

Rb-Sr and Sm-Nd isotopic variations in dissected crustal xenoliths

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Abstract—Four crustal xenoliths from Scotland have been dissected in order to evaluate the effects of magma-xenolith interaction on Rb-Sr and Sm-Nd isotopic systematics, and hence the use of the xenoliths for accurate determination of source compositions. Each xenolith was sampled from the edge to the center. The Nd isotopic compositions are virtually uniform across each xenolith; however, there are significant variations in Rb, Sr, and REE concentration, as well as Rb/Sr and Sm/Nd ratios and Sr isotopic composition. Most of this variation appears to be inherited from the protolith. However, one sample shows systematic variations in Rb and Sr concentration and Sr isotopic composition with distance from the xenolith edges due to interaction with the host magma. This sample displays petrographic evidence for grain boundary melt infiltration. These data are consistent with theoretical considerations of transport times and diffusivities that place limits on the amount of modification that is possible via diffusion alone. The calculated effects of transport of even large amounts of crustal xenoliths on the compositions of xenoliths and host magmas will be small, provided transport times are short.

Overall, the isotopic variations within the xenolith suite mirror the source rocks from which they were extracted and are inferred to reflect variations in the deep crust of Scotland. However, small samples, such as most crustal xenoliths, may be of little use in defining meaningful crustal residence ages because of modal variations in metamorphic minerals and isotopic re-equilibration. The one xenolith studied has extremely variable Sm/Nd ratios due to the presence of garnet and yields a broad range of geologically meaningless Nd model ages, but yields an Sm-Nd isochron age of 396 ± 49 Ma. The Loch Roag xenolith has a large range of Rb/Sr ratio, and the Sr isotopic data define an isochron age of 416 ± 28 Ma. Both ages are taken to represent the timing of metamorphism and/or cooling from 600°C at the end of the Caledonian Orogeny in Scotland.

INTRODUCTION

A KNOWLEDGE OF THE COMPOSITION of the lower continental crust is critical to our understanding of many large-scale geologic processes and problems, including crustal underplating, high-grade metamorphism, the chemical composition of the bulk continental crust, and the nature of the Moho.

Geophysical measurements, such as of bulk densities of minerals and rocks (CHRISTENSEN and FOUNTAIN, 1975; HALL and AL-HADDAD, 1979; HALL and SIMMONS, 1979; JACKSON and ARCULUS, 1984), can be combined with V_p data obtained from seismic refraction studies to place constraints on the structure of the lower crust. However, a detailed picture can be obtained only from the analysis of samples excavated from lower crustal depths. Exposed high-grade gneiss terrains are dominantly Archean or Proterozoic in age, and have intermediate chemical compositions (WEAVER and TARNEY, 1980a,b). In contrast, the lower crust as sampled by granulite-facies xenoliths appears to be dominated by mafic compositions (GRIFFIN and O'REILLY, 1986, 1987). While this difference might be due partly to a change in the composition of the crust with time, thermobarometry calculations suggest that many granulite-facies, high-grade gneiss terrains were formed at a shallower depth than mafic granulite-facies xenoliths (BOHLEN and MEZGER, 1989). Therefore, mafic granulite-facies xenoliths have the greatest potential for representing genuine lower crustal samples; and, arguably, the biggest advances in our knowledge of the lower

crust in recent years can be linked with geochemical and petrological studies of such samples.

However, xenoliths are not always pristine. Transport processes may cause chemical and physical modification of xenoliths (RUDNICK, 1993). SELVERSTONE (1982) has shown that xenoliths with a diameter of about 10 cm can be heated up to the temperature of the host magma within 3 h of entrainment. Decompression of xenoliths may result in the development of microcracks and partial melting along grain boundaries (GARVEY and ROBINSON, 1984; JONES et al., 1983; PADOVANI and CARTER, 1977). Xenoliths can also show evidence of alteration by fluids during transport (EHRENBERG and GRIFFIN, 1979; EMERY et al., 1985; GRIFFIN et al., 1979; PADOVANI et al., 1982; ROGERS, 1977; ROGERS and HAWKESWORTH, 1982; RUDNICK and TAYLOR, 1987). Such effects are easily observable in thin section. Despite the obvious potential for magma-xenolith interaction, many recent isotopic studies of crustal xenoliths (DOWNES and LEYRELOUP, 1986; HARTE et al., 1981; HAWKESWORTH et al., 1983; LEE et al., 1985; MCCULLOCH et al., 1982; RUIZ et al., 1988) have not evaluated this as a possible cause of artifacts that may obscure the original trace element and isotopic compositions.

In this study, four xenolith samples from different localities from NW to SE across Scotland have been dissected to study the variations in Rb, Sr, Sm, and Nd concentrations, and Nd and Sr isotopic composition, within individual samples. This study attempts to address the questions of whether there is

partial exchange between the xenoliths and the host magma, how it affects the use of the xenoliths as representative lower crustal samples, and the nature of the mass transfer processes involved. Scotland is an ideal location for such an experiment because the deep crust has been well studied using both geophysical and geochemical techniques (BAMFORD et al., 1978; DAVIES et al., 1984; HALL et al., 1983; HALLIDAY, 1984; HALLIDAY and STEPHENS, 1984; HALLIDAY et al., 1979, 1984, 1985, 1993; HUNTER et al., 1984; JACOB et al., 1985; PIDGEON and AFTALION, 1978; STEPHENS and HALLIDAY, 1984; UPTON et al., 1976, 1983, 1984). It is generally argued that there is a granulite-facies lower crust, the age of which increases to Archean toward the NW across Scotland. It was shown in a recent study of crustal xenoliths found across the various tectonic zones of Scotland (Fig. 1) that the isotopic compositions of Nd, Sr, and Pb are highly variable, both generally and within individual localities (HALLIDAY et al., 1985, 1993).

PETROGRAPHY

Petrographic descriptions of various examples from the Scottish xenolith suite were presented by GRAHAM and UPTON (1978), UPTON et al. (1976, 1983, 1984), HUNTER et al. (1984), and HALLIDAY et al. (1984, 1993). The four crustal xenoliths analyzed in this study are from three different areas in Scotland: LR114 is from Loch Roag, in the Outer Hebrides; STP23 is from Streap Comlaidh, in the Highlands; and FD155 and CYH33 are from Fidra and Coalyard Hill in the Midland Valley (Fig. 1). The host rocks for LR114 and STP23 are monchiquite dikes, and the host rocks for FD155 and CYH33 are basanites. The host igneous rocks are part of the late Carboniferous (ca. 300 Ma) igneous activity of northern Britain with the exception of the host to LR114, the Loch Roag dike, which was emplaced at ca. 50 Ma (R. M. Macintyre, pers. commun. 1985).

Meta-gabbroic and mafic granulite xenoliths are ubiquitous throughout the various Scottish xenolith localities. Garnet is absent from nearly all of these assemblages, limiting direct geobarometric

estimates. However, most are thought to be representative of the lower crust (currently at depths of ca. 20–30 km in this region) because of the coexistence of ultramafic assemblages, in particular spinel lherzolites (UPTON et al., 1983; MENZIES et al., 1987; MENZIES and HALLIDAY, 1988), and the correlation with seismic velocities (BAMFORD et al., 1978).

The xenoliths tend to be small (rarely > 10 cm diameter), and the feldspathic samples sometimes show evidence of partial melting and incipient disaggregation. The mafic granulites are mostly granular textured; some, such as the Streap sample studied here, exhibit modal layering and mineral lamination. The overall range is from pyroxenite to anorthosite, and relic igneous textures are common. The whole-rock chemistry for the suites at Fidra and Loch Roag, together with a large unpublished data base of mineral chemistry, suggest a common igneous parentage for each locality, suggesting that they probably represent disrupted layered mafic complexes within the deep crust. Three samples have been recovered from Fidra (E. Midland Valley) in which evidence of former garnet is present in the form of sp-pl-ol symplectites, textures typical of decompressional reactions associated with uplift. The rarity of garnet indicates that most xenoliths probably equilibrated at depths of less than 30 km and that there was not much crust of mafic composition below this depth during the Permo-Carboniferous. Petrographic evidence of interaction and partial melting is usually confined to the outer 1–5 mm of xenoliths.

At Loch Roag (Isle of Lewis, Outer Hebrides), the modal mineralogy of the mafic xenoliths is as follows: plagioclase (pl) 50–85%; clinopyroxene (cpx) 10–35%; orthopyroxene (opx) 0–15%; titanomagnetite (Ti-mt) 0–10%; potassium feldspar (ksp) 0–10%; apatite (ap) 0–1%; zircon (zc) is occasionally observed. The plagioclase compositional range is $An_{42.3}Or_{3.1}$ to $An_{26.7}Or_{5.6}$. The xenoliths are 4–15 cm in diameter with a grain size of 1–5 mm. Most samples exhibit a granular texture with weak modal layering. Plagioclase and spinel show local reaction with minor zonation in plagioclase within 100–200 μ m of adjacent spinel. Sample LR114 is a mafic two-pyroxene granulite. The modal mineralogy is approximately 50% plagioclase and alkali feldspar, 30% clinopyroxene, 10% orthopyroxene, less than 10% magnetite, and a minor amount of serpentinous pseudomorphs after orthopyroxene. The xenolith has interacted slightly with the host. A thin rim can be seen on one side of the xenolith, where pyroxene is more Ti-rich and plagioclase shows reaction to analcime. There is some analcime, Ti-augite, kaersutite, and Ti-magnetite (i.e., host mineralogy) along the grain boundaries, which indicates that melt infiltration has taken place. However, the interaction is slight. The only other sign of alteration is the occasional very thin (< 1 mm) late fracture, in-filled with alteration products cutting across the fabric of the xenolith and also penetrating the host. Pyroxenes have thin exsolution lamellae of spinel on cleavages, which is a common characteristic for the crustal xenoliths at Loch Roag.

