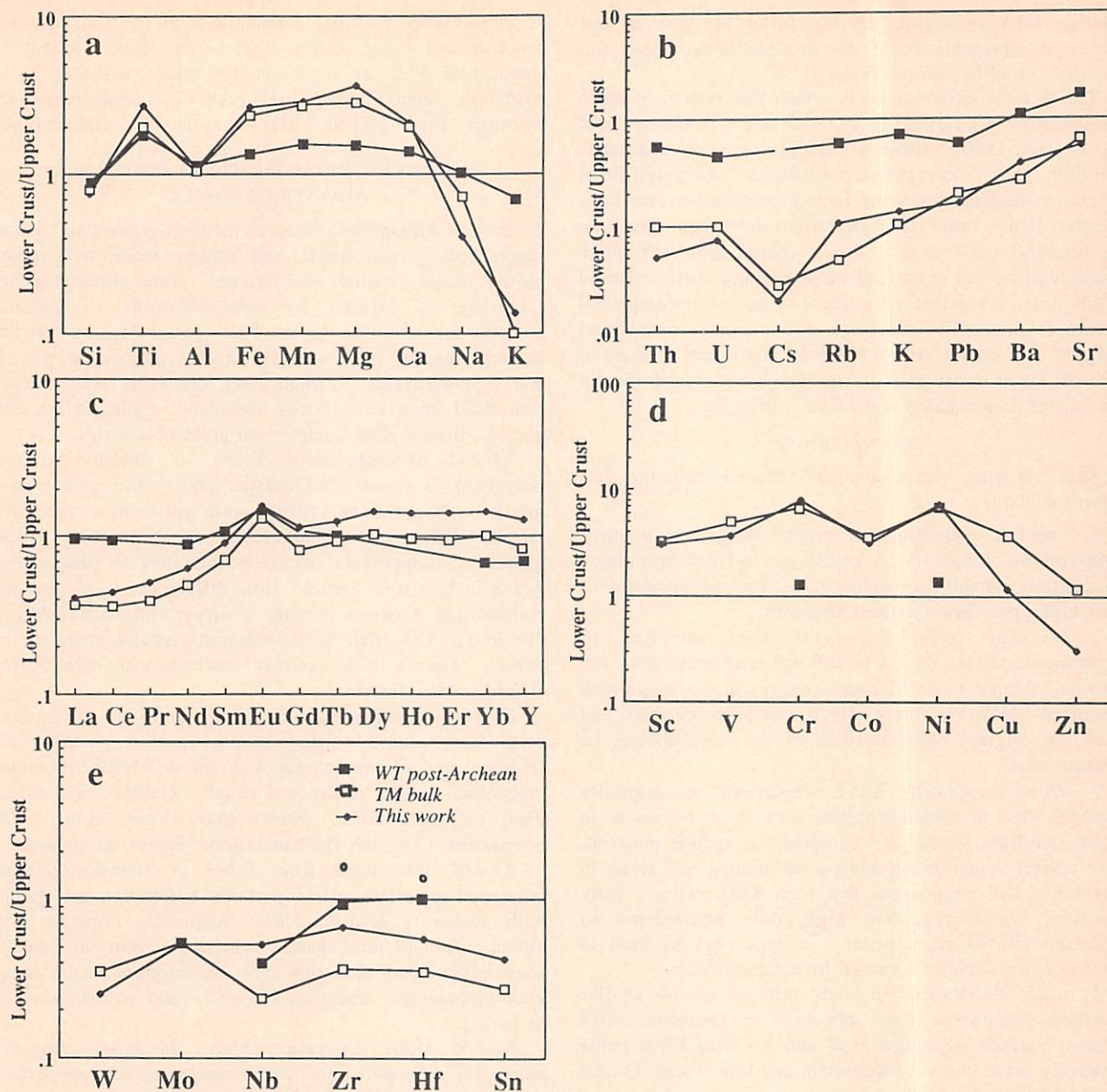


Fig. 14. Lower crustal compositional models normalized to upper crust of Taylor and McLennan [1985]. (a) Major elements, (b) alkali elements and Th and U, (c) REE and Y, (d) transition metals, and (e) high field strength cations. Zr and Hf abundances are shown with 85-107 included (open symbols, plotting above the line = 1.0) and with 85-107 excluded (solid symbols).

minor proportion of the observed lithologies. Therefore although the compositions are similar, the processes giving rise to the lower crust are not. This suggests that cumulate processes may be more important in intracrustal differentiation and crustal evolution than supposed by Taylor and McLennan [1985].

