THE ROLES OF LHERZOLITE AND GARNET PYROXENITE IN THE CONSTITUTION OF THE UPPER MANTLE

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Some xenoliths from Salt Lake Crater contain lherzolite in contact with garnet pyroxenite. Previous workers have interpreted these xenoliths to be fragments of deep-seated regions where upper mantle peridotite (lherzolite) was intruded by a melt which crystallized to the pyroxenite assemblage. Textures at the lherzolite-pyroxenite contact indicate that *lherzolite* is younger, having formed from pre-existing garnet pyroxenite. Lherzolite orthopyroxene is derived from both pyroxenite pyroxenes. Lherzolite olivine has formed at the expense of pyroxenite orthopyroxene. Lherzolite chrome diopside and spinel originate as remnants of pyroxenite clinopyroxene and spinel, most of which are consumed in the transformation event.

Rare earth data support the proposed view. Application of crystal/liquid distribution coefficients to pyroxenite "pre-garnet" clinopyroxene REE patterns gives hypothetical liquids unlike any Hawaiian basalt. If the pyroxenites crystallized from such liquids, pyroxenite formation and the current Hawaiian vulcanism are probably genetically unrelated. Garnet pyroxenites have REE patterns capable of giving rise to tholeiitic REE patterns on partial melting. Clinopyroxenes from the lherzolite and pyroxenite portions of a single xenolith have very similar REE distributions. This is consistent with the textural observation that the chrome diopside forms from pyroxenite clinopyroxene. It is inconsistent with the view that lherzolite has been intruded by a melt which crystallized at depth to the pyroxenite assemblage.

Sr isotopic data suggest: (1) some garnet pyroxenites have developed and maintained Sr isotopic disequilibrium for a period of time longer than the age of current Hawaiian vulcanism. (2) Xenoliths and host basalt have different Sr^{87}/Sr^{86} ratios, suggesting the absence of a genetic relationship.

The similarity of lherzolite inclusions at Salt Lake Crater to those around the world suggests that garnet pyroxenite, not lherzolite, may be the dominant rock in the upper mantle. This is in accord with geophysical evidence that the density of the upper mantle may be greater than that allowed by the pyrolite model.

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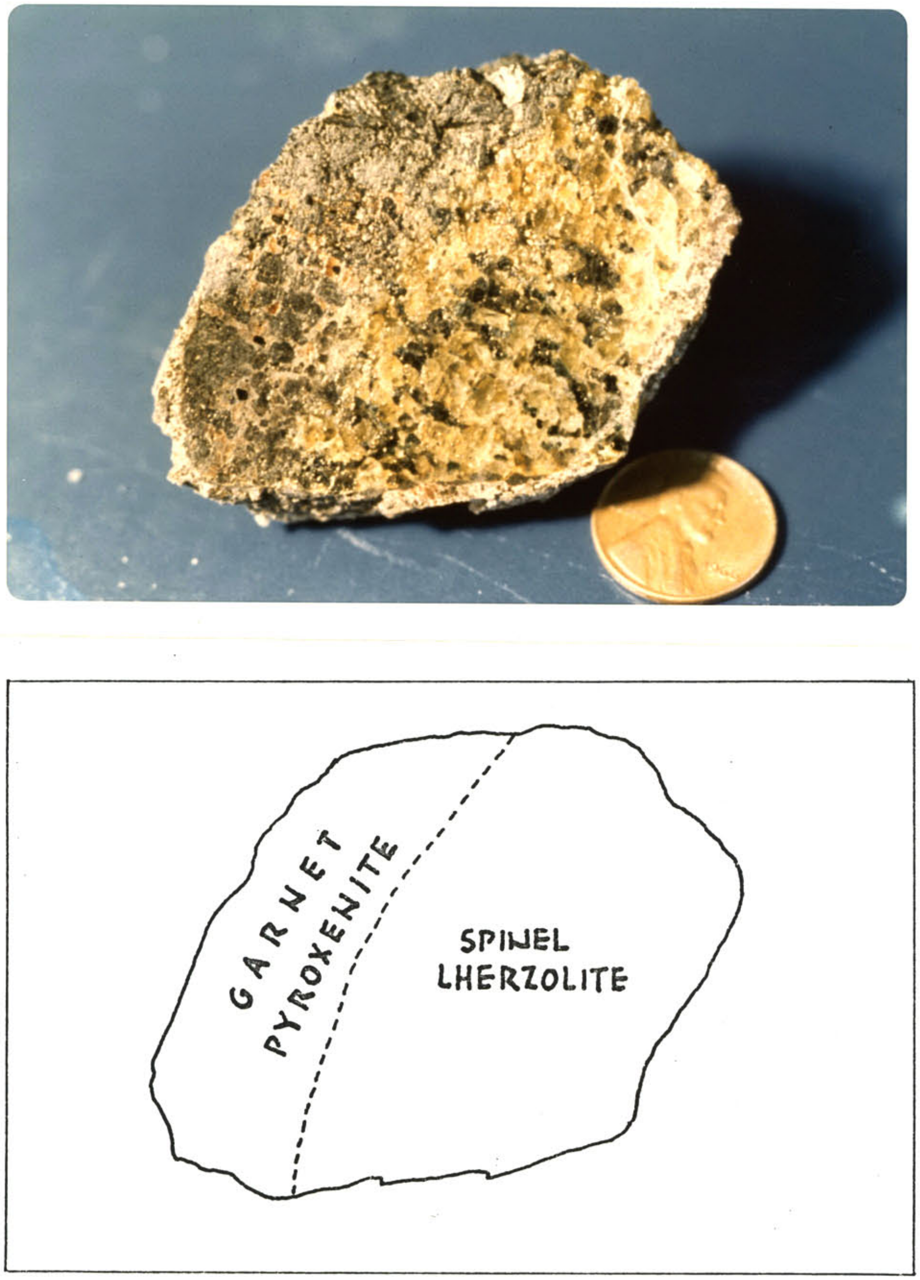

Garnet pyroxenite and spinel lherzolite in contact within a single xenolith from Salt Lake Crater, Oahu.

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ABSTRACT

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CHAPTER I

RESULTS OF PREVIOUS STUDIES OF LHERZOLITE XENOLITHS

Lherzolite xenoliths in basalts have been known for over a century, yet their origin is still a matter of considerable debate. The debate centers on whether the inclusions are genetically related to their host basalt, or whether they were accidentally incorporated in the ascending magma. Several modes of origin for genetically related xenoliths can be envisioned: cumulates, residues from partial fusion, and parental material. Exotic xenoliths can also be of several types including primitive upper mantle material, residue from other melting events, cumulates from unrelated events, and crystallization of entrapped melts. Chapter I summarizes some of the significant studies of these inclusions.

Ross, Foster and Myers (1954) determined major and minor element compositions for minerals from lherzolite xenoliths from widely separated localities around the world (Hawaii, Arizona, Manchuria, California, Alaska, Mexico, Austria, Germany and Africa). All xenoliths analysed consist of the four minerals - olivine, enstatite, chrome diopside and spinel. A plot showing the range in major element composition in olivines, enstatites and chrome diopsides is shown in Figure I-1. The remarkable result is that these silicate minerals show very limited compositional variation despite their geographic dispersion. The range in MgO content, for example, is less than two weight percent in olivines and enstatites, and less than three weight percent in chrome diopsides. The other major elements show similar uniformity. The authors interpret the lherzolites' identical mineralogy and compositional uniformity to indicate that the xenoliths represent fragments of a worldwide upper mantle peridotite zone. Contrary to the notion that the xenoliths have segregated from basaltic liquids at depth, they note than in many inclusions, the grain size is considerably larger than that of typical basalt phenocrysts. Also, the Cr_2O_3 contents of the pyroxenes and particularly, the spinels, are too high to have resulted from basaltic liquids usually containing less than 0.1% Cr_2O_3. Spinels, however, from the same xenoliths show large compositional variation. No interpretation of spinel heterogeneity is given. Carter (1969) noted a similar situation in the lherzolites from Kilbourne Hole, New Mexico. The results reported in this paper for the Salt Lake Crater lherzolites also show non-uniform spinels in rocks with relatively uniform olivine, enstatite and chrome diopside. A discussion of this condition is given in Chapter II.