The samples from Streap Comlaidh in the central Highlands have modes of 30–60% pl; 25–40% cpx; 10–20% opx; < 10% Ti-mt and < 1% ap. There is little compositional zonation in individual phases. Sample STP23 contains about 40% clinopyroxene, 10% orthopyroxene, 45% plagioclase, less than 5% titanomagnetite, and minor amounts of apatite and sulfide. The xenolith displays a granular texture with modal layering on a cm scale. The grain size is less than 1 mm. Clinopyroxene shows fine lamellae of spinel along the cleavages. Plagioclase ranges in composition from $An_{41}Or_4$ to $An_{27}Or_6$. The xenolith has embayed margins, and melt infiltration in the form of fine-grained basalt is seen along some grain margins. Clinopyroxene is enriched in Ti where it is adjacent to the host. The last equilibration temperature is about 850°C, based on two-pyroxene thermometry.

The xenoliths from Fidra in East Lothian, Midland Valley, which have been described in detail by HUNTER et al. (1984), have the following modal mineralogy: 40–85% pl; 0–50% cpx; < 5% opx; 0–15% Ti-mt; 0–5% olivine (ol); and < 2% ap. The plagioclase exhibits a compositional range from $An_{51}Or_1$ to $An_{31}Or_{10}$. The xenoliths are 2–15 cm in diameter with a grain size of 1–8 mm but usually < 3 mm. Modal layering is ubiquitous on a cm-scale. Granular textures are present in most, but relic cumulus texture is evident in some. There is evidence of former garnet (now a pl + ol + sp symplectite) in some samples. Many samples show evidence of clinopyroxene and orthopyroxene reacting with plagioclase and titanomagnetite to form granular olivine. There are also local spinel-plagioclase reactions with

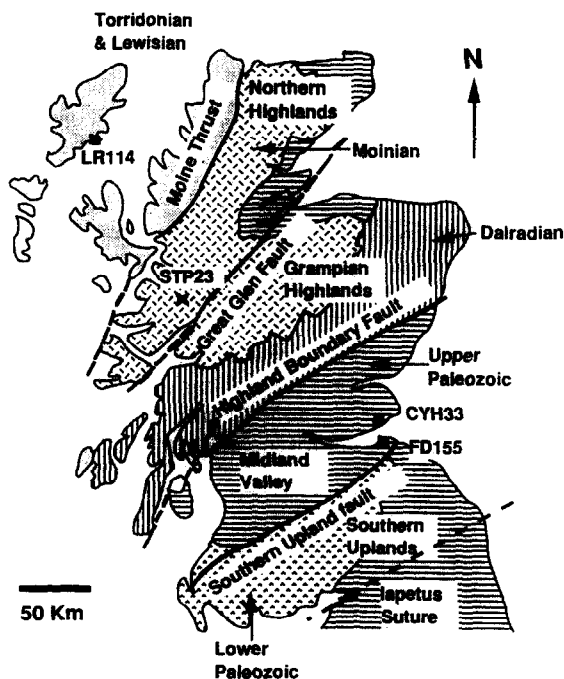


FIG. 1. Geological sketch map of Scotland and northern England showing the major stratigraphic divisions and xenolith locations.

plagioclase zonation adjacent to spinel. Decompression melting is evident in many with subsequent quench crystallization of anorthoclase/titanomagnetite intergrowths. Interaction with the host is generally limited to the outer 0.5 mm of xenolith; in these areas, clinopyroxene tends to be richer in Ti. Sample FD155 is a layered meta-anorthosite composed of about 85% highly strained plagioclase in crystals up to 5 mm, trace amounts of clinopyroxene, and some secondary alteration products (mostly prehnite), including anhedral magnetite.

The felsic granulites are most abundant in the Midland Valley, with similar lithologies present through the central Irish province (STROGEN, 1974). The felsic granulites are granular textured but, in general, much more deformed than the mafic granulites with ribbon texture, grain-size reduction, and strain exhibited by quartz and fractured plagioclase. Mineral lamination is enhanced by the deformation with strong fabrics present in some samples; grain size is varied, again as a function of deformation. Undeformed samples have grain sizes in the range 1–8 mm. Modal layering is abundant and, since it is largely responsible for mineralogical variation within a given suite, modal compositions are somewhat meaningless. The suite has meta-sedimentary protoliths in most cases, perhaps modified through intracrustal melting at some stage. However, some of the trondhjemitic assemblages may represent intermediate igneous rocks. Typical assemblages are: $qz + pl \pm ksp \pm ga$, and $qz + pl + ksp + ga \pm sillimanite$ (sill) $\pm rutile$ (rut) $\pm graphite$ (graph) $\pm zoisite$ (zo) $\pm ap$ (apatite). The abundance of graphite in many samples is indicative of the supracrustal origin of many of these aluminous xenoliths. Garnet, when present, is alm-py with <5 mol% gr (in contrast to mafic granulites, where gr is typically 10–30 mol% in the few garnetiferous examples). The composition of plagioclase is typically in the range of $An_{20-40}Or_{2-10}$. In general, felsic xenoliths tend to be more altered and particularly prone to carbonation. Much of this is due to relatively low-temperature circulation within the loosely consolidated vents which form the hosts for many of the Midland Valley samples. Commonly, individual phases remain fresh (although plagioclase sometimes is incipiently altered to clay minerals), while the carbonate forms as thin veins and along grain boundaries.

Sample CYH33 is a garnet-bearing gneiss and is the only garnetiferous felsic granulite retrieved from Coalyard Hill. It displays coarse modal layering, with concentrations of garnet and granular plagioclase alternating with ribbons of quartz and feldspar and granular textured quartz and feldspar. Felsic ribbons are up to 3 mm in size, and garnets are usually 1 mm. Quartz comprises about 10%, varying from large porphyroclastic (approximately 6 mm in diameter) to smaller, dominantly equigranular (<1 mm) grains with equilibrated neoblasts, together with plagioclase (andesine) and perthite. The garnets are idiomorphic and rounded with scalloped edges. They possess altered kelyphitic rims with chlorite corona. There is fine exsolution of rutile in the garnet. Plagioclase is fine-grained and equigranular in garnet domains but is coarser and fractured in quartz domains. The deformation in quartz and plagioclase has partially annealed. Titanomagnetite, rutile, apatite, and zoisite are also present; the latter is quite abundant as rounded grains. Locally, calcite is abundant as veins and pseudomorphs. Some pale greenish chloritic pseudomorphs re-

place clinopyroxene with trace amounts of biotite and opaque oxide. However, this rock is typical of felsic granulites found throughout the Midland Valley of Scotland and probably represents the meta-sedimentary protolith into which the host mafic magmas were emplaced. The alteration in this sample may reflect retrograde metamorphism, and this probably should have affected the Rb-Sr systematics in particular.

WHOLE ROCK CHEMICAL COMPOSITIONS

Major and trace element analyses and Nd and Sr isotope geochemistry on representative segments from the original whole rocks from near the center of each xenolith were obtained by XRF at the University of Edinburgh, ICP at the University of London, and the Scottish Universities Research and Reactor Center as part of a larger study of the entire Scottish xenolith suite, to be presented in full elsewhere (HALLIDAY et al., 1993). The major and trace element data, as well as Nd and Sr isotopic data for these samples are presented in Tables 1 and 2, along with major element analyses for some of the smaller segments studied here. The small sample size used in this study limited the opportunities for acquiring comprehensive compositional data for each segment.

Whereas mafic xenoliths from individual localities exhibit considerable variation in major element composition, there are certain distinctive trace element characteristics. The Loch Roag xenoliths display depletion in U, with variable enrichments in Rb. The xenoliths from Fidra are highly variable in composition but commonly display positive Eu anomalies, consistent with their cumulate textures, and are relatively enriched in Ba, K, and Sr. Unlike the other suites, they are not strongly Nb depleted and have more of an alkali basaltic trace element pattern. The Sr concentrations at Loch Roag and Fidra are a function of alumina content consistent with feldspar accumulation. The prevalence of feldspar in these suites is an important factor in the control of large ion lithophile element distributions.

The general features of the trace element compositions of the four samples studied here are illustrated in Fig. 2. The low abundances of Th in all the samples and Y and HREEs in FD155 are at the limits of measurement by XRF and ICP and have been deleted from Fig. 2 because the data produce excess scatter. All of the xenoliths display some degree of LREE enrichment but have relatively flat HREE patterns. The mafic xenoliths from Loch Roag and Strep Comlaidh tend to be particularly U depleted. The pronounced Sr, Eu, and Ti anomalies in sample FD155 from Fidra are as expected from a cumulate meta-anorthosite dominated by plagioclase and titanomagnetite. The Loch Roag sample displays the Rb and K enrichment characteristic of xenoliths from this site. This enrichment at Loch Roag is consistent with the presence of alkali feldspar in these samples. However the K/U and Ba/Rb ratios of LR114 are extreme; they are 131,000 and 3.9, respectively. Enrichment may reflect fractionation from alkali-rich parent magmas, as are common in this region, a metasomatic alteration of the protolith rocks, or interaction with alkali-rich host magmas during transport to the surface. However, the K/U and Ba/Rb ratios are difficult to explain simply by fractionation of alkali basalt. While the combined evidence suggests that

Table 1: Major element concentrations for representative xenolith samples (Halliday et al., 1993) and selected serial samples (this study).

	LR114*	LR114A	LR114B	LR114C	LR114D	STP23*	STP23F	FD155*	FD155A	FD155C	FD155D	CYH33*	CYH33E	CYH33F
SiO ₂	48.90	43.33	47.17	47.95	48.29	47.10	47.27	43.70	42.58	46.79	48.31	50.70	52.32	49.01
TiO ₂	1.17	2.68	1.28	1.20	1.22	1.29	1.13	3.28	3.04	2.63	2.24	1.22	1.02	1.30
Al ₂ O ₃	14.10	13.67	13.75	13.40	14.50	13.80	13.87	21.50	13.57	22.74	23.29	18.40	16.67	18.32
Fe ₂ O ₃	13.40	9.51	15.06	14.42	13.26	13.50	14.93	14.00	14.03	11.81	10.27	7.49	5.93	8.04
MnO	0.20	0.19	0.19	0.21	0.17	0.22	0.24	0.07	0.18	0.06	0.05	0.18	0.15	0.16
MgO	6.95	8.84	6.81	6.96	6.79	7.77	8.54	1.18	9.59	0.68	0.59	4.17	3.56	4.60
CaO	11.40	9.92	10.92	11.44	11.22	10.00	9.42	5.87	9.42	5.85	6.03	9.57	9.04	7.70
Na ₂ O	2.75	1.94	2.36	2.25	2.43	2.83	2.36	5.35	3.40	5.51	5.66	2.73	2.61	2.46
K ₂ O	1.01	3.47	1.02	0.86	0.69	0.40	0.33	1.17	0.64	1.25	1.19	1.88	1.95	2.07
P ₂ O ₅	0.09	1.35	0.10	0.10	0.11	0.07	0.03	0.11	0.62	0.10	0.11	0.09	0.19	0.10
LOI	2.81	3.77	1.23	0.92	1.30	3.14	1.30	3.28	2.30	1.64	1.35	--	5.79	4.94
Total	99.97	98.67	99.89	99.71	99.98	99.79	99.42	99.37	99.37	99.06	99.09	99.71	99.23	98.70

All obtained by fused disc XRF method. Those marked with * are whole rock analyses of representative aliquots generally sampled from near the center of each xenolith.