The large differences between the McBride average composition and the post-Archean lower crust proposed by Weaver and Tarney [1984] are attributable to their use of average orogenic andesite as representative of the post-Archean lower crust. The xenoliths from this province and elsewhere are, on average, much more mafic than this. Thus ferromagnesian elements are higher in the xenoliths (2-6 times, Figure 14d), and alkali elements dramatically

lower (0.1-0.2 times, Figure 14b) than orogenic andesite. Similar findings have been reported by Kay and Kay [1985] in their studies of the Aleutian island arc, where the lower crust is dominated by mafic cumulates derived from olivine tholeiite magmas. It is also noteworthy that the degree of LREE enrichment observed in the xenoliths is much lower than that postulated by Weaver and Tarney [1984], suggesting that REE are significantly fractionated between the upper and lower crusts. If the average lower crust composition from Weaver and Tarney's model is used (i.e., a mixture of Archean and post-Archean lower crusts), then the discrepancies between the model composition and the xenolith-derived composition become even more severe. This reflects the major compositional differences

between felsic granulite terrains (used by Weaver and Tarney as representative of the Archean lower crust) and granulite xenolith compositions.

The tectonic environment in which this section of north Queensland crust formed is under debate [Henderson and Stephenson, 1980]. However, the presence of voluminous 270-320 Ma calc-alkaline volcanics and high-level intrusions suggests that some form of subduction may have occurred at this time. It is clear from the zircon ages from the xenoliths that extensive lower crustal growth, through basalt addition and reworking of preexisting crust, occurred at this time. Therefore the regions where most continental growth is presently occurring, at convergent continental margins, are characterized by dominantly mafic, mantle-derived lower crust with tectonically and magmatically intercalated intermediate and felsic lithologies.

CONCLUSIONS

The following conclusions are reached from the data presented above.

1. Careful evaluation of major and trace element compositions, coupled with petrologic and geochronologic information, enables recognition of the petrogenesis of most high-grade lower crustal xenoliths.

2. In some xenoliths, preferential sampling of metamorphic phases due to metamorphic differentiation and the small sample size will cause changes in the whole-rock chemistry. However, these effects can be recognized, and often the original characteristics of the xenolith can be reconstructed.

3. When evaluating LILE depletions in granulite xenoliths due to metamorphism, care must be taken to avoid xenoliths which are cumulate or restite material, since crystal accumulation and partial melting will result in very low LILE abundances and high K/Rb ratios. Only xenoliths which represent high-grade equivalents to unmetamorphosed supracrustal rock types can be used to evaluate LILE depletions caused by metamorphism.

4. LILE abundances in such xenoliths show similar depletion patterns to those observed in granulite facies terrains: variable depletion in K and Rb with K/Rb ratios negatively correlated with K content and low overall Th and U contents with variable Th/U ratios.

5. A Ba- and Sr-rich lower crust is not supported by the McBride data, and only xenoliths carried by kimberlites show such enrichments. It is likely that these enrichments are due to contamination by the host kimberlite and do not represent an *in situ* lower crustal feature.

6. The lower crustal xenoliths from Hill 32 formed through a variety of processes: crystallization of mantle-derived basaltic melts, crystallization of locally derived felsic melts, crystal accumulation from mafic and felsic magmas, high-grade metamorphism of supracrustal lithologies, and partial melt extraction. Such a diverse lithologic assemblage reflects the heterogeneous nature of the lower crust in Phanerozoic continental margins.

7. A weighted mean composition of the lower crust beneath the McBride province, based upon the xenoliths from Hill 32, shows that the lower crust is mafic, with a major and trace element composition similar to that proposed by Taylor and McLennan [1985].

8. Simple models for lower crustal genesis in which the lower crust forms primarily through magmatic underplating

or intracrustal melting cannot account for all of the xenolith data. It is likely that both these processes operated as well as depletion of LILE (including heat-producing elements) by high-grade metamorphism and tectonic intercalation of supracrustal assemblages.

APPENDIX: PETROGRAPHIC DESCRIPTIONS OF ANALYZED SAMPLES

83-157 Metapelite. Fine- to medium-grained polygonal plagioclase, garnet, quartz, and orthopyroxene, with minor biotite, rutile, apatite, and zircon. Rare clinopyroxene. Layering is defined by concentrations of garnet-orthopyroxene±biotite layers from plagioclase-quartz-rich layers. Acicular rutile occurs in quartz, plagioclase, garnet, and orthopyroxene. Garnets have kelyphitic rims. CO₂-rich fluid inclusions locally abundant in plagioclase and garnet. Brown glass common on grain boundaries.