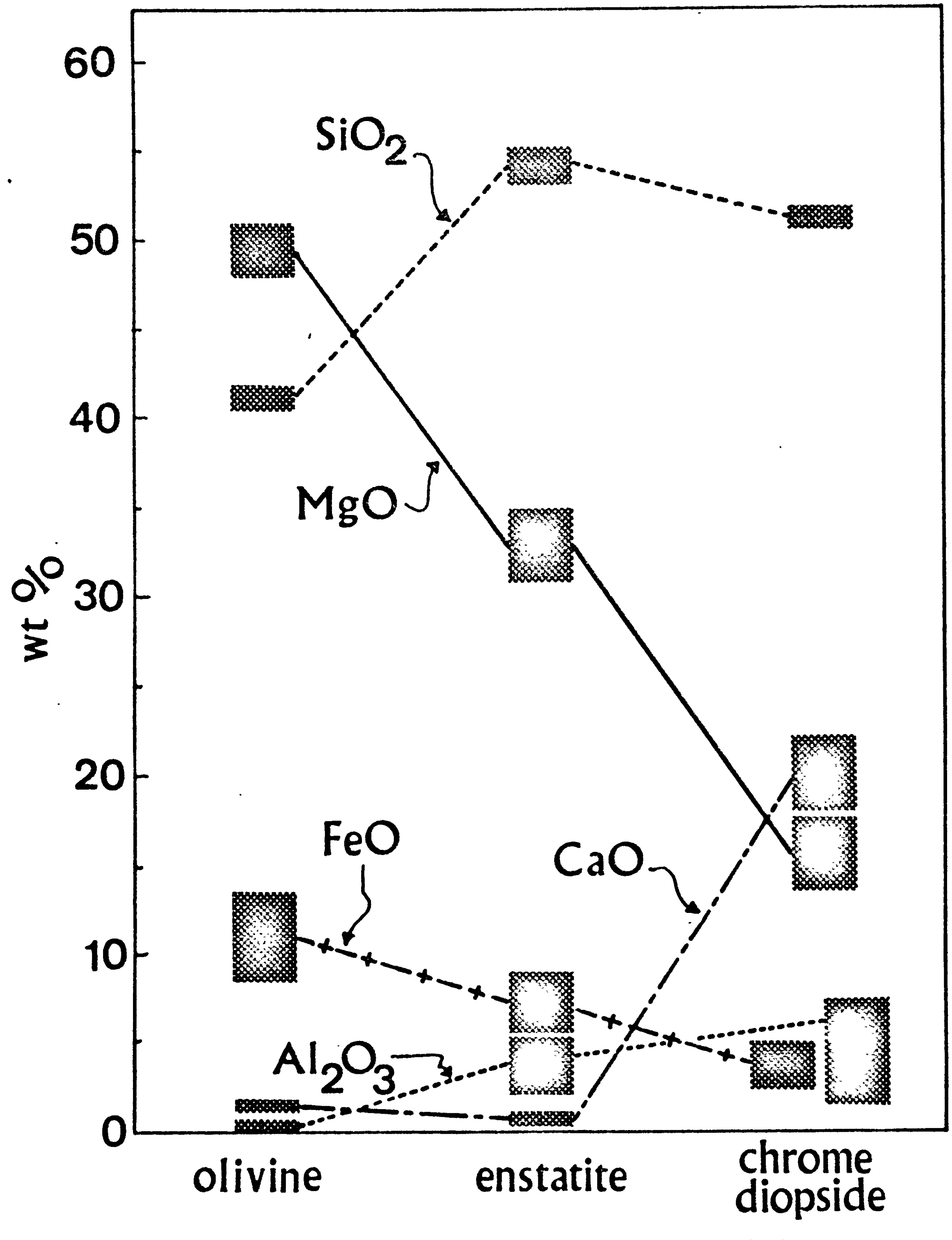


Figure I-1: Ranges in major element composition for the silicate minerals from lherzolite inclusions in basalts. Data include all analyses of Ross *et al.*, 1954, and White, 1966.

The three petrological groups into which the Hawaiian basalts are generally classified (tholeiite, alkali basalt, and highly undersaturated nepheline and melilite-bearing basalts) contain distinctive xenolith suites (White, 1966). Tholeiite inclusions are small and sparse, and are apparently agglomerations of minerals found occurring individually as phenocrysts in tholeiites. The xenoliths in the alkali basalt suite are more varied, and are composed dominantly of dunite, wehrlite (olivine and clinopyroxene), and gabbro. A number of textural and compositional features of this xenolith group suggests they may be cumulates from magma chambers at depth.

In the Hawaiian Islands, spinel lherzolite (olivine, chrome diopside and spinel) occurs preferentially in the late stage undersaturated basalts. White found evidence in the lherzolites arguing against an origin as cumulates from basaltic magma. Unlike cumulate rocks, lherzolite xenoliths show remarkably restricted mineral chemistry and modal proportions. Poikilitic pyroxenes are not found in lherzolites, and plagioclase is very rare. Glass is commonly found in lherzolites as the result of incipient partial melting caused by the enclosing basalt. A cumulate would not be expected to melt in the liquid from which it crystallized without a drastic change in environment. Evidence of solid state recrystallization prior to incorporation in the basalt is common (exsolution in pyroxenes, deformation banding in olivines and pyroxenes). White notes the constant association of lherzolites and very undersaturated basalts, and considers the association as strong evidence favoring a genetic relationship between the inclusions and the host rocks. Harris, Reay and White (1966) note, however, that the undersaturated basalts are more pyroclastic and explosive than alkali basalts or tholeiites, and would be more likely to carry entrained fragments to the surface. Hence, a genetic relationship, although possible, is not necessary. It is also possible that the undersaturated basalts originate at greater depths than the lherzolites and the association could be the result of the presence of lherzolite in the path of the rapidly ascending undersaturated basalt. The very low potassium contents of lherzolites have been noted by several authors (Ross *et al.*, 1954; Kushiro and Kuno, 1963; Oxburgh, 1964), and is felt by White to indicate that lherzolite inclusions probably are not fragments of the primitive upper mantle. Rather, they are residues from fusion or fragments of infusible parts of a heterogeneous mantle.

Recently, Green and co-workers have conducted an extensive geochemical investigation of a selected group of lherzolite xenoliths from Victoria, Australia. Using neutron activation and gamma-ray spectrometry, Green, Morgan and Heier (1968) analyzed the lherzolites and their host basalts for K, Th, and U. They found that the lherzolites differed little from their host rocks in Th/U, but showed marked differences in K/Th and K/U. They state that the large differences in K/Th and K/U can be interpreted in one of two ways. The xenoliths either are mantle fragments incorporated accidently in a genetically unrelated basalt; or, if the xenoliths represent residue related to the host basalt production, potassium is partitioned into the liquid more strongly than either Th or U. The fact that the K/Th and K/U ratios show little variation despite considerable variation in the xenoliths' contents and relative proportions of hornblende, phlogopite and apatite (accessory minerals with expected high K, Th or U contents), favors an accidental origin. Glass-bearing xenoliths do not show element ratios intermediate between glass-free inclusions and the host rocks, suggesting that the lherzolite contents of K, Th and U are not dominated by material absorbed from the host basalts. Some of the lherzolites analyzed contain sufficient U and Th to be considered possible parent material for low-potassium tholeiites. None contains sufficient K to be parental to anything but the very low potassium tholeiites. They are felt to be residue left after the extraction from pyrolite of an undersaturated magma or after the selective removal of minor incompatible elements by reaction of the conduit walls with a permeating melt.

Using the fission track method, Kleeman and others (1969) analyzed nine lherzolites from the same group for their uranium concentrations. Most of the rocks analyzed showed incipient partial melting at contacts between the clinopyroxene and spinel. In the glass formed at these sites, a second clinopyroxene with euhedral outline is found apparently crystallized from the liquid formed from the original spinel and clinopyroxene. A clinopyroxene-liquid distribution coefficient for uranium can be determined from the measured uranium concentrations. When the coefficient is applied to the uranium content of the unaltered original clinopyroxene, hypothetical liquids result having uranium contents 10-100 times the levels found in basalts. This observation is interpreted as evidence that the lherzolites containing these high uranium clinopyroxenes are samples of primordial upper mantle material. A pyrolite upper mantle containing 15-20% high uranium clinopyroxene could give rise to the uranium contents found in various basaltic rock types, when partial melting percentages previously proposed for those rock types by Green and Ringwood (1967) are applied to the parent pyrolite composition. Other lherzolites from Victoria show primary clinopyroxene uranium contents on the order of a few tens of parts per billion; these are interpreted as possible residue from partial melting processes at depth.

Cooper and Green (1970) reported the results of a lead isotopic study of lherzolites from the same suite. They found that the host basalts have Pb206/Pb204, and Pb208/Pb204 ratios with small but real variation, while the corresponding lherzolite ratios show greater variability. The lherzolite ratios range from basanite values to ones depleted in radiogenic Pb206 and Pb208 relative to Pb204. Although the variation could possibly be explained by mixing different proportions of basanite lead and the end-member lherzolite lead, it is inescapable that the lherzolites and basanites show clearly different lead isotopic compositions. They must, therefore, have come from different immediate source areas in the upper mantle. The data preclude a genetic relationship between the peridotites and the basalts.

A number of studies of the strontium isotopic compositions of lherzolites and their host rocks have been made. Most conclude that the lherzolites are genetically unrelated to their host rocks. Leggo and Hutchison (1969) found lherzolite inclusions from the Massif Central, France, having generally higher and more variable Sr^{87}/Sr^{86} ratios than their hosts. They proposed that those with the highest ratios may have been derived from a zone similar to alpine peridotites (characteristically having high Sr^{87}/Sr^{86} ratios), while those with lower $^{87}/^{86}$ ratios may be derived from a postulated pervasive upper mantle peridotite zone. The dissimilarity in Sr isotopic composition between host and inclusions is interpreted as excluding a cognate origin for the lherzolites. Stueber (1969) has analyzed a whole-rock lherzolite inclusion from Camperdown, Australia, and its constituent olivine, enstatite and chrome diopside, for their Rb, Sr, and Sr^{87}/Sr^{86} contents. He found that the data form an isochron plot with a definitely non-zero age. Since the basalt containing the lherzolite is Pleistocene to Recent, the measurably non-zero age for the inclusion is strong evidence for an accidental origin. Stueber and Murthy (1968) were the first to report a strontium study of ultramafic rocks, and found similar Sr^{87}/Sr^{86} ratios in the inclusions and their hosts. Such a result is equivocal in choosing between a cognate and an accidental origin for the inclusions.