Table 2: Trace element concentrations for xenoliths (ppm).

	LR114	STP23	FD155	CYH33
Ni	61	61	39	74
Cr	80	282	17	210
V	317	396	308	159
Sc	47	52	239	2
Cu	70	43	177	47
Zn	96	82	61	28
Sr*	201.6	216.5	2204	541.1
Rb*	79.9	5.08	15.8	21.6
Zr	70	78	45	210
Nb	5	5	11	8
Y	27	29	6	39
Ba	311	187	524	465
U*	0.064	0.081	0.226	0.228
Pb*	2.30	1.46	1.63	6.80
⁸⁷ Sr/ ⁸⁶ Sr*	0.71256±7	0.71195±4	0.70313±6	0.71125±4
La	16.9	7.75	8.49	16.7
Ce	31.2	20.1	16.0	32.0
Nd*	21.23	15.37	7.15	15.12
Sm*	4.58	3.98	1.22	4.17
Eu	1.15	1.29	1.12	1.35
Gd	3.54	3.7	--	4.98
Dy	3.95	4.36	--	6.07
Er	2.24	2.72	--	4.17
Yb	2.05	2.53	--	4.54
¹⁴³ Nd/ ¹⁴⁴ Nd*	0.51149±3	0.51196±4	0.51262±3	0.51202±4

* Taken from Halliday et al. (1993). Concentrations measured by isotope dilution. Isotopic compositions normalized for interlaboratory differences. Analytical uncertainties include those of normalization.

All the trace element concentrations, except for REEs, were measured by pressed powder XRF method, while REEs, except for Sm and Nd, were measured by ICP.

mafic xenolith FD155 was part of a fractionated igneous suite, the exact origin of STP23 is less clear and LR114, while originally igneous in origin, may have suffered alkali enrichment at some stage.

The felsic xenoliths are undoubtedly varied in origin. The silica contents of the felsic samples generally lie in the range of 60–90 wt% SiO₂. Many of these are metasedimentary gneisses. CYH33 is poor in silica and high in alumina, Sr, and Pb, with high K/U and Nb/U ratios, and is probably restitic from crustal melting, the felsic lamellae representing melt segregations. However, CYH33 displays a slight HREE enrichment, consistent with the presence of garnet.

TECHNIQUES

Figure 3 shows outline sketches of the four xenoliths used in this study, together with the manner in which they were sampled. The xenoliths were cut in the middle, and an approximately 8-mm thick slice was taken. The slice was then divided into 3–7 subslices; and, in some cases, the immediate host rock was sampled as well (LR114, STP23 and FD155).

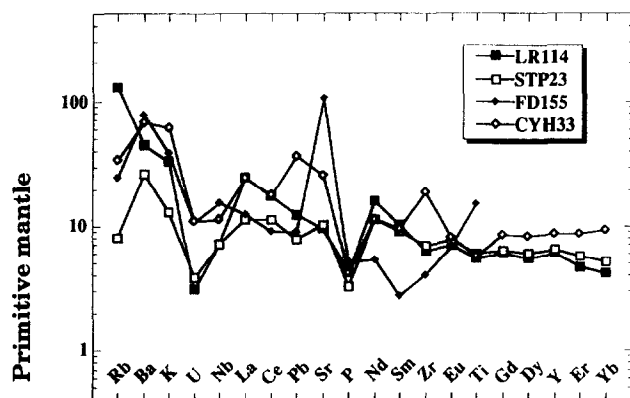


FIG. 2. Primitive mantle-normalized trace element patterns for bulk xenoliths. Normalizing values were taken from SUN and McDONOUGH (1989).

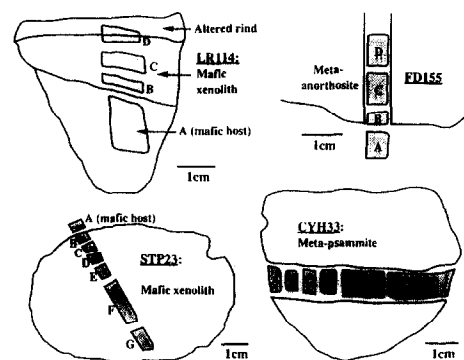


FIG. 3. Schematic sketch of each xenolith with sampling distribution.

Major element analyses were acquired for segments of sufficient size using a Philips PW1480 Automatic X-ray Spectrometer after fusion with Li borate (Spectroflux 105). The analyses were calibrated with a range of USGS and CRPG standards. All isotopic compositions and isotope dilution measurements were made in the Radiogenic Isotope Geochemistry Laboratory of the University of Michigan. About 30 mg of powder from each slice was dissolved in HF, HNO₃, and HCl. Isotope dilution measurements were made on aliquots of about 20% of the total solution utilizing mixed ⁸⁷Rb/⁸⁴Sr and ¹⁴⁹Sm/¹⁵⁰Nd spikes. The separation of Rb, Sr, and REEs was achieved using 3 mL of Bio-Rad 50W X8 200–400 mesh resin in a quartz column. Separation of Nd and Sm from the other REEs was achieved using a 3-mL column of PTFE powder coated with 2-Diethylhexyl orthophosphoric acid. Routine total process blanks were approximately 50 pg Rb, 100 pg Sr, and <70 pg Nd.

Analyses were performed on two VG Sector thermal ionization mass spectrometers with multiple Faraday collectors. Both machines give a value of $0.71025 \pm 2 (2\sigma)$ for the ⁸⁷Sr/⁸⁶Sr ratio of the NBS987 Sr standard and $0.51185 \pm 2 (2\sigma)$ for the ¹⁴³Nd/¹⁴⁴Nd ratio of the La Jolla Nd standard. Strontium was loaded onto a zone-refined Re single filament with a slurry of Ta₂O₅ in 5% H₃PO₄. Rubidium was loaded on the side filament of a triple Ta filament in 4M HNO₃. Both Sm and Nd were loaded in 10% HNO₃ on the side filament of a triple Ta-Re-Ta filament assembly.

SPATIAL VARIATIONS IN COMPOSITION

Major Element Chemistry

The major element compositions of some of the segments studied here are given in Table 1. From these, and the data for the representative whole rocks in Table 1, it can be seen that the different segments of each xenolith are broadly similar. The only notable variation in LR114 is in potassium. The two FD155 segments are lower in Fe, Mg, and Ti than the whole rock. The two segments and whole rock of CYH33 display significant differences in Fe, Mg, Ca, Ti, and P, consistent with the coarse grain size, modal layering, and patchy alteration. It is safe to assume that LR114 and STP23 are reasonably homogeneous at the scale sampled, with the exception of potassium in the former. In contrast FD155 contains a small degree of variability, and CYH33 is heterogeneous.

Concentrations of Rb, Sr, Sm, and Nd

The results of isotope dilution Rb, Sr, Sm, and Nd measurements are presented in Table 3. The concentrations of Rb, Sr, and Nd are plotted as a function of segment distance from the center of each xenolith in Fig. 4, except for FD155,

Table 3: Isotopic ratio and isotopic dilution data for crustal xenoliths and their hosts.

Sample	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}^*$	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}^*$	$\epsilon(\text{Nd})_t$	T_{DM} Ga
LR114 A†	89.4	1730	0.1488	0.705887 ± 13	12.1	86.2	0.08513	0.512355 ± 11	-4.7	0.9
LR114 B	72.5	225	0.9276	0.711340 ± 13	4.57	20.4	0.1358	0.511556 ± 10	-20.6	3.1
LR114 C	69.5	193	1.038	0.711827 ± 20	4.99	21.9	0.1376	0.511578 ± 11	-20.3	3.1
LR114 D	40.7	190	0.6161	0.709404 ± 21	4.77	24.4	0.1180	0.511474 ± 7	-22.3	2.7
STP23 A†	61.5	1150	0.1538	0.706946 ± 30	9.53	52.3	0.1101	0.512541 ± 17	+1.4	0.9
STP23 B	5.39	237	0.0655	0.712174 ± 29	2.44	9.74	0.1518	0.511947 ± 16	-11.8	2.9
STP23 C	4.88	209	0.0674	0.712203 ± 31	1.75	6.35	0.1664	0.511961 ± 11	-12.1	3.8
STP23 D	4.25	188	0.0650	0.712403 ± 38	3.25	11.7	0.1683	0.511966 ± 8	-13.5	4.2
STP23 E	3.71	151	0.0707	0.712522 ± 23	4.99	18.0	0.1675	0.511977 ± 20	-11.8	3.8
STP23 F	3.93	175	0.0648	0.712395 ± 23	3.74	13.2	0.1710	0.511961 ± 14	-12.2	4.2
STP23 G	5.80	242	0.0687	0.711607 ± 13	5.63	22.8	0.1490	0.511838 ± 10	-13.8	3.1
FD155 A†	7.30	1077	0.01952	0.703772 ± 15	7.98	39.1	0.1234	0.512718 ± 11	+4.4	0.7
FD155 B	22.7	2190	0.02982	0.703447 ± 20	1.35	7.70	0.1058	0.512655 ± 10	+3.9	0.7
FD155 C	16.9	2180	0.02227	0.703292 ± 15	0.83	5.15	0.09679	0.512651 ± 11	+4.1	0.7
FD155 D	13.8	2392	0.01657	0.703142 ± 14	0.82	5.20	0.09509	0.512641 ± 17	+3.9	0.7
CYH33 A	20.7	598	0.09956	0.710654 ± 13	5.78	28.7	0.1217	0.511899 ± 12	-11.5	2.1
CYH33 B	28.0	506	0.1593	0.710639 ± 17	6.87	30.1	0.1378	0.511891 ± 33	-12.3	2.5
CYH33 C	29.0	537	0.1552	0.710912 ± 11	4.27	20.4	0.1266	0.511902 ± 11	-11.7	2.2
CYH33 D	28.4	553	0.1480	0.711283 ± 11	5.22	23.8	0.1325	0.511917 ± 11	-11.6	2.3
CYH33 E	24.6	519	0.1365	0.711316 ± 24	4.43	18.6	0.1438	0.511938 ± 18	-11.6	2.6
CYH33 F	24.3	478	0.1463	0.711496 ± 28	4.31	13.7	0.1908	0.512072 ± 13	-10.9	7.1

† Host composition.