85-101 Metasediment. Fine- to medium-grained polygonal to seriate plagioclase, quartz, and garnet with minor orthopyroxene, rutile, and sillimanite and rare zircon. Sillimanite occurs solely as oval grains within garnet. Orthopyroxene occurs as thin rims on plagioclase, which in turn rims garnet. Note differences in plagioclase composition between coronas (~An₂₇) and matrix (An₄₂) (Table 3). CO₂-rich fluid inclusions locally abundant in garnet. Garnet is pervasively cracked, with dark brown staining along fractures.

83-160 Metagranodiorite. Fine-grained, polygonal plagioclase, quartz, orthopyroxene, garnet, perthitic K-feldspar, and clinopyroxene with minor biotite, ilmenite, magnetite, apatite, rutile, and zircon. Garnets have thick, dark, kelyphitic rims. Brown glass occurs along grain boundaries. CO₂-rich fluid inclusions present throughout.

83-162 Meta-adamellite. Fine- to medium-grained, polygonal to seriate quartz, perthitic K-feldspar, and garnet with accessory apatite, rutile, magnetite, ilmenite, and zircon. Thin (2 mm), garnet-rich layers occur throughout. Garnets have dark alteration rims; brown glass occurs along grain boundaries. Abundant CO₂-rich fluid inclusions occur in garnet.

85-100 Mafic Granulite (Melt). Medium-grained, polygonal orthopyroxene, plagioclase, and clinopyroxene with accessory rutile, biotite, ilmenite, apatite, and zircon. An orthopyroxene-rich band (with ~70% orthopyroxene) forms 3/4 of sample. Biotite occurs as small grains interstitial to orthopyroxene. Orthopyroxenes are stained brown and contain acicular rutile inclusions. Clinopyroxene stained yellow. CO₂-rich fluid inclusions locally abundant in clinopyroxene.

85-108 Mafic Granulite (Melt). Fine-grained, polygonal plagioclase, clinopyroxene, and quartz with xenoblastic garnet. Accessory phases include orthopyroxene, rutile, apatite, and biotite. Garnet has kelyphitic rims. Clinopyroxene contains fine-grained exsolution lamellae of orthopyroxene.

85-120 Mafic Granulite (Melt). Finely layered (on mm scale), fine-grained polygonal plagioclase, orthopyroxene, and clinopyroxene with minor amphibole (which rims clinopyroxene), rutile, ilmenite, and rare zircon. Clinopyroxene contains orthopyroxene exsolution lamellae. Light brown melt on grain boundaries. CO₂-rich fluid inclusions occur as trails within plagioclase and pyroxenes.

83-158 *Mafic Granulite (Cumulate)*. Coarse-grained clinopyroxene, garnet, and plagioclase with minor amphibole (near rims of clinopyroxenes), ilmenite, and orthopyroxene and rare zircon. Clinopyroxene contains abundant exsolution lamellae of orthopyroxene, plagioclase, and garnet. Xenoblastic groundmass garnet contains acicular rutile inclusions and has kelyphitic rims. Plagioclases are optically zoned. Trails of CO₂-rich fluid inclusions occur throughout.

85-106 *Mafic Granulite (Cumulate)*. Coarse-grained clinopyroxene, amphibole, garnet, scapolite, ilmenite, and plagioclase. Garnets are xenoblastic and contain acicular rutile inclusions. Garnets rim ilmenite and are often observed between scapolite and clinopyroxene. Amphibole is interstitial to scapolite and clinopyroxene. Scapolite grain boundaries rimmed by fine-grained, tabular, anorthite-rich plagioclase.

83-159 *Mafic Granulite (Restite)*. Medium- to coarse-grained polygonal garnet, plagioclase, and clinopyroxene with accessory amphibole, rutile, apatite, and zircon. Garnets are nearly completely replaced by kelyphite. Clinopyroxenes contain orthopyroxene exsolution lamellae and abundant CO₂-rich fluid inclusions.

85-107 *Mafic Granulite (Cumulate)*. Fine-grained, polygonal plagioclase, garnet, quartz, and clinopyroxene with accessory ilmenite, apatite, and zircon. Garnets have thick kelyphitic rims which contain radially arranged, tabular crystals of plagioclase, pyroxenes, and spinel which are up to 0.25 mm long. Fluid inclusions are rare.

85-114 *Mafic Granulite (Restite)*. Medium-grained, polygonal garnet, clinopyroxene, and plagioclase with accessory, tabular rutile (up to 3 mm long). Garnets contain acicular rutile inclusions and are nearly completely replaced by fine-grained kelyphite. Dark melts with quenched crystals occur along grain boundaries. Plagioclase is strongly optically zoned. No fluid inclusions observed.

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