Kuno (1969) has analyzed lherzolite and garnet pyroxenite samples from Salt Lake Crater, Hawaii, for their major elements. In contrast to other authors (*e.g.*, Ross *et al.*, 1954) who have stressed the uniformity of lherzolite inclusions, Kuno notes the variation in lherzolite chemistry. They range in MgO/FeO from 6.0 to 2.3, and show systematic variation of the other major elements with the MgO/FeO ratio. He considers the lherzolites fragments of a worldwide upper mantle peridotite zone. Their variation in composition may have been produced by crystal settling in the past, though the estimated composition of the parent melt is unlike any known mafic magma. Kuno suggests that the lherzolites with low MgO/FeO ratios are the potential source of basaltic magmas, and that others may be residual. Ringwood's pyrolite has a composition comparable to an intermediate member of the Hawaiian lherzolite series (Fig. II-19).

The garnet pyroxenites are considered basaltic melts which have been trapped at depth and have undergone subsolidus recrystallization on cooling to their present assemblages. The compostional variation shown by pyroxenites is distinct from the variation in the lherzolite series. Kuno, however, does not stress the observation that together, the lherzolites and pyroxenites form smooth and often linear compositional trends with the pyroxenite group as one end member. The possible significance of this continuity in composition between the two groups is considered in Chapter II.

Carter (1969) has determined major element compositions for spinels from lherzolite and pyroxenite inclusions from Kilbourne Hole, New Mexico. Pyroxenite spinels are Al-rich and Cr-poor, and homogeneous. Lherzolite spinels have widely varying composition within a given xenolith, from relatively Al-rich to Cr-rich. He attributes the wide range in lherzolite spinel composition to their crystallization over a wide range of temperatures or oxygen fugacities. Another possible explanation proposed is the existence of a large solvus in the spinel prism. Although there is a solvus along the hercynite-magnetite join below 860°C (Turnock and Eugster, 1962), formation temperatures of these xenoliths were probably higher (950-1000°C), based on typical lherzolite clinopyroxene composition data. Further, there is no indication of a solvus along the spinel-picrochromite join - the join near which all Kilbourne Hole spinel compositions lie (Muan and Soyima, 1959). The data of Carter's study compare very closely to spinel compositions from lherzolite and pyroxenite xenoliths from Salt Lake Crater (Fig. II-9 of this study) and hence may be related in the manner proposed in subsequent chapters for the Hawaiian inclusions.

Jackson and Wright (1970) have determined major element compositions for xenoliths and host basalts from the Honolulu Series, Hawaii. They find two potential parent-residue pairs in the xenoliths which may be related to the generation of the Honolulu basalts. (1) Pyroxene-rich lherzolite → olivine-rich lherzolite + basalt; and (2) spinel-bearing garnet websterite → garnet websterite + basalt. Neither parent material contains sufficient TiO_2, K_2O nor P_2O_5 to account for the basalt levels of these elements, which may be provided through selective leaching of the wall rock. No xenolith pairs can account for the production of tholeiite in the manner: "parent = basalt + residue". Either the parent material for tholeiites has been entirely consumed, or a multistage process is necessary. Garnet lherzolite, a relatively rare xenolith type, is considered the best parent for the tholeiites.

The following chapters consider the contact between lherzolite and pyroxenite in Salt Lake Crater bimodal xenoliths. Chapter II describes textural and major element data; Chapter III and Chapter IV consider the contributions of rare earth and strontium isotopic data respectively to an understanding of the contact relationship.

CHAPTER II

PETROGRAPHIC RELATIONSHIPS IN SALT LAKE CRATER XENOLITHS CONTAINING LHERZOLITE AND GARNET PYROXENITE IN CONTACT

Introduction

Jackson (1966, 1968) has shown that the xenolith suite at Salt Lake Crater, Hawaii, has a basically bimodal distribution. One group of inclusions is typical spinel lherzolite of the type discussed in Chapter I. Occurring in nearly equal volume are inclusions of garnet pyroxenite, consisting mainly of clinopyroxene and garnet, with smaller amounts of orthopyroxene, spinel, olivine, and often phlogopite. A small proportion of inclusions contain the two materials intimately associated within a single hand specimen. Jackson was first to mention these two-assemblage rocks. Kuno (1970) reports major element compositions for associated lherzolite and garnet pyroxenite from such a xenolith. He interprets the lherzolite as a fragment of the pervasive upper mantle peridotite zone intruded by a melt which crystallized at depth to the pyroxenite assemblage. Jackson and Beeson (in press) note that lherzolite minerals exhibit abundant deformation banding, while associated pyroxenites do not. They do not elaborate, but consider this observation evidence that the lherzolite was deformed prior to its association with the pyroxenite, and hence that it predated the pyroxenite.

The petrography of these two-assemblage xenoliths is reported in three parts. Sections *1* and *2* describe the pyroxenite and lherzolite as they appear in regions well removed from the contacts separating them, and a third section is devoted to the appearance of the contact regions themselves.

1. Lherzolite Portions

A petrographic description of lherzolites from the Hawaiian Islands and from some localities in the western U.S. is given by White (1966, p.263). The lherzolites in this study are similar to the Hawaiian lherzolites of White's study.

Wholly lherzolite xenoliths show a granoblastic fabric of large, fresh, roughly equant olivine anhedra, with smaller amounts of enstatite, chrome diopside and spinel. As described by White, the orthopyroxene occurs in part as large porphyroblasts characterized by relatively small amounts of clinopyroxene exsolution along (100), and by ordered exsolution of thin euhedral flakes of spinel. Small rounded olivine blebs commonly occur within the large enstatites, and small rounded orthopyroxene blebs are found enclosed by olivine. Some included blebs of each extinguish together within a single host grain of the other, though many do not. Interstitial to the large orthopyroxenes and olivines, are smaller grains of the same minerals, and it is commonly found that small, rounded neighboring grains of enstatite separated from one another by olivine do extinguish together. The suggestion is that they may be parts of a formerly larger contiguous orthopyroxene grain. Chrome diopside is generally less abundant than enstatite, and is often found closely associated with or entirely enclosing grains of spinel (Fig. II-1). The spinel is generally fine grained and ameboid in shape, although small euhedral spinels are sometimes found enclosed by chrome diopside, and enstatite as well. Spinel color ranges from greenish-black to reddish brown, in transmitted light. Pyroxene grain size in the intergrowths around spinels is generally smaller (0.1 to 0.3 mm) than the average (1 to 5 mm). Exsolution in chrome diopside is minor, and occurs as thin lamellae, in contrast to the advanced exsolution in pyroxenite clinopyroxenes described in a later section.

Contacts between minerals are generally very sharp, with little crack-filling material between mineral grains. An exception is the observation that the chrome diopside commonly displays a "spongy" periphery (Fig. II-2) resulting from patterns of tiny inclusions of a black isotropic

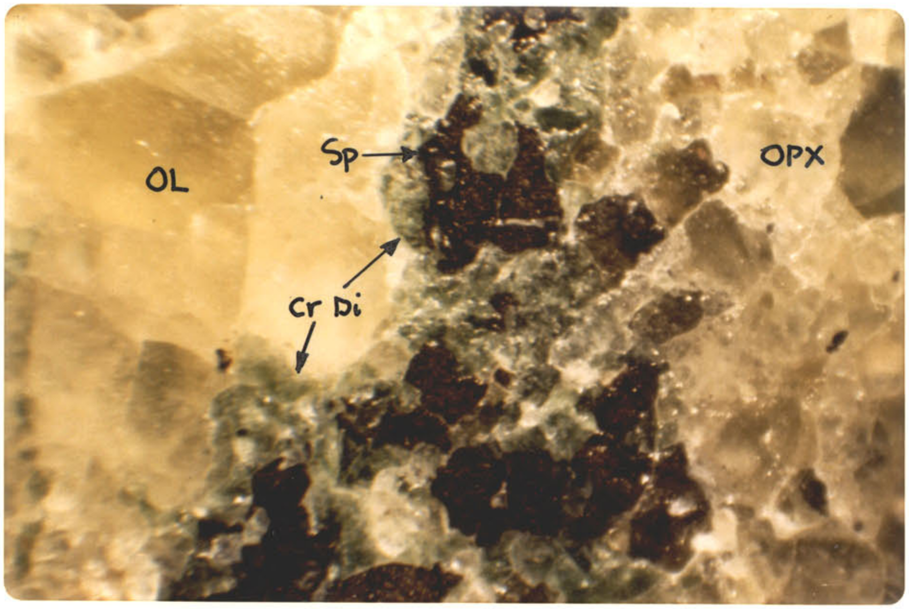


Figure II-1: Chrome diopside enclosing spinel. (Lherzolite portion of R7444; view in reflected light of a polished surface.)