* The uncertainties shown above are 2 sigma errors. All Sr isotopic analyses were normalized to $^{86}\text{Sr}/^{88}\text{Sr}=0.1194$, and Nd isotopic analyses were normalized to $^{146}\text{Nd}/^{144}\text{Nd}=0.7219$. $\lambda^{87}\text{Rb}=1.42\times 10^{-11}\text{y}^{-1}$, $\lambda^{147}\text{Sm}=6.54\times 10^{-12}\text{y}^{-1}$, $^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}}=0.512638$, $^{147}\text{Sm}/^{144}\text{Nd}_{\text{CHUR}}=0.1966$.

which is plotted with distance from the contact between the host rock and the xenolith. The concentration of Sm is not plotted because it varies in a similar manner to Nd concentration.

The concentrations of Rb, Sr, and Nd within LR114 are very similar for segment B (close to the edge) and C (at the center), whereas D is lower in Rb (Fig. 4a). The segments (including the representative whole rock) with higher Rb also have lower K/Rb and higher Rb/Sr. The host, in contrast, has higher K/Rb and lower Rb/Sr. Therefore, if the Rb enrichment were caused by interaction with the host, it would need to be selective relative to K and Sr.

The concentrations of Rb and Sr in STP23 increase from the center to the margins of the xenolith (Fig. 4b). This strongly suggests the partial incorporation of these elements from the host magma, which has higher concentrations of both Rb and Sr. Since the variations are nearly symmetrical, the mass transfer must have taken place after the entrainment of the xenolith, as opposed to in a wall rock environment. Although it appears that the outer parts of the xenolith have suffered enrichments in Rb and Sr, no such effect is apparent in the Nd concentrations. Samples B, C, and D have lower concentrations of Nd even though they are adjacent to the host magma, which is enriched in Nd. This difference is partly because of the higher concentrations of Rb and Sr in the host relative to the xenolith. For example, for sample F, closest to the center, $\text{Rb}_{\text{Host}}/\text{Rb}_{\text{F}} = 15.6$, $\text{Sr}_{\text{Host}}/\text{Sr}_{\text{F}} = 6.6$, and $\text{Nd}_{\text{Host}}/\text{Nd}_{\text{F}} = 4.0$. However, the degree of variability in Nd is actually greater than that for Rb or Sr, presumably reflecting protolith heterogeneities. As such, the effects of small additions of Nd are undetectable.

FD155 (Fig. 4c) has been sampled in three areas. Sample B at the margin of the xenolith has the highest Rb concentration and is higher than that of the host magma. Sample B also has a higher concentration of Nd than C and D, which

are almost identical. The concentration of Sr is highest in D. These data provide no clear evidence that exchange with the host magma has affected the Rb, Sr, or Nd concentrations; but the sampling renders the results somewhat inconclusive.

CYH33 (Fig. 4d) does not display systematic variations in Rb, Sr, or Nd. Segments A and F have the lowest Rb contents, while segment C has the highest. The large variations in Rb and Sr in different portions might be ascribable to modal heterogeneity or low-temperature alteration. Segments A and B have higher Nd contents, which almost certainly reflects modal differences.

Nd and Sr Isotopic Compositions

The isotopic compositions of Nd and Sr for each segment are presented in Table 3. Only one xenolith, STP23, displays a systematic variation in ($^{87}\text{Sr}/^{86}\text{Sr}$)_t ratio (where *t* = time of eruption) with distance from the center of the xenolith that can unequivocally be ascribed to interaction with the host (Fig. 5). STP23 shows a gradual increase in ($^{87}\text{Sr}/^{86}\text{Sr}$)_t toward the center of the xenolith, coupled with a decrease in Sr concentration. The variations in Sr isotopic ratios are not coupled with the variations in Rb/Sr ratios. For example, there is a decrease in ($^{87}\text{Sr}/^{86}\text{Sr}$)_t at the edge of the xenolith (segments B & G); but the Rb/Sr ratios of these two samples are not significantly different from other samples, providing evidence that Sr from the host magma has been incorporated at the margins of the xenolith, and that these variations are not inherited from the xenolith protolith. This is consistent with the petrographic evidence of melt infiltration at the margins of this xenolith. FD155 shows an increase of ($^{87}\text{Sr}/^{86}\text{Sr}$)_t from the mafic host toward the center of the xenolith. This correlates with Rb/Sr ratio and is therefore possibly inherited from the protolith. The Rb-Sr systematics may also have been affected by secondary alteration. The plot of

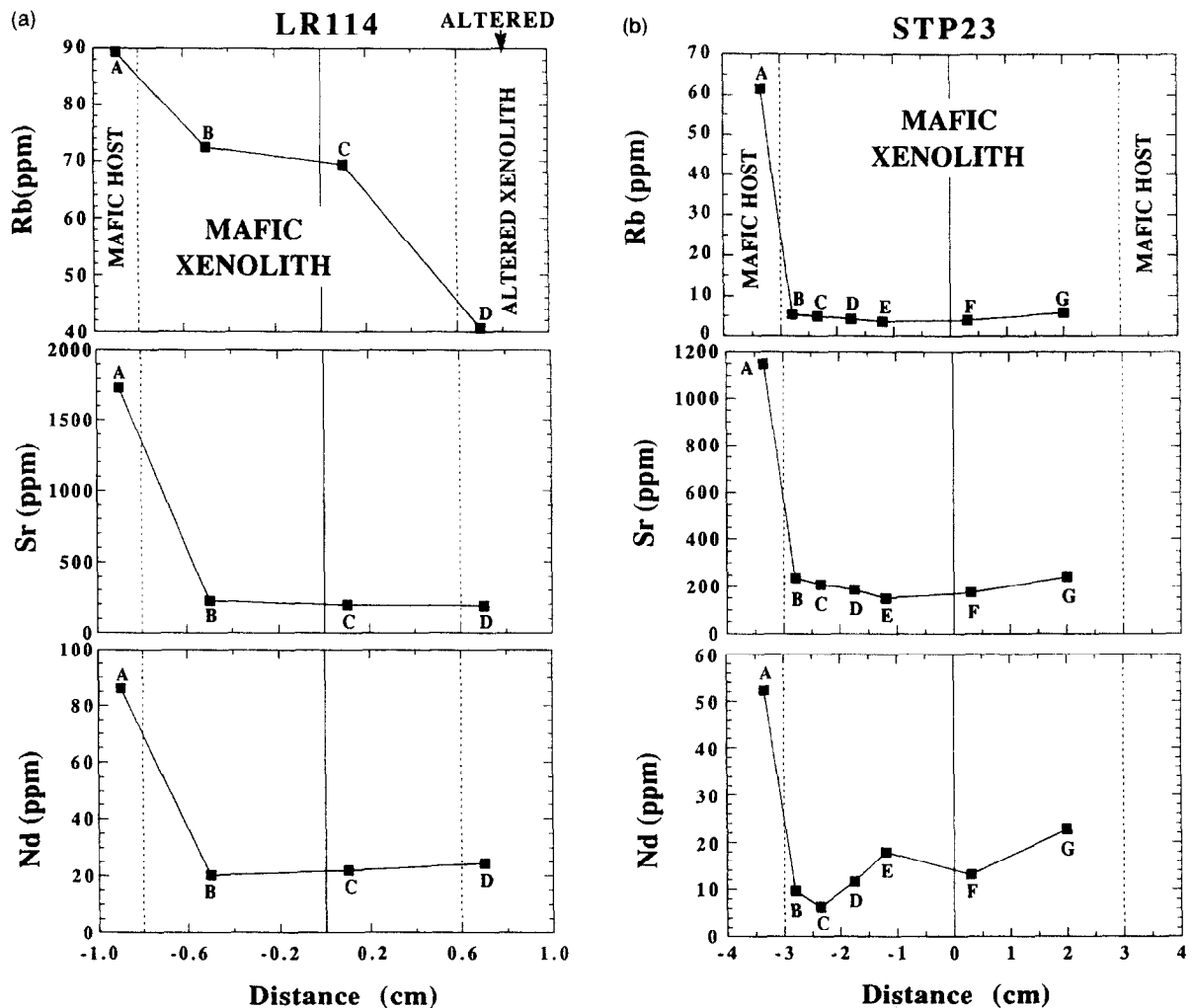


FIG. 4. Rb, Sr, and Nd concentrations plotted vs. the distance relative to the center of each xenolith, except for FD155, which is plotted from the edge. The vertical lines in the center represent the center of the xenolith, and the broken lines represent the two edges.

CYH33 displays an increase of ($^{87}\text{Sr}/^{86}\text{Sr}$), from B to F. This appears to be inversely correlated with the variations in Rb/Sr ratio but positively correlated with variations in Nd isotopic composition. This possibly reflects growth of garnet with a high Sm/Nd ratio from alumina-rich portions that were previously rich in biotite with a high Rb/Sr ratio. However, again, no great reliance can be placed on this because of the secondary alteration of garnet and feldspar.

The most striking feature of the data set is that the $\epsilon(\text{Nd})$ values of each xenolith show very little internal variation (Table 3; Fig. 5). The relatively uniform Nd isotopic compositions are an unambiguous indication that there has been little mass transfer of the REEs between the magmas and the xenoliths. In each case, the xenolith and host have distinct Nd isotopic compositions. The same can be anticipated for the felsic xenolith CYH33, with very unradiogenic Nd compared with Midland Valley Carboniferous basaltic igneous rocks. Hence the internal uniformity was not brought about through equilibration with the host magma.