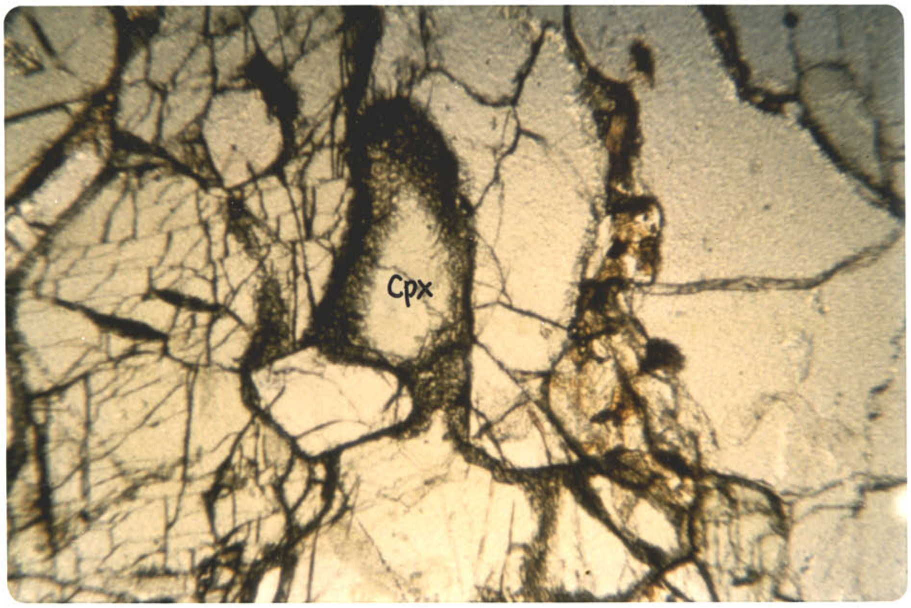


Figure II-2: Lherzolite chrome diopside (cpx) in R7444 showing peripheral inclusions of tiny glass beadlets. (plane polarized light) material whose index of refraction is lower than that of the clinopyroxene. The chrome diopside extinguishes uniformly through such material, which appears to be clusters of glass beadlets showing no optical evidence of altering the host pyroxene. White (1966) and Wilshire and Binns (1961) have reported similar peripheral material in chrome diopsides. In the Australian lherzolites of the latter study, the spongy material appears to be decomposed glass. White found the clinopyroxene depleted in Na and Al in regions near the glass droplets. In the Hawaiian lherzolites, narrow trains of similar glass beadlets are found in the olivine and orthopyroxene. They appear to be fluid droplets trapped along fractures which have since healed. The olivine, and orthopyroxene, like the chrome diopside, show no optical evidence of alteration by this material.

The compositions of spinel grains found interstitial to pyroxenes show remarkably high variation in Cr_2O_3/Al_2O_3 even among spinel grains within a single xenolith. Wide range in spinel composition is a feature both of near-contact lherzolite material (R7444, Table II-1), and xenoliths showing no attached pyroxenite (R7397, Table II-2; also Ross, Foster and Myers, 1954; Carter, 1969). Orthopyroxene grains with enclosed spinels have higher alumina contents (based on qualitative probe scanning) than neighboring grains with no spinel inclusions. The unusual fea-

TABLE II-1. Major element compositions for 4 separate spinel grains from the lherzolite portion of R7444 (Wt.%); Total iron as FeO.

	1	2	3	4
MgO	15.60(±0.48)	15.50(±0.51)	16.81(±0.55)	17.00(±0.48)
Al_2O_3	48.55(±1.50)	47.75(±1.84)	46.96(±1.85)	52.83(±1.92)
Cr_2O_3	17.04(±1.35)	15.25(±0.48)	15.38(±0.49)	11.74(±0.38)
FeO	17.51(±0.20)	21.57(±0.26)	21.99(±0.26)	20.45(±0.24)
TOTAL	98.70	100.07	101.14	102.02
$\frac{Cr_2O_3}{Al_2O_3}$	0.351	0.319	0.328	0.222

TABLE II-2. Major element compositions for 5 separate spinel grains from the lherzolite portion of R7397 (Wt.%); Total iron as FeO.

	1	2	3	4	5
MgO	19.55(±0.45)	20.49(±0.34)	20.29(±0.41)	19.05(±0.52)	19.66(±0.45)
Al_2O_3	49.36(±1.83)	55.20(±1.90)	50.47(±1.84)	46.02(±1.79)	47.92(±1.83)
Cr_2O_3	18.02(±0.31)	12.56(±0.23)	15.03(±0.26)	21.66(±0.36)	17.51(±0.30)
FeO	12.63(±0.16)	11.73(±0.15)	12.35(±0.16)	13.46(±0.17)	12.81(±0.16)
TOTAL	99.56	99.98	98.14	100.19	97.90
$\frac{Cr_2O_3}{Al_2O_3}$	0.365	0.228	0.298	0.471	0.365

ture of wide compositional variation in spinels associated with rather uniform silicates and the presence of deformation textures in lherzolite, while not in pyroxenite, are considered in detail in a later section, for their bearing on an understanding of the relative ages of the lherzolite and pyroxenite.

2. Pyroxenite Portions

The pyroxenites as a group show greater variation in the relative proportions of their constituent minerals than do lherzolites, yet all the observed pyroxenites share a common set of textures. These textures have recorded a large amount of information concerning the rocks' subsolidus histories.

Pyroxenite mineralogies are dominated by two aluminous pyroxenes, with clinopyroxene generally more abundant than orthopyroxene (Jackson and Wright, 1970; Beeson and Jackson, in press). Pyroxene constitutes between 60 and 80% of the rock, with the remainder consisting of pyrope-rich garnet, green aluminous spinel, and generally a small proportion of olivine with a higher fayalite content than lherzolite olivine (White, 1966). Small flakes of phlogopite and blebs of amphibole are present in trace amounts in most pyroxenites, along with traces of pyrrhotite and ilmenite.

Within a given inclusion, clinopyroxenes vary considerably in grain size, in turbidity, and degree of orthopyroxene exsolution. Large clinopyroxenes have irregular anhedral shape, and owe their rather turbid appearance to abundant trains and patches of glass droplets, to tiny discrete grains and exsolution patterns of a dark green-brown spinel, and to the presence of small amounts of included phlogopite and amphibole. Large amounts of exsolved orthopyroxene characterize the large clinopyroxenes, as noted by Yoder and Tilley (1962), Jackson and Wright (1970) and Beeson and Jackson (in press). The exsolution is initially lamellar, but grains showing advanced orthopyroxene exsolution have developed a complicated and often "zebra-like" myrmekitic pattern (Fig. II-3).

Beeson and Jackson have determined the major element chemistry of exsolved orthopyroxenes and their host clinopyroxenes in four pyroxenite xenoliths from Salt Lake Crater. Using the relative modal proportions of exsolved and host material within single pyroxene grains, they have reconstructed the compositions of clinopyroxenes prior to the exsolution of orthopyroxene. When the reconstructed compositions are plotted on the 30 kb enstatite-diopside solvus (Davis and Boyd, 1966) they show equilibration temperatures between 1300 and 1400°C (Fig. II-4). As Beeson and Jackson note, this temperature range is not necessarily

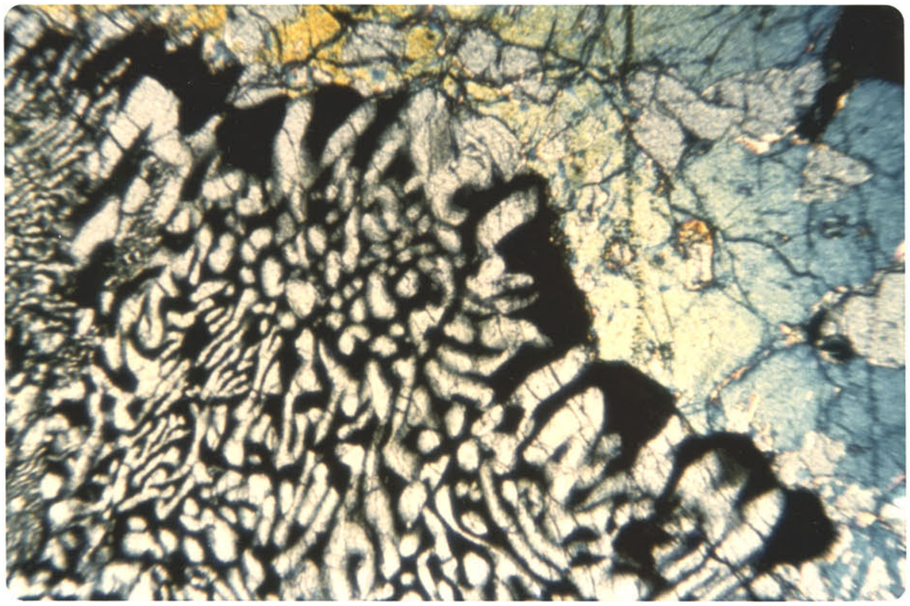


Figure II-3: Pyroxenite clinopyroxene (in R7399-SLC6) showing abundant exsolution of orthopyroxene from host clinopyroxene. Exsolved orthopyroxene appears light; host clinopyroxene dark.

that of the initial equilibration of the pyroxene, but represents a minimum temperature estimate of the oldest environment recorded in the rocks' present textures. If the initial crystallization took place at higher temperatures but was followed by exsolution which formed separate orthopyroxene and clinopyroxene grains, the rock would have a similar appearance. (Salt Lake lherzolite chrome diopsides have very little exsolved orthopyroxene and show equilibration temperatures about 950-1000°C [Fig. II-4; Table II-3]). Beeson and Jackson have also plotted the measured compositions of exsolved orthopyroxene blebs, and their clinopyroxene hosts on the same diagram and have found a re-equilibration temperature range of 1080-1150°C for the pyroxenites. Whether the measured and reconstructed compositions represent a cooling history along the solvus (Beeson and Jackson) or the spinodal as suggested by Wones (personal communication, 1970) is debatable. It is clear in either case that the pyroxenite textures have recorded a history involving temperatures considerably higher than the material had at its time of incorporation into the Salt Lake eruption. The lherzolite shows no evidence of such a high temperature past. The effect of Fe on the En-Di solvus has been studied by Green (unpublished data; personal communication, T. R. McGetchin, 1970). The positions of the solvus for clinopyroxenes with two Fe contents are shown

TABLE II-3. Partial major element analyses for chrome diopsides from R7404, and from lherzolite portion of R7444.