The xenoliths are also internally fairly uniform in Sm-Nd model age (Table 3), with the exception of the margins of

STP23 and the felsic xenolith CYH33, which displays large variations. In the former case, the model age variations do not reflect differing crustal residence times but rather xenolith-magma interaction. In the latter case, the variations reflect the growth of garnet, which fractionates Sm/Nd ratios.

DIFFUSION VERSUS MELT INFILTRATION

There is now a considerable body of data available on diffusion in silicate minerals and silicate melts (HOFMANN, 1980; HOFMANN and MAGARITZ, 1977; JAMBON and CARRON, 1976; MAGARITZ and HOFMANN, 1978a,b; SHAW, 1974; SMITH, 1974; WASSERBURG et al., 1964; WATSON et al., 1985; YODER, 1973). Among three different kinds of diffusion models—chemical, tracer, and self-diffusion—chemical diffusion is the most suitable way of modelling the chemical variations observed in this study.

Diffusion coefficients of Rb, Sr, and REEs in relevant silicate liquid and crystal phases have been measured by KASPER (1974), HOFMANN and MAGARITZ (1977), MISRA and VENKATUSBRAMANIAN (1977), MAGARITZ and HOFMANN (1978a,b), SNEERINGER et al. (1984), WATSON et al. (1985),

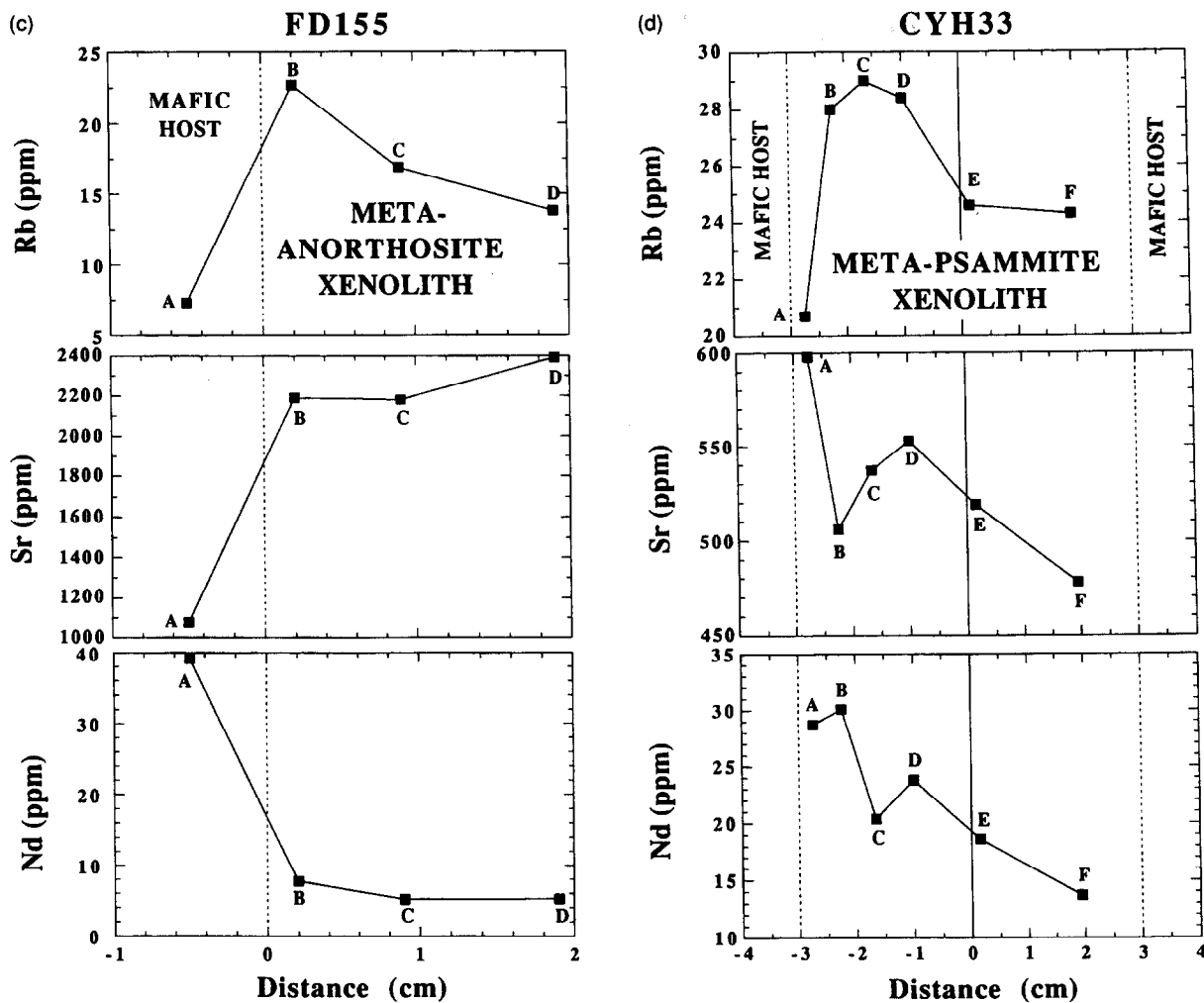


FIG. 4. (Continued)

GILETTI (1991), and WATSON and CHERNIAK (1991). In general, for basaltic compositions at a temperature of 1250°C, the diffusion coefficients of these elements are expected to be on the order of 10^{-6} or 10^{-7} $\text{cm}^2 \text{s}^{-1}$. Rare earth elements and Sr have similar values of 1.4×10^{-7} and 1.6×10^{-7} $\text{cm}^2 \text{s}^{-1}$, respectively (HOFMANN and MAGARITZ, 1977; MAGARITZ and HOFMANN, 1978a,b). Diffusion coefficients for Rb are not available for this temperature range. However, by analogy with lower temperature data, 3×10^{-7} $\text{cm}^2 \text{s}^{-1}$ is probably a reasonable value for Rb. Diffusion coefficients for these elements in silicate minerals are very small, i.e., less than 10^{-13} $\text{cm}^2 \text{s}^{-1}$ in diopside and less than 10^{-10} $\text{cm}^2 \text{s}^{-1}$ in plagioclase at a temperature of 1250°C (KASPER, 1974; SNEERINGER et al., 1984). Clinopyroxene and plagioclase are the most common phases in mafic xenoliths and also of this study. Since crustal xenoliths could possibly be mixtures of both crystals and small amounts of intergranular melt, the best way to model the data is to use two extreme liquid and solid models.

The diffusion transport distance can be modelled with the relation $X = (Dt)^{1/2}$ (X is the mean distance in one direction; D is diffusion coefficient in $\text{cm}^2 \text{s}^{-1}$; t is time in seconds; CRANK, 1975). The time for Rb to diffuse approximately 1

cm in a basaltic melt, at a temperature of 1250°C, is more than 900 h. It would take more than two times longer for Sr to diffuse through the same distance. In contrast, it would take about 300,000 y for these elements to diffuse through the same distance in diopside, at a temperature of 1250°C. The time for these elements to diffuse the same distance in plagioclase is about 300 y. Studies by KUSHIRO et al. (1976) and SPERA (1980), showed that dense xenoliths can be retained in suspension in a basaltic or kimberlitic magma for no more than 50 h. The inclusion assemblage in the Strep Comlaidh dike includes spinel lherzolites and pyroxene megacrysts, so ascent times must have been rapid. In such a short time, even in a melt phase, these elements can diffuse no more than a few mms. In the solid phases, the effects of diffusion would be negligible. It is well known that volatiles can affect the structure of silicate melts and thus affect the diffusion coefficient. SHAW (1974) found that increasing water content may increase diffusivities. However, YODER (1973) noted that the presence of water does not increase the diffusivities drastically. According to SMITH (1974), even in the presence of volatiles, diffusion coefficients will not change by more than an order of magnitude. To account for the isotopic and chemical variations in STP23 by diffusion, the magma

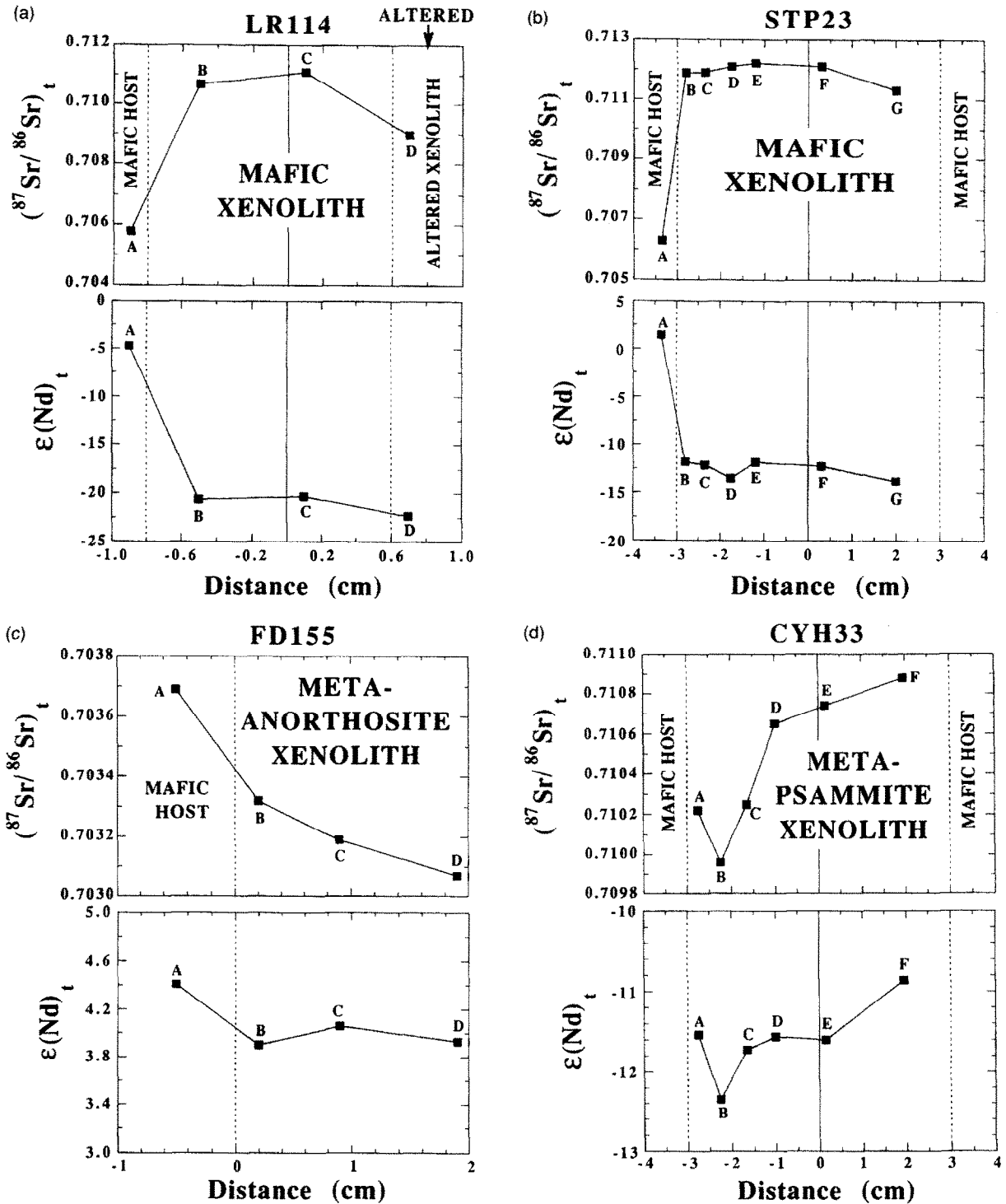


FIG. 5. Plots of $(^{87}\text{Sr}/^{86}\text{Sr})_t$ and $\epsilon(\text{Nd})_t$ vs. distances relative to the center of each xenoliths, except for FD155, which was plotted from the edge.

would need to exchange with the xenolith in a magma chamber, or in conduits, for some time prior to being brought to the surface. This model requires a constantly erupting magma for long periods, or a magma that stayed in a crustal reservoir for a long time, where most of the exchange took place.

A more realistic mechanism for producing the profile observed in STP23 is melt infiltration. WATSON (1982) showed

from surface-energy considerations that a magma body in chemical and thermal equilibrium with its surroundings will tend to infiltrate through dry grain boundaries of the wall rock at a rate of 1–2 mm/day. This rate is roughly the same as the diffusion rate in basaltic melt. Similar results were predicted by TURCOTTE and AHERN (1978) on the basis of buoyancy. Of the xenoliths from this study, STP23 shows

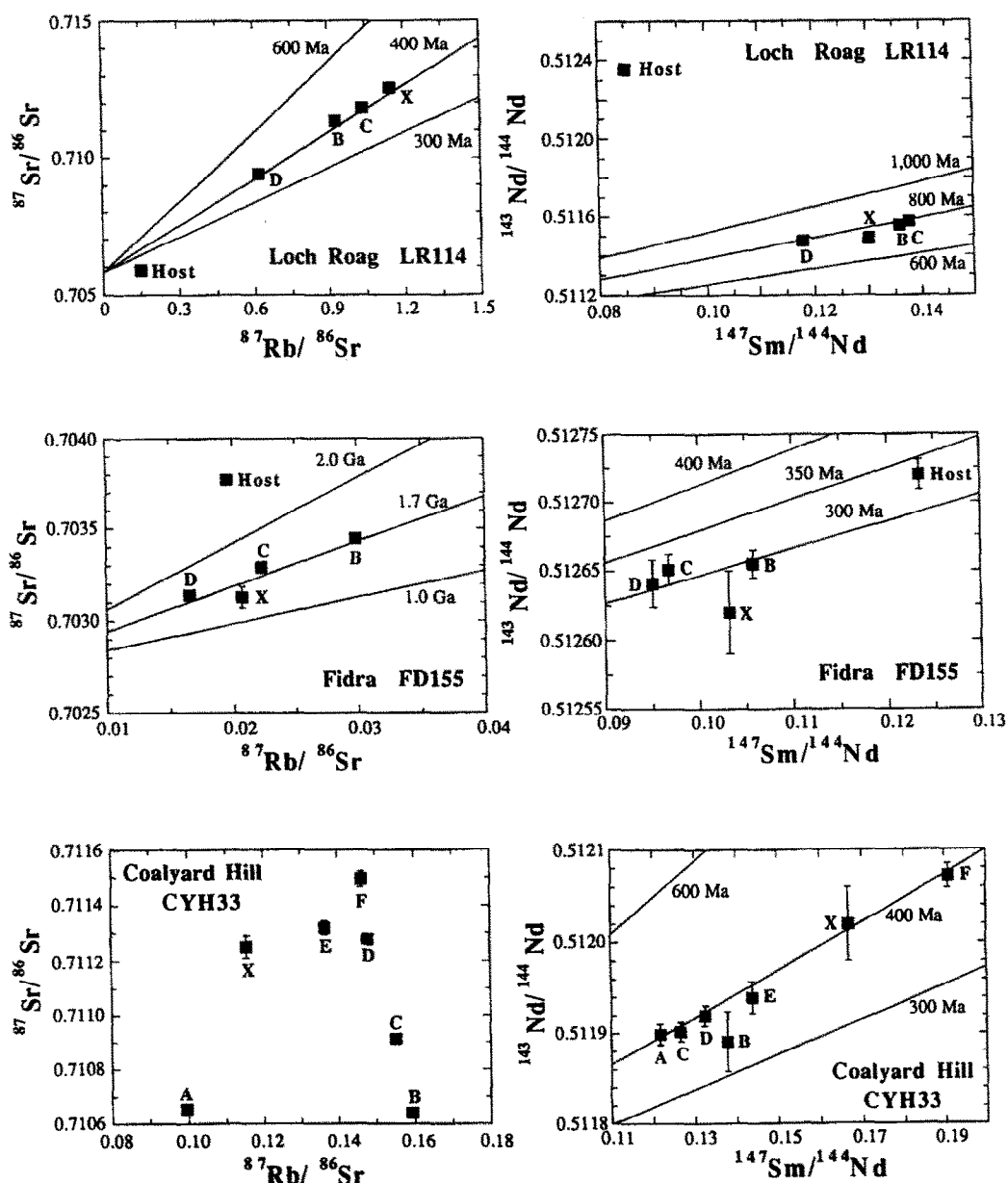


FIG. 6. Plots of $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$, respectively, for all the xenoliths (except STP23). The error bars shown are 2σ errors. No error bar is shown for samples for which the size of their symbols is greater than the 2σ errors. Samples marked with X are for representative whole-rock analyses (HALLIDAY et al., 1993).

the clearest petrographic evidence of melt infiltration, and this mechanism appears to have been responsible for the small amount of chemical and isotopic variation.

In summary, the data from this study show that negligible amounts of interaction between host liquid and entrained crustal xenoliths have taken place. Unambiguous evidence of modification of a xenolith is found only where there is petrographic evidence of melt infiltration. In a similar manner, exchange with xenoliths cannot significantly alter the concentration and isotopic composition of Sr in the host magmas. This can be evaluated for STP23 as the most extreme example, assuming STP23E represents a pristine composition and the chemical gradients observed in the xenolith are the same from segment E in all directions. Even if the magma entrained 20% of such xenoliths, neither the Sr con-

centration nor the Sr isotopic composition of the host magma is likely to be significantly affected.

EQUILIBRATION AGES

If the only xenolith that was affected by melt infiltration is STP23, the isotopic compositions and parent/daughter ratios of the other serially sectioned samples might be expected to provide evidence of the timing of last equilibration over the scale sampled. The $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{147}\text{Sm}/^{144}\text{Nd}$ ratios for every segment of each sample, except STP23, which has suffered melt infiltration, are plotted against their respective $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in Fig. 6. None of the data define perfect isochrons.

The spread in Rb/Sr ratios for LR114 produces a (scattered) apparent isochron age of 416 ± 28 Ma (with an MSWD

of 7.8). The composition of the host does not lie along the linear array defined by the compositions of the xenoliths. The Rb enrichment and high Rb/Sr ratios found in these xenoliths appear to have been produced at approximately 420 Ma. The Rb-Sr age is unlikely to be a result of metasomatism by the host magma. Since the host has a lower Rb/Sr ratio than the xenolith, there would have to have been preferential addition of Rb relative to Sr. Such an effect is not seen in the two samples B and C of LR114, from the edge and center of the xenolith, respectively. In fact, the edge has a lower Rb/Sr ratio than the center. The 420 Ma isochron almost certainly reflects the time of last Sr isotopic equilibration in the deep crust. Given the closure temperature for Sr in alkali feldspar (WATSON and CHERNIAK, 1991), these data probably define the time the system last cooled through approximately 600°C (depending on the cooling rate). The foreland region in NW Scotland was characterized by ultrapotassic magmatism at this time (HALLIDAY et al., 1987). It is therefore possible that this xenolith represents metamorphosed igneous rocks from a deep crustal magma chamber formed at ca. 420 Ma. Alternatively, mantle-derived potassium-rich melts or fluids may have been instrumental in pervasively modifying the deep crust in this region at this time. The latter is supported by the low Ba/Rb ratio of this sample. The spread in Sm/Nd is small, resulting in a less precise age estimate of 758 ± 260 Ma (MSWD = 6.5). However, the Sm-Nd data do indicate that minor, but significant, Nd isotopic heterogeneity was preserved on a small scale. This is consistent with a model of modification of an older protolith at about 420 Ma.

In the case of FD155, the age defined by the Sr isotopic data is 1.63 ± 0.39 Ga (MSWD = 8.7). However, the Nd isotopic composition was uniform at the time of magmatism (100 ± 350 Ma; MSWD = 2.7). The apparent "age" defined by the Rb-Sr data is considerably in excess of the Sm-Nd model age (ca. 0.7 Ga). This is difficult to explain unless the Rb-Sr age reflects inherited heterogeneity or some disturbance to the systematics. The range in Rb/Sr ratio is small; therefore, the Rb-Sr "isochron age" is probably dominated by the effects of later disturbances, rather than by in-situ radiogenic additions.