	R7404	*R7444*
SiO_2	47.61	51.47
Al_2O_3	4.23	4.93
FeO	3.53	3.31
MgO	15.60	14.53
CaO	20.08	19.66
TiO_2	0.80	0.15
Cr_2O_3	1.71	1.05
MnO	0.10	0.13
Na_2O	1.87	1.66
TOTAL	95.53	96.89
mol fraction Ca/(Ca + Mg)	0.960	0.985
$\frac{MgO}{FeO+MgO}$	0.823	0.814

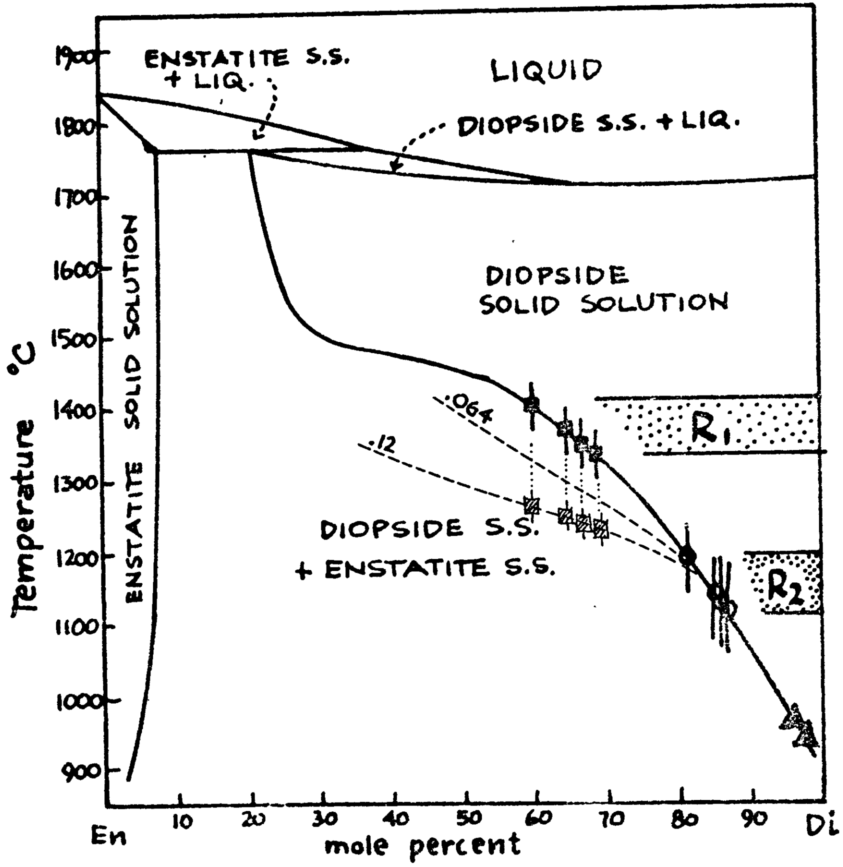


Figure II-4: Compositions of pyroxenite and lherzolite clinopyroxenes plotted on the En-Di solvus (Davis & Boyd, 1966). Solid triangles are lherzolite chrome diopsides (this study). Open circles and solid squares are for pyroxenite clinopyroxenes from Beeson & Jackson, in press (see text). Dashed solvi are for Fe/(Ca+Mg+Fe) = 0.064 and 0.12 (D.H. Green, unpublished data; personal communication, T. R. McGetchin, 1970).

as dashed lines in Fig. II-4. The upper solvus corresponds to Fe/(Ca+Mg+Fe) = 0.064, a value typical of lherzolite chrome diopsides. The lower solvus is for typical pyroxenite clinopyroxenes with Fe/(Ca+Mg+Fe) = 0.12. The difference between initial pyroxenite clinopyroxene crystallization temperatures and the corresponding lherzolite chrome diopside temperatures is lowered by this consideration. The distinction, itself, that pyroxenites contain evidence of a high temperature past, and that lherzolites do not, remains unchanged.

Smaller clinopyroxene grains occur in the same rocks, but are considerably less turbid in general. They show little or no orthopyroxene exsolution, and have sharper and more rounded outlines. These grains may represent a more advanced stage of exsolution - one involving the formation of separate ortho- and clinopyroxene grains, rather than blebs of exsolved orthopyroxene within contiguous clinopyroxenes. This hypothesis is supported by the observation that neighboring small orthopyroxene grains often extinguish in unison, despite being separated from one another by clinopyroxene.

Amphibole and phlogopite occur in trace amounts, and often together within a single pyroxene grain. They almost invariably show the same relationship to the pyroxene host. Both hydrous phases are found either at the contacts between exsolved blebs of orthopyroxene and their host clinopyroxene, or at the edges of exsolved clinopyroxene in orthopyroxene. This constant association suggests a secondary origin for the mica and amphibole. Although inconclusive in some cases, orthopyroxene rather than clinopyroxene seems to be preferentially replaced.

The degree of turbidity in the orthopyroxenes seems to increase less with grain size than in clinopyroxene, and for grains of a given size, orthopyroxene is less turbid than clinopyroxene. Compared with clinopyroxene, orthopyroxene shows considerably less exsolution, as would be expected from the small temperature dependence of composition of enstatite coexisting with diopside (Davis and Boyd, 1966; Fig. II-4). The exsolution is largely confined to thin lamellae along (100), although these lamellae broaden in places to become subhedral blebs of clinopyroxene. Reconstructed original orthopyroxene compositions are not very useful in specifying ancient temperature environments because the enstatite limb of the solvus is nearly vertical.

Other than tiny flakes and grains of spinel which occur as exsolution and inclusions within the pyroxenes, all pyroxenite spinels occur as large, irregular anhedra (up to 3 mm), ranging in color from olive green to black. The spinels are always separated from pyroxene by rims of garnet (Fig. II-5)(Green, 1966; Jackson and Wright, 1970; Beeson and Jackson, in press). Partial rims of olivine

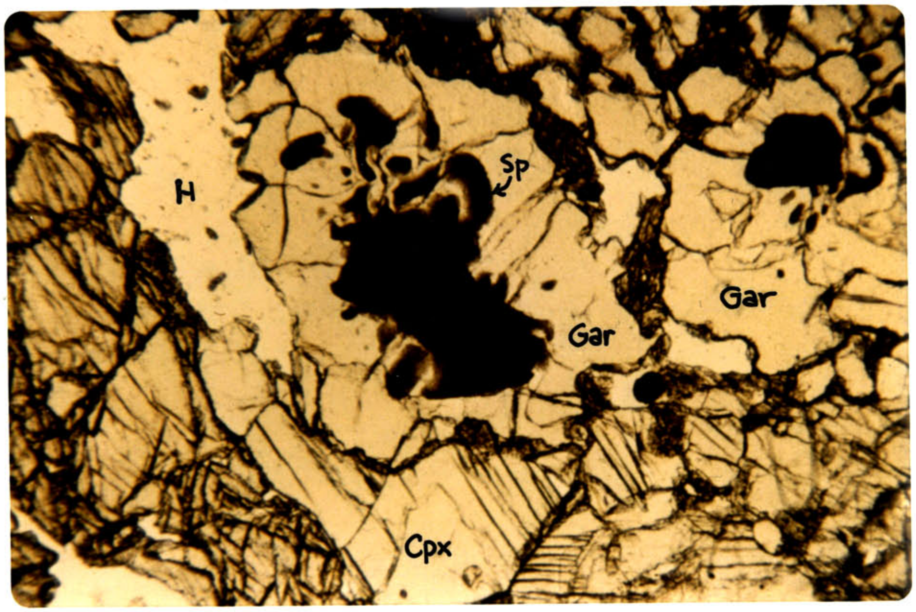
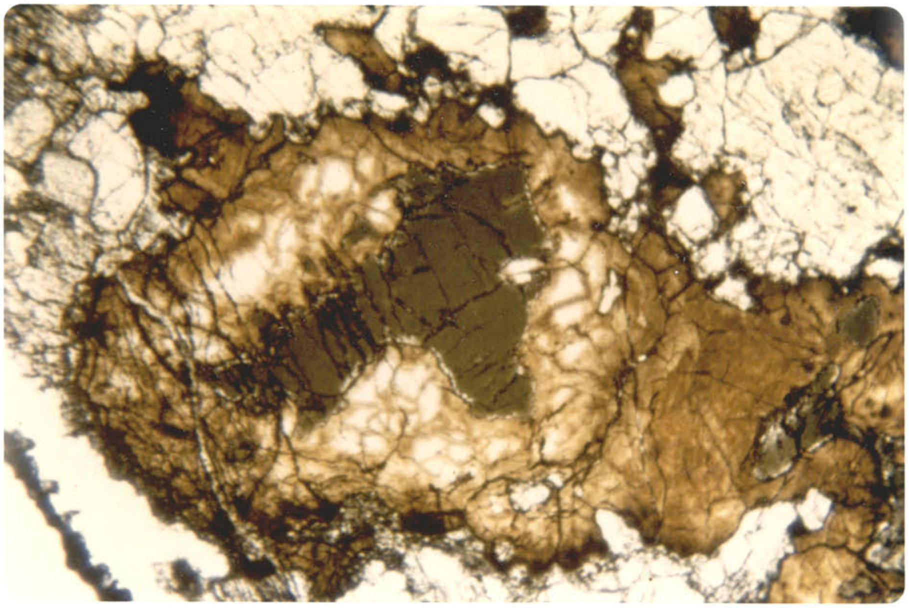
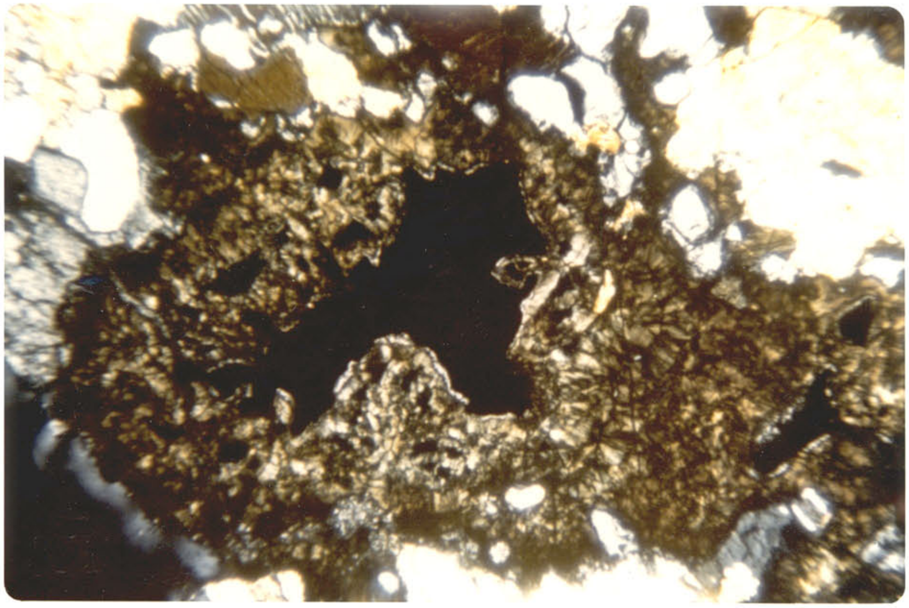


Figure II-5: Garnet coronas surrounding corroded spinels. Clinopyroxene encloses the garnet. (H is a hole in the thin section.)

are found around the garnet in a few cases. Garnet is also found as rounded blebs exsolving directly from pyroxene. Most exsolved garnet in clinopyroxene appears superimposed upon earlier orthopyroxene exsolution lamellae and blebs.

Garnet rarely exhibits clear outlines against neighboring grains. At garnet-clinopyroxene contacts, material is found ranging in color from amber to black, which invariably has an index of refraction less than the pyroxene or the garnet. Under crossed polars, this material ranges from isotropic to fibrously birefringent, and in many cases leaves little of the garnet unaffected (Figs. II-6, II-7). Amber material has a composition determined with the electron microprobe, which is similar to the garnet composition, except for the possibility of a slight enrichment in the K_2O (Table II-4). Black regions, generally showing more complete consumption of the garnet in thin section, are also similar to the garnet in composition, and also show excess K_2O. The analyses must be improved, as discussed in Chapter VI. Both from textural and compositional standpoints, the melt is composed to a very large degree of garnet, with relatively little of the clinopyroxene being involved. In the large garnets surrounding corroded spinels, amber glass has formed along cracks, leaving fresh, unaffected garnet in the regions between the cracks (Fig. II-6). The suggestion is that melting has been initiated by the introduction of a





Figures II-6 & II-7: Appearance in plane polarized light (II-6), and under crossed polars (II-7) of a partially melted garnet corona in R7444 garnet pyroxenite. Note melting occurring preferentially along fractures, leaving fresh garnet in between.

TABLE II-4. Partial major element analyses for garnet and glass in the pyroxenite portion of R7444*.

	1	*2*	*3*	*4*
	Garnet	*Typical Salt Lake garnet*†	*Amber glass*	*Black glass*
SiO_2	38.00	41.26	38.01	41.52
Al_2O_3	21.76	23.16	22.34	22.08
FeO	10.85	13.29	10.28	11.11
MgO	14.98	17.89	15.95	17.03
CaO	4.84	4.87	5.03	4.77
Cr_2O_3	0.13	0.23	0.25	0.17
K_2O	<0.01	...	0.04	0.09
TOTAL	90.57	101.38	91.90	96.77

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* The low sums for columns *1* and *3* are probably due to improper beam focusing, used inadvertently in early analyses.

† From Beeson and Jackson (in press) 68-SAL26 "reacted from spinel".

vapor or liquid along fractures, rather than by the addition of the heat needed to reach the pyroxenite's dry solidus. This notion is reinforced by the observation that the exsolution record of a long subsolidus cooling history is retained in rocks showing evidence of having undergone partial melting.

3. Textures at the Pyroxenite-Lherzolite Contact

Within c. 5 mm of the lherzolite-pyroxenite contact (hereafter referred to as the L-P contact), the lherzolite has a somewhat different appearance. The most striking difference at first glance is the much larger amount of intergranular material and included glass droplets in the lherzolite near the L-P contact. Aside from having more included glass, olivines differ relatively little in appearance in nearing the L-P contact. The largest difference is displayed by the chrome diopside. Near the contact, it is strongly turbid; glass droplet trains and fringes are especially well developed, and patterns of exsolved spinel euhedra (a feature rarely found in lherzolite far from the contact) are common. The clinopyroxene is noticeably less green, is more abundant modally, and shows a higher degree of orthopyroxene exsolution. The large orthopyroxenes also show considerably more exsolved spinel and included glass droplets as the L-P contact is neared. Spinel is characteristically found in close association with or totally enclosed by clinopyroxene and clinopyroxene-orthopyroxene intergrowths, as in lherzolites in general. The modal abundance of spinel seems to decrease with distance from the L-P contact. Qualitative electron microprobe analyses have shown that Mg/Fe in lherzolite olivines varies with distance from the L-P contact. Count ratios for Mg/Fe are shown for olivine and chrome diopside grains in R7444 lherzolite in Table II-5 and Table II-6. Olivine compositions are more Fe-rich at the contact. Chrome diopsides show no systematic variation in Mg/Fe in this preliminary investigation.

A striking texture at the L-P contact involves spinel (Fig. II-8). The figure shows a spinel grain exactly astride the L-P contact in R7444-SLC48. The portion on the left (in the pyroxenite portion) is a green aluminous, low Cr spinel, identical in appearance to other spinels occurring within the pyroxenite portion. The amber to black material surrounding the green part of the spinel grain is the glass formed from a thin garnet corona in the manner described in the previous section. The glass, in turn, is surrounded by the pyroxenite clinopyroxene. Toward the right, the spinel changes color through deep brown to black, with a garnet corona being absent from these portions of the spinel grain. Chrome diopside is found partially enclosing the black portions of the spinel, in a manner similar to regions in

TABLE II-5. X-ray counts (90 sec.) from Mg and Fe spectrometers for 7 olivine grains in R7444L as functions of distance from the L-P contact.

Mg	*Fe*	*Mg/Fe*	*Distance from contact*
31262	20236	1.545	less than 1 mm
33044	20789	1.589	" "
32778	20626	1.589	" "
32119	20410	1.574	" "
33867	19099	1.773	about 5 mm
34355	19007	1.807	" "
33726	19393	1.739	" "

TABLE II-6. X-ray counts (90 sec.) from Mg and Fe spectrometers for 5 chrome diopside grains in R7444L as functions of distance from the L-P contact.

Mg	*Fe*	*Mg/Fe*	*Distance from contact*
11042	6428	1.717	less than 1 mm
10739	6435	1.668	" "
10205	6079	1.678	about 3 mm
10688	5334	2.003	" "
10534	5967	1.765	about 5 mm

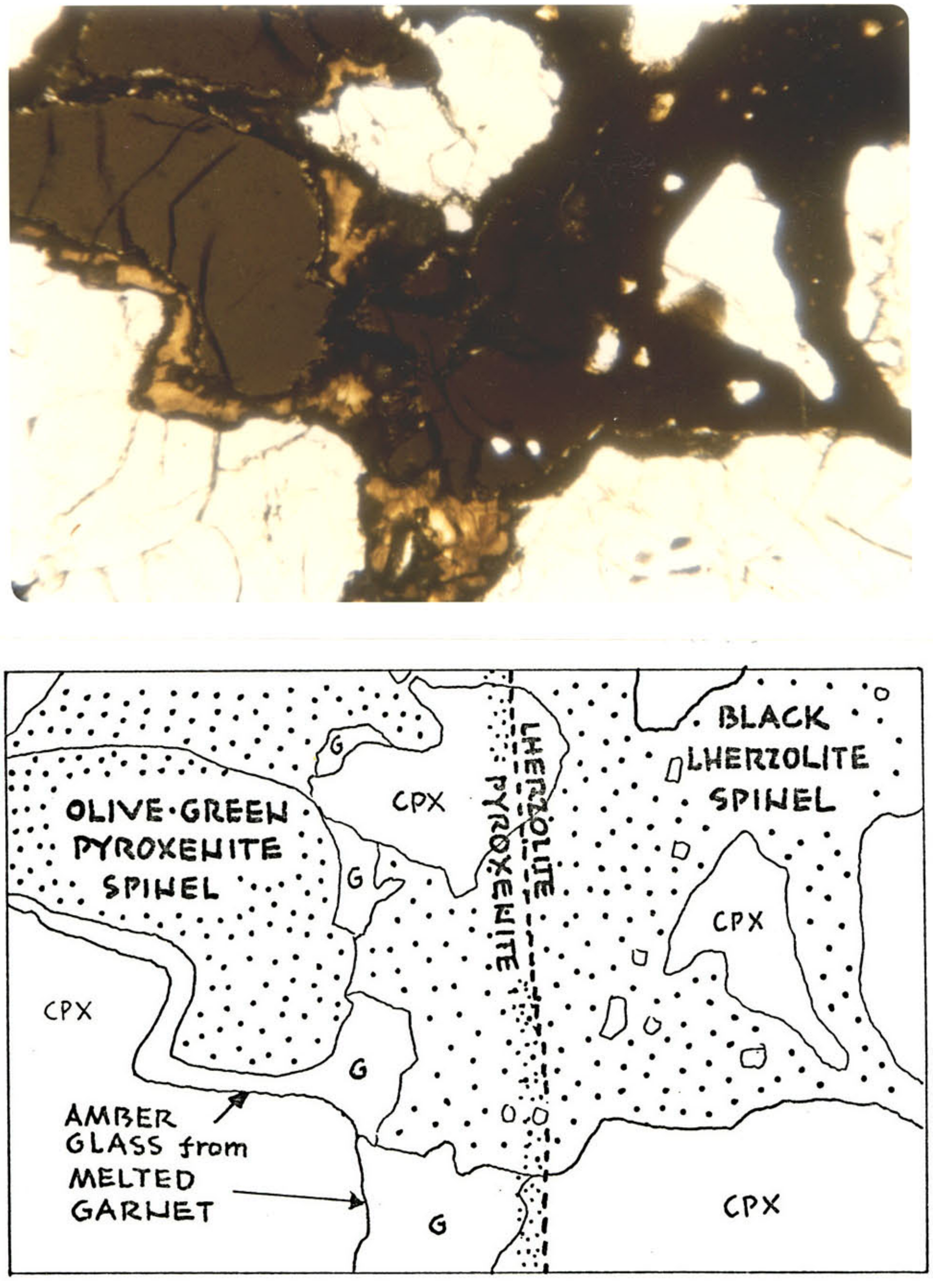


Figure II-8: Appearance of a spinel grain astride the lherzolite-pyroxenite contact in R7444. (see text) wholly lherzolite xenoliths. Microcrystalline chrome diopside grains are found within the black spinel, a feature found very rarely farther within the lherzolite. Electron microprobe analyses of both types of spinel grains near the L-P contact in R7444-SLC48 (Table II-7) show the green pyroxenite spinel to be homogeneous with a composition about $(Mg_{.82}Fe_{.18})(Al_{.89}Fe_{.09}Cr_{.02})_2O_4$. The black grains are heterogeneous, and considerably richer in Cr_2O_3 and poorer in Al_2O_3 than the green spinel. Table II-8 and Fig. II-9 show the variation in Cr_2O_3, Al_2O_3, and Fe_2O_3 for spinels (assumed ideal) in the pyroxenite and lherzolite portions of R7444-SLC48. The noteworthy feature is that the two groups of spinels follow a smooth trend taken together. Carter (1969) has analyzed spinels from a group of lherzolites and pyroxenites from Kilbourne Hole, New Mexico, and has found a nearly identical distribution in spinel compositions from green low-Cr spinels to brown-red high-Cr spinels in the lherzolites. The superposition of the Salt Lake Crater and Kilbourne Hole spinel data (Fig. II-9) suggests that whatever process related the lherzolites and pyroxenites at Salt Lake Crater, a similar process has probably been active in the development of the material found as inclusions at Kilbourne Hole.

Two types of pyroxene grains appear along the L-P contact. Figs. II-10 and II-11 show the appearance in plane

3LE II-7. Major element analyses for spinels from the pyroxenite and lherzolit portions of R7444.

	Pyroxenite spinels			*Lherzolite spinels*			
	1	*2*	*3*	*1*	*2*	*3*	*4*
)	22.06	22.94	20.60	15.60	15.50	16.81	17.00
$_2O_3$	57.75	57.78	57.12	48.55	47.75	46.96	52.83
$_2O_3$	2.28	2.16	2.46	17.04	15.25	15.38	11.74
)	14.49	13.61	14.26	17.51	21.57	21.99	20.45
ΓAL	96.58	96.49	94.44	98.70	100.07	101.14	102.02

TABLE II-8. Calculated molecular percentages of trivalent cations in spinels from lherzolites and pyroxenites from Salt Lake Crater. For calculation purposes, spinels were assumed ideal (RO R_2O_3).

Source		*Fe*$^{+3}$	*Al*$^{+3}$	*Cr*$^{+3}$
R7444 Pyroxenite	1	9.86	87.82	2.32
	2	9.50	88.19	2.30
	3	8.25	89.17	2.58
R7444 Lherzolite	1	2.39	79.02	18.59
	2	6.40	77.09	16.51
	3	8.77	74.79	16.44
	4	5.88	81.91	12.21
R7397 Lherzolite	1	3.16	70.00	25.55
	2	2.52	84.58	12.90
	3	4.42	79.66	15.91
	4	3.62	73.25	23.12
	5	3.87	77.52	18.61

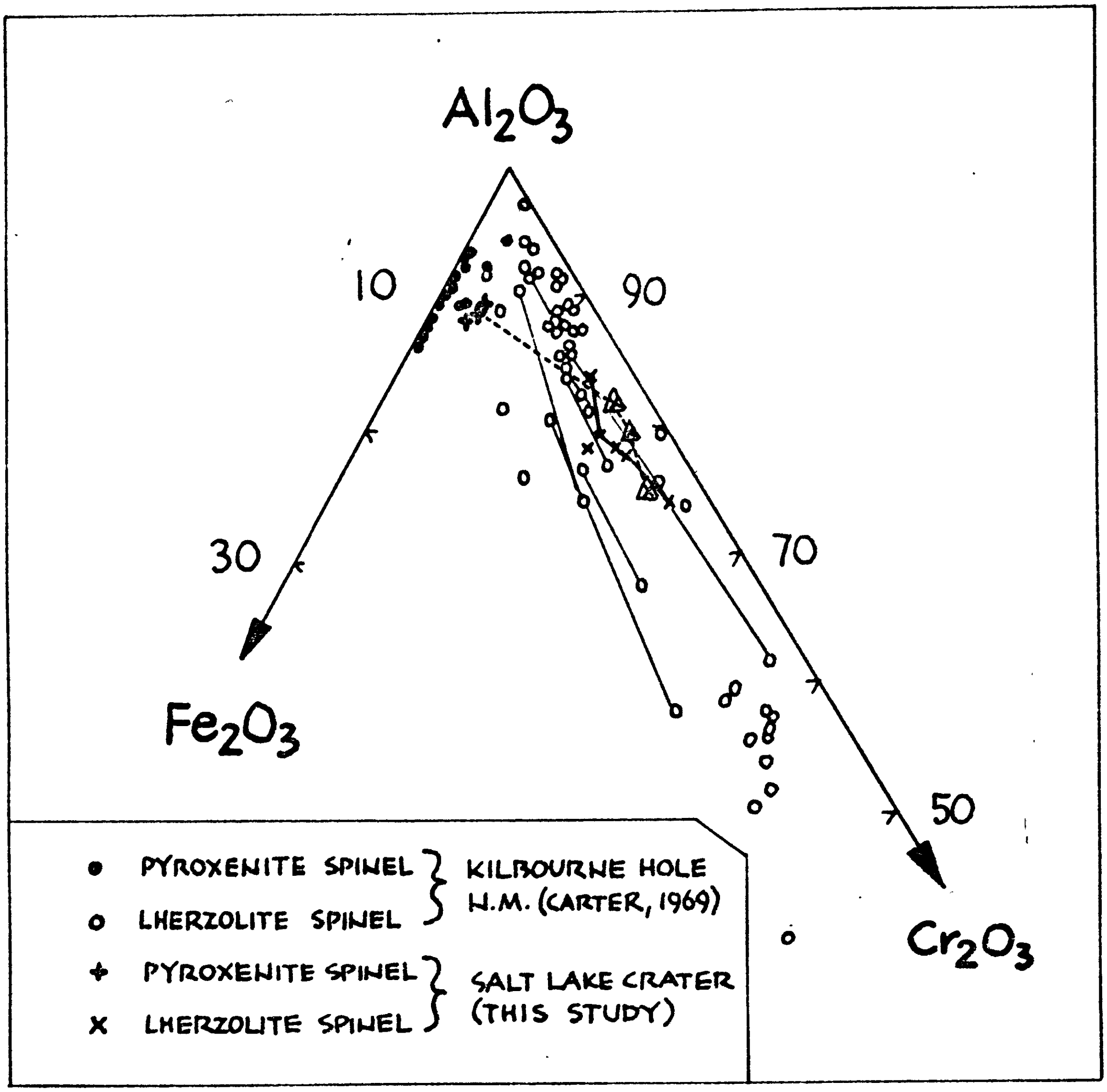


Figure II-9: Mole fraction of Fe_2O_3, Al_2O_3 and Cr_2O_3 in spinels from pyroxenites and lherzolites from Kilbourne Hole and Salt Lake Crater. Tie lines connect spinels from single xenoliths. Dashed tie line is for R7444-SLC48.

polarized light and under crossed polars respectively of a Type I pyroxene grain astride the L-P contact in R7629-SLC76. In Fig.II-10, it is seen that the left side of the grain (the part in the pyroxenite) is more turbid. Under crossed polars (Fig. II-11), the turbid portion of the grain is found to be clinopyroxene with a high degree of orthopyroxene exsolution. The clearer part of the grain (in the lherzolite) is entirely orthopyroxene, and is crystallographically continuous with the orthopyroxene exsolution in the turbid pyroxenite portion of the grain. The portion of the grain which is entirely orthopyroxene shows undulatory extinction and deformation banding, neither of which are displayed by clinopyroxene portion of the same grain.

A second type of pyroxene grain (Type II) is entirely orthopyroxene. The portions of these grains lying in the pyroxenite portion are again more turbid, and contain exsolved spinel and clinopyroxene. Their appearance is very similar to orthopyroxene grains well within the pyroxenite. The portions of these grains in the lherzolite, are less turbid, and contain little or no exsolution of either kind (Fig. II-12). Rounded olivine grains are found both as inclusions and embayments in the portions of these orthopyroxene grains within the lherzolite. In some instances, single olivine grains along the L-P contact separate small orthopyroxene grains all of which extinguish

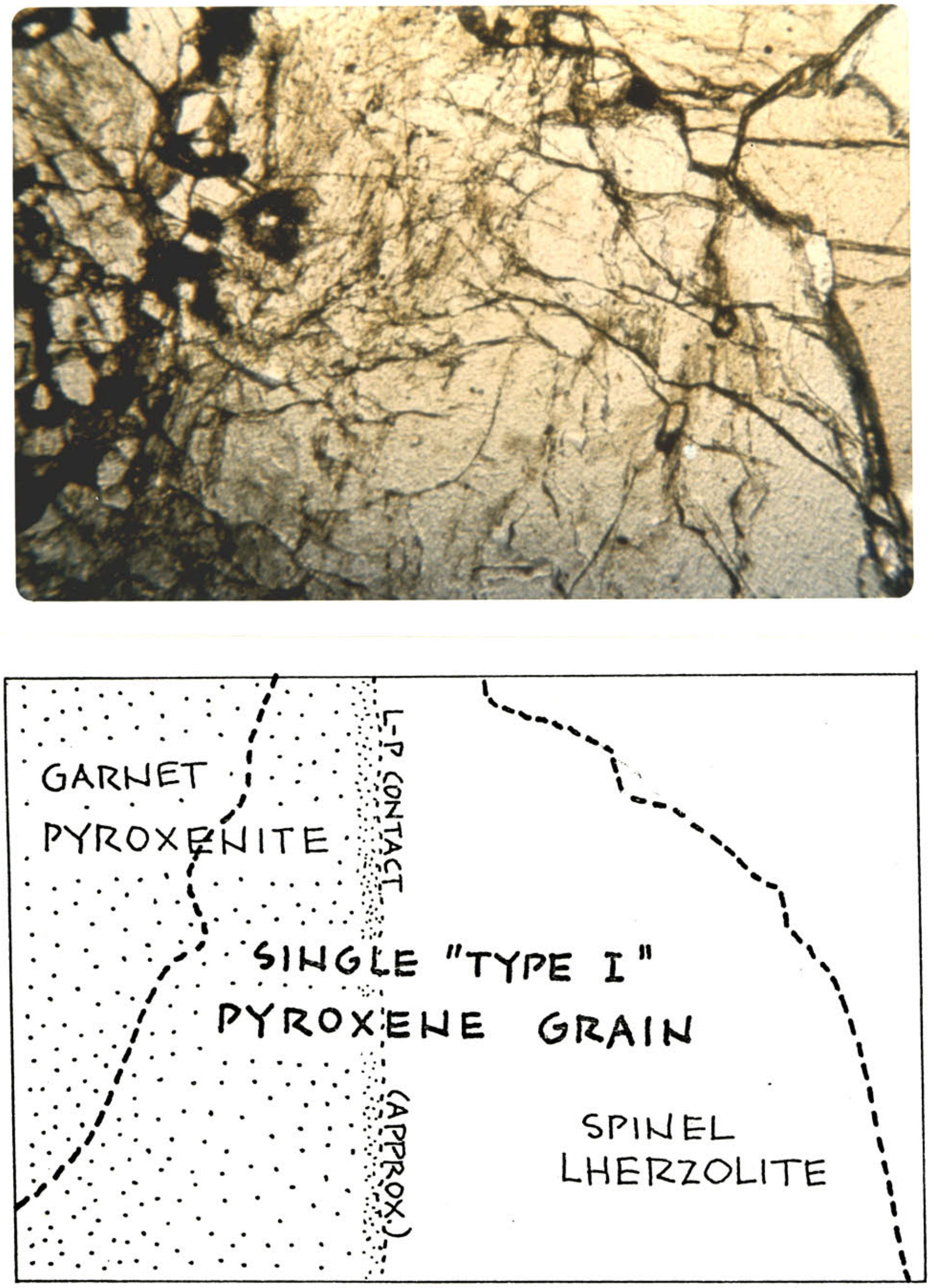


Figure II-10: Appearance in plane polarized light of a "type I" pyroxene grain astride the lherzolite-pyroxenite contact (R7629-SLC76). (Figure II-11 shows the same grain viewed under crossed polars.)

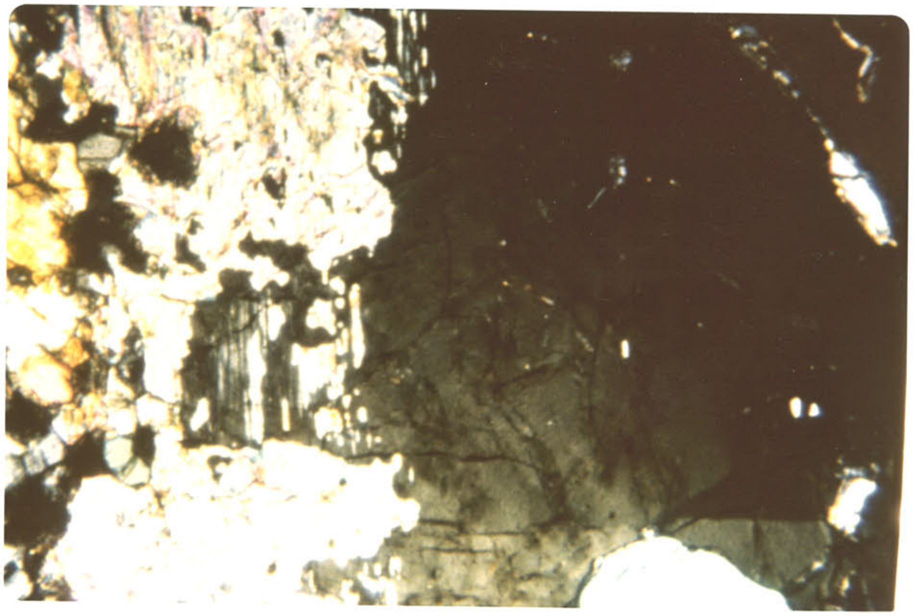


Figure II-11: Appearance under crossed polars of a "type I" pyroxene grain astride the lherzolite-pyroxenite contact. (Figure II-10 shows the same grain in plane polarized light.)

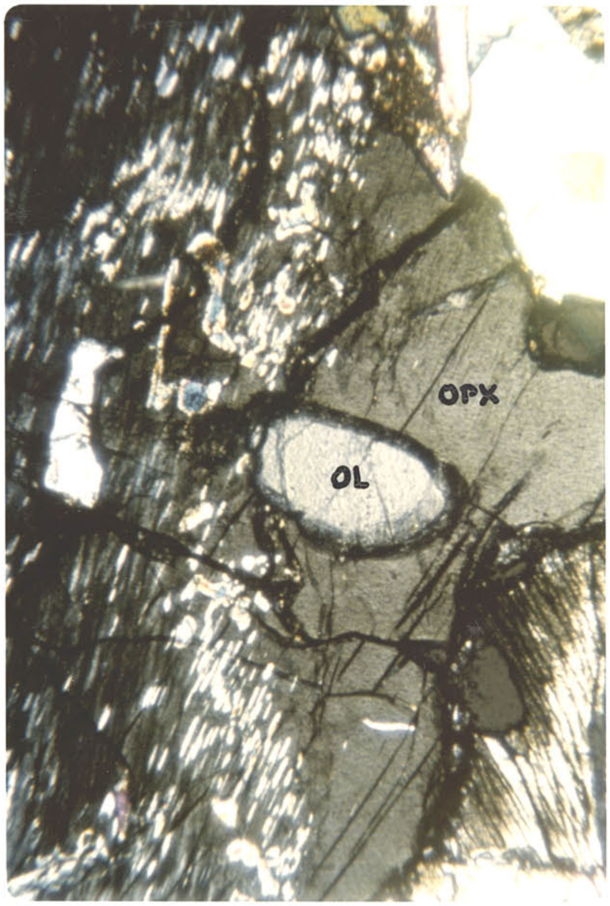


Figure II-12: Appearance of a "type II" pyroxene grain astride the lherzolite-pyroxenite contact (R7629-SLC76). The contact runs vertically through the middle of the photograph, with the pyroxenite portion on the left, the lherzolite on the right.

in unison and have the same interference figure (Fig. II-13). The suggestion is that the orthopyroxene grains are parts of a formerly larger contiguous grain, which may have been partially consumed in the growth of the intervening olivine.

Interpretation of the Textures

The goal of the textural study has been to reconstruct the histories of the rocks. The discussion which follows considers the history of the pyroxenites separately, and then evaluates the three possible age relationships between the pyroxenite and the lherzolite.

1. Petrologic Histories of the Pyroxenites

Green and Ringwood (1967) have studied the stability fields of aluminous peridotite in the pressure range 10-40 kb. Although the pyrolite composition differs considerably from a pyroxenitic composition, their experiments were performed with materials having a pyrolitic composition less 40% olivine, for ease in identification of products. Olivine shows very limited solid solution with pyroxene; thus, olivine need only be present but not necessarily in the pyrolitic proportion (about 60%) for the results to be applicable to the pyrolitic composition. The