The Rb-Sr isotopic data for CYH33 scatter drastically, consistent with the alteration of this sample. However, the Sm-Nd system is relatively well behaved. CYH33 is the only xenolith to display a reasonable spread in Sm/Nd ratio and Nd isotopic composition and yields an apparent age of 396 ± 49 Ma (MSWD = 2.0). This age (as with the 416 ± 28 Ma Rb-Sr age of LR114) coincides with the timing of the major granite plutonism in Scotland. COHEN et al. (1988) have suggested that the closure temperature for Sm-Nd in garnet is $>900^\circ\text{C}$. If this is true, the most likely explanation for the CYH33 data is that the metamorphism is young (Phanerozoic). If the closure temperature is $<700^\circ\text{C}$ (MEZGER et al., 1993), the ages may instead reflect the time of last cooling from high temperatures for the crust represented by the xenoliths.

In the case of both the Loch Roag and the Coalyard Hill samples, it appears that useful geochronological information on the protoliths is preserved in the isotopic heterogeneity of the xenoliths. While these results must be regarded as pre-

liminary, the ages obtained are taken to indicate that the sections of lower crust under the Outer Hebrides and under the Midland Valley were disturbed in late Silurian and early Devonian times, a conclusion that is hardly surprising given that this represents the time of formation of the majority of granites in Scotland (PIDGEON and AFTALION, 1978, HALLIDAY et al., 1979).

CONCLUSIONS

The four crustal xenoliths analyzed in this study show spatial variations in the concentrations of Rb, Sr, and Nd, and Nd and Sr isotopic composition. These variations are inherited in part from the protolith but, in one case, have been modified by melt infiltration from the host magma. Serial sections of a garnetiferous felsic xenolith and an anorthositic have disturbed Rb-Sr systematics, but these may be related to retrograde metamorphism and low-temperature alteration. The diffusion coefficients for Rb and Sr are sufficiently small in silicates so that it would have taken much longer than the transport times to produce the effects observed in the xenoliths by volume diffusion. Melt infiltration appears to be the only mechanism capable of significantly changing the composition of the xenolith during transit over typical transport times. The results of this study lend confidence to the interpretation of the isotopic and chemical compositions of xenoliths transported in basaltic magmas as reflecting their source regions. This study highlights the potential problems of interpreting Sm-Nd model ages from metamorphic rocks, particularly in garnetiferous gneisses with strongly fractionated Sm/Nd ratios in which the sample slice is not representative of the bulk composition (RUDNICK, 1990).

The xenoliths with large variations in parent/daughter ratio (Rb/Sr in the Loch Roag sample; Sm/Nd in the Coalyard Hill sample) appear to preserve geologically meaningful isochron ages. The "closure temperature" of a cm-scale sectioned whole rock system would be a useful concept if the assemblage were relatively simple in terms of the host minerals of the major, diffusing species. This is the case in the examples discussed here. That is, at Loch Roag, the phases hosting Rb and Sr are simply alkali feldspar and plagioclase; and the apparent age would be a function of the closure temperature for the phase with higher closure temperature. (For a recent discussion of this, see GILETTI, 1991.) In the case of the Coalyard Hill sample, the phase controlling the range in Sm/Nd is garnet. Until accurate closure temperatures for Sm-Nd in garnet and feldspar are determined, it will remain unclear which phase limits the closure temperature for the whole-rock system. The ca. 400-Ma ages obtained in this study probably date either the last major metamorphism and establishment of isotopic and textural equilibrium or cooling from relatively high temperatures ($>600^\circ\text{C}$, depending on the exact closure temperature for Sr in feldspar and Nd in feldspar and garnet). Finally, the effects of entrainment of even relatively large volumes of xenoliths on the Nd and Sr isotopic compositions of the basaltic host magmas is generally small.

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REFERENCES

- BAMFORD D., NUNN K., PRODEHL C., and JACOB B. (1978) LISPB—IV. Crustal structure of northern Britain. *Geophys. J. Roy. Astr. Soc.* **54**, 43–60.
- BOHLEN S. R. and MEZGER K. (1989) Origin of granulite terrains and the formation of the lowermost continental crust. *Science* **244**, 326–329.
- CHRISTENSEN N. I. and FOUNTAIN D. M. (1975) Constitution of the lower continental crust: Based on experimental studies of seismic velocities in granulite. *GSA Bull.* **86**, 227–236.
- COHEN A. S., O'NIONS R. K., SIEGENTHALER R., and GRIFFIN W. L. (1988) Chronology of the pressure-temperature history recorded by a granulite terrain. *Contrib. Mineral. Petrol.* **98**, 303–311.
- CRANK J. (1975) *The Mathematics of Diffusion*, 2nd ed. Oxford Univ. Press.
- DAVIES G. R., UPTON B. G. J., and STROGEN P. (1984) Sr and Nd isotope evidence for age and origin of crustal xenoliths from the Midland Valley of Scotland and central Ireland (abstr.). *Trans. Roy. Soc. Edinb.: Earth Sci.* **75**, 297.
- DOWNES H. and LEYRELOUP A. (1986) Granulitic xenoliths from the French Massif Central—Petrology, Sr and Nd isotope systematics, and model age estimates. In *The Nature of the Lower Continental Crust* (ed. J. B. DAWSON, et al.); *GSA Spec. Publ.* **24**, pp. 319–330.
- EHRENBERG S. N. and GRIFFIN W. L. (1979) Garnet granulite and associated xenoliths in minette and serpentinite diatremes of the Colorado Plateau. *Geology* **7**, 483–487.
- EMERY J. V., ARCULUS R. J., and HENDERSON C. E. (1985) A possible mineral host for the unusual Sr- and Ba-enrichments in some lower crustal granulites and eclogites. *Eos* **66**, 387.
- GARVEY O. C. and ROBINSON D. N. (1984) The formation of kelyphite and associated subkelyphite sculptured surfaces on pyrope from kimberlite. In *Kimberlites and Related Rocks* (ed. J. KORNPROBST), pp. 371–382. Elsevier.
- GILETTI B. J. (1991) Rb and Sr diffusion in alkali feldspars, with implications for cooling histories of rocks. *Geochim. Cosmochim. Acta* **55**, 1331–1343.
- GRAHAM A. M. and UPTON B. G. J. (1978) Gneisses in diatremes, Scottish Midland Valley: Petrology and tectonic implications. *J. Geol. Soc. London* **135**, 219–228.
- GRIFFIN W. L. and O'REILLY S. Y. (1986) The lower crust in eastern Australia: Xenolith evidence. In *The Nature of the Lower Continental Crust* (ed. J. B. DAWSON et al.); *GSA Spec. Publ.* **24**, pp. 363–374.
- GRIFFIN W. L. and O'REILLY S. Y. (1987) The composition of the lower crust and the nature of the continental Moho—Xenolith evidence. In *Mantle Xenoliths* (ed. P. H. NIXON). John Wiley & Sons Ltd.
- GRIFFIN W. L., CARSWELL D. A., and NIXON P. H. (1979) Lower-crustal granulites and eclogites from Lesotho, Southern Africa. In *The Mantle Sample: Inclusions in Kimberlites* (ed. F. R. BOYD and H. O. A. MEYER). AGU, Washington, DC.
- HALL J. and AL-HADDAD F. M. (1979) Variation of effective seismic velocities of minerals with pressure and its use in velocity prediction. *Geophys. J. Roy. Astr. Soc.* **57**, 107–118.
- HALL J. and SIMMONS G. (1979) Seismic velocities of Lewisian metamorphic rocks at pressures to 8 Kbar: Relationship to crustal layering in North Britain. *Geophys. J. Roy. Astr. Soc.* **58**, 337–347.
- HALL J., POWELL D. W., WARNER M. R., EL-ISA Z. H. M., ADE-SANYA O., and BLUCK B. J. (1983) Seismological evidence for shallow crystalline basement in the Southern Uplands of Scotland. *Nature* **305**, 418–420.
- HALLIDAY A. N. (1984) Coupled Sm-Nd and U-Pb systematics in late Caledonian granites and the basement under northern Britain. *Nature* **307**, 229–233.
- HALLIDAY A. N. and STEPHENS W. E. (1984) Crustal controls on the genesis of the 400 Ma old Caledonian granites. *Phys. Earth. Planet. Interiors* **35**, 89–104.
- HALLIDAY A. N., AFTALION M., VAN BREEMEN O., and JOCELYN J. (1979) Petrogenetic significance of Rb-Sr and U-Pb isotopic systems in the ca 400 Ma old British Isles granitoids and their hosts. In *The Caledonides of the British Isles—Reviewed* (ed. A. L. HARRIS, C. H. HOLLAND, and B. E. LEAKE); *GSA Spec. Publ.* **8**, pp. 653–662.
- HALLIDAY A. N., AFTALION M., UPTON B. G. J., ASPEN P., and JOCELYN J. (1984) U-Pb isotopic ages from a granulite facies xenolith from Partan Craig in the Midland Valley of Scotland. *Trans. Roy. Soc. Edinb.: Earth Sci.* **75**, 71–74.
- HALLIDAY A. N., STEPHENS W. E., HUNTER R. H., MENZIES M. A., DICKIN A. P., and HAMILTON P. J. (1985) Isotopic and chemical constraints on the building of the deep Scottish lithosphere. *Scottish J. Geol.* **21**, 465–491.
- HALLIDAY A. N., AFTALION M., PARSONS I., DICKIN A. P., and JOHNSON M. R. W. (1987) Syn-orogenic alkaline magmatism and its relationship to the Moine Thrust Zone and the thermal state of the Lithosphere in NW Scotland. *J. Geol. Soc. London* **144**, 611–617.
- HALLIDAY A. N., DICKIN A. P., DAVIES G. R., DEMPSTER T. J., HAMILTON P. J., HUNTER R. H., and UPTON B. G. J. (1993) Formation and composition of the lower continental crust: Evidence from Scottish xenolith suites. *J. Geophys. Res.* (in press).
- HARTE B., JACKSON P. M., and MACINTYRE R. M. (1981) Age of mineral equilibria in granulite facies nodules from kimberlites. *Nature* **291**, 147–148.
- HAWKESWORTH C. J., ERLANK A. J., MARSH J. S., MENZIES M. A., and VAN CALSTEREN P. (1983) Evolution of the continental lithosphere: Evidence from volcanics and xenoliths in southern Africa. In *Continental Basalts and Mantle Xenoliths* (ed. C. J. HAWKESWORTH and M. J. NORRY), pp. 111–138. Shiva.
- HOFMANN A. W. (1980) Diffusion in natural silicate melts: A critical review. In *Physics of Magmatic Processes* (ed. R. B. HARGRAVES), pp. 385–418, Princeton Univ. Press.
- HOFMANN A. W. and MAGARITZ M. (1977) Diffusion of Ca, Sr, Ba, and Co in a basalt melt: Implications for the geochemistry of the mantle. *J. Geophys. Res.* **82**, 5432–5440.
- HUNTER R. H., UPTON B. G. J., and ASPEN P. (1984) Meta-igneous granulite and ultramafic xenoliths from basalts of the Midland Valley of Scotland: Petrology and mineralogy of the lower crust and the upper mantle. *Trans. Roy. Soc. Edinb.: Earth Sci.* **75**, 75–84.
- JACKSON I. and ARCULUS R. J. (1984) Laboratory wave velocity measurements on lower crustal xenoliths from Calcutteroo, South Australia. *Tectonophysics* **101**, 185–197.
- JACOB A. W. B., KAMINSKI W., MURPHY T., PHILIPS W. E. A., and PRODEHL C. (1985) A crustal model for a northeast-southwest profile through Ireland. *Tectonophysics* **113**, 75–103.
- JAMBON A. and CARRON J.-P. (1976) Diffusion of Na, K, Rb, and Cs in glasses of albite and orthoclase composition. *Geochim. Cosmochim. Acta* **40**, 897–903.
- JONES A. P., SMITH J. V., DAWSON J. B., and HANSEN E. C. (1983) Metamorphism, partial melting, and K-metasomatism of garnet-scapolite-kyanite granulite xenoliths from Lashine, Tanzania. *J. Geology* **91**, 143–166.
- KASPER R. B. (1974) Cation diffusion in a low albite. *GSA Abstr. Prog.* **6**, 815.
- KUSHIRO I., YODER H. S., JR., and MYSÉN B. O. (1976) Viscosities of basalt and andesite melts at high pressures. *J. Geophys. Res.* **81**, 6351–6356.
- LEEMAN W. P., MENZIES M. A., MATTY D. J., and EMBEE G. F. (1985) Strontium, neodymium, and lead isotopic compositions of deep crustal xenoliths from the Snake River Plain: Evidence for Archean basement. *Earth Planet. Sci. Lett.* **75**, 354–368.

- MAGARITZ M. and HOFMANN A. W. (1978a) Diffusion of Sr, Ba, and Na in obsidian. *Geochim. Cosmochim. Acta* **42**, 595–605.
- MAGARITZ M. and HOFMANN A. W. (1978b) Diffusion of Eu and Gd in basalt and obsidian. *Geochim. Cosmochim. Acta* **42**, 847–858.
- MCCULLOCH M. T., ARCULUS R. J., CHAPPELL B. W., and FERGUSON J. (1982) Isotopic and geochemical studies of nodules in kimberlite have implications for the lower continental crust. *Nature* **300**, 166–169.
- MENZIES M. and HALLIDAY A. N. (1988) Chemical provinciality in the Scottish sub-crustal lithosphere. In *Oceanic and Continental Lithosphere: Similarities and Differences* (ed. M. A. MENZIES and K. G. COX); *J. Petrol. Spec. Lithosphere Issue*, pp. 275–302.
- MENZIES M., HALLIDAY A. N., PALACZ Z., HUNTER R. H., UPTON B. G. J., ASPEN P., and HAWKESWORTH C. J. (1987) Evidence from mantle xenoliths for an enriched lithospheric keel under the outer Hebrides. *Nature* **325**, 44–47.
- MEZGER K., ESSENE E. J., and HALLIDAY A. N. (1993) Closure temperature of the Sm-Nd system in metamorphic garnets. *Earth Planet. Sci. Lett.* (in press).
- MISRA N. K. and VENKATASBRAMANIAN V. S. (1977) Strontium diffusion in feldspars: A laboratory study. *Geochim. Cosmochim. Acta* **41**, 837–838.
- PADOVANI E. R. and CARTER J. L. (1977) Non-equilibrium partial fusion due to decompression and thermal effects in crustal xenoliths. In *Magma Genesis* (ed. H. J. B. DICK); *Oregon Dept. Geol. Mineral. Indust. Bull.* **96**, 43–57.
- PADOVANI E. R., HALL J., and SIMMONS G. (1982) Constraints on crustal hydration below the Colorado Plateau from Vp measurements on crustal xenoliths. *Tectonophysics* **84**, 313–328.
- PIDGEON R. T. and AFTALION M. (1978) Cogenetic and inherited zircon U-Pb systems in granites: Paleozoic granites of Scotland and England. In *Crustal Evolution in Northwestern Britain and Adjacent Regions* (ed. D. R. BOWES and B. E. LEAKE); *Geol. J. Spec. Issue No. 10*, 183–220.
- ROGERS N. W. (1977) Granulite xenoliths from Lesotho kimberlites and the lower continental crust. *Nature* **270**, 681–684.
- ROGERS N. W. and HAWKESWORTH C. J. (1982) Proterozoic age and cumulate origin for granulite xenoliths, Lesotho. *Nature* **299**, 409–413.
- RUDNICK R. L. (1990) Nd and Sr isotopic compositions of lower-crustal xenoliths from north Queensland, Australia: Implications for Nd model ages and crustal growth processes. *Chem. Geol.* **83**, 195–208.
- RUDNICK R. L. (1993) Xenolith—Samples of the lower continental crust. In *Continental Lower Crust* (ed. D. FOUNTAIN, R. J. ARCULUS, and R. W. KAY) (in press).
- RUDNICK R. L. and TAYLOR S. R. (1987) The composition and petrogenesis of the lower crust: A xenolith study. *J. Geophys. Res.* **92**, 13,891–14,005.
- RUIZ J., PATCHETT P. J., and ARCULUS R. J. (1988) Nd-Sr isotope composition of lower crustal xenoliths—Evidence for the origin of mid-Tertiary felsic volcanics in Mexico. *Contrib. Mineral. Petrol.* **99**, 36–43.
- SELVERSTONE J. (1982) Fluid inclusions as petrogenetic indicators in granulite xenoliths, Pali-Aike volcanic field, Chile. *Contrib. Mineral. Petrol.* **79**, 28–36.
- SHAW H. R. (1974) Diffusion of H₂O in granitic liquids. Part I. Experimental data; Part II. Mass transfer in magma chambers. In *Geochemical Transport and Kinetics* (ed. A. W. HOFMANN et al.); *Carnegie Inst. Wash. Publ.* **634**.
- SMITH H. D. (1974) An experimental study of the diffusion of Na, K, and Rb in magmatic silicate liquids. Ph.D. diss., Univ. Oregon.
- SNEERINGER M., HART S. R., and SHIMIZU N. (1984) Strontium and samarium diffusion in diopside. *Geochim. Cosmochim. Acta* **48**, 1589–1608.
- SPERA F. J. (1980) Aspects of magma transport. In *Physics of Magmatic Processes* (ed. R. B. HARGRAVES), pp. 265–323, Princeton Univ. Press.
- STEPHENS W. E. and HALLIDAY A. N. (1984) Geochemical contrasts between late Caledonian granitoid plutons of northern, central, and southern Scotland. *Trans. Roy. Soc. Edinb.: Earth Sci.* **75**, 259–273.
- STROGEN P. (1974) The sub-Palaeozoic basement in central Ireland. *Nature* **250**, 562–563.
- SUN S.-S. and MCDONOUGH W. F. (1989) Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes. In *Magmatism in the Ocean Basins* (ed. A. D. SAUNDERS and M. J. NORRY); *GSA Spec. Publ.* **42**, pp. 313–345.
- TURCOTTE D. L. and AHERN J. L. (1978) A porous flow for magma migration in the asthenosphere. *J. Geophys. Res.* **83**, 767–772.
- UPTON B. G. J., ASPEN P., GRAHAM A. M., and CHAPMAN N. A. (1976) Pre-Palaeozoic basement of the Scottish Midland Valley. *Nature* **260**, 517–518.
- UPTON B. G. J., ASPEN P., and CHAPMAN N. A. (1983) The upper mantle and deep crust beneath the British Isles: Evidence from inclusions in volcanic rocks. *J. Geol. Soc.* **140**, 105–122.
- UPTON B. G. J., ASPEN P., and HUNTER R. H. (1984) Xenoliths and their implications for the deep geology of the Midland Valley of Scotland and adjacent regions. *Trans. Roy. Soc. Edinb.: Earth Sci.* **75**, 65–70.
- WASSERBURG G. J., ALBEE A. L., and LANPHERE M. A. (1964) Migration of radiogenic strontium during metamorphism. *J. Geophys. Res.* **69**, 4395–4401.
- WATSON E. B. (1982) Melt infiltration and magma evolution. *Geology* **10**, 236–240.
- WATSON E. B. and CHERNIAK D. J. (1991) Diffusion of Strontium in K-feldspar measured by Rutherford backscattering spectroscopy. *Eos* **72**, 309.
- WATSON E. B., HARRISON T. M., and RYERSON F. J. (1985) Diffusion of Sm, Sr, and Pb in fluorapatite. *Geochim. Cosmochim. Acta* **49**, 1813–1823.
- WEAVER B. L. and TARNEY J. (1980a) Continental crust composition and nature of the lower crust: Constraints from mantle Nd-Sr isotope correlation. *Nature* **286**, 342–346.
- WEAVER B. L. and TARNEY J. (1980b) Rare earth geochemistry of Lewisian granulite-facies gneisses, northwest Scotland: Implications for the petrogenesis of the Archean lower continental crust. *Earth Planet. Sci. Lett.* **51**, 279–296.
- YODER H. S., JR. (1973) Contemporaneous basaltic and rhyolitic magmas. *Amer. Mineral.* **58**, 153–171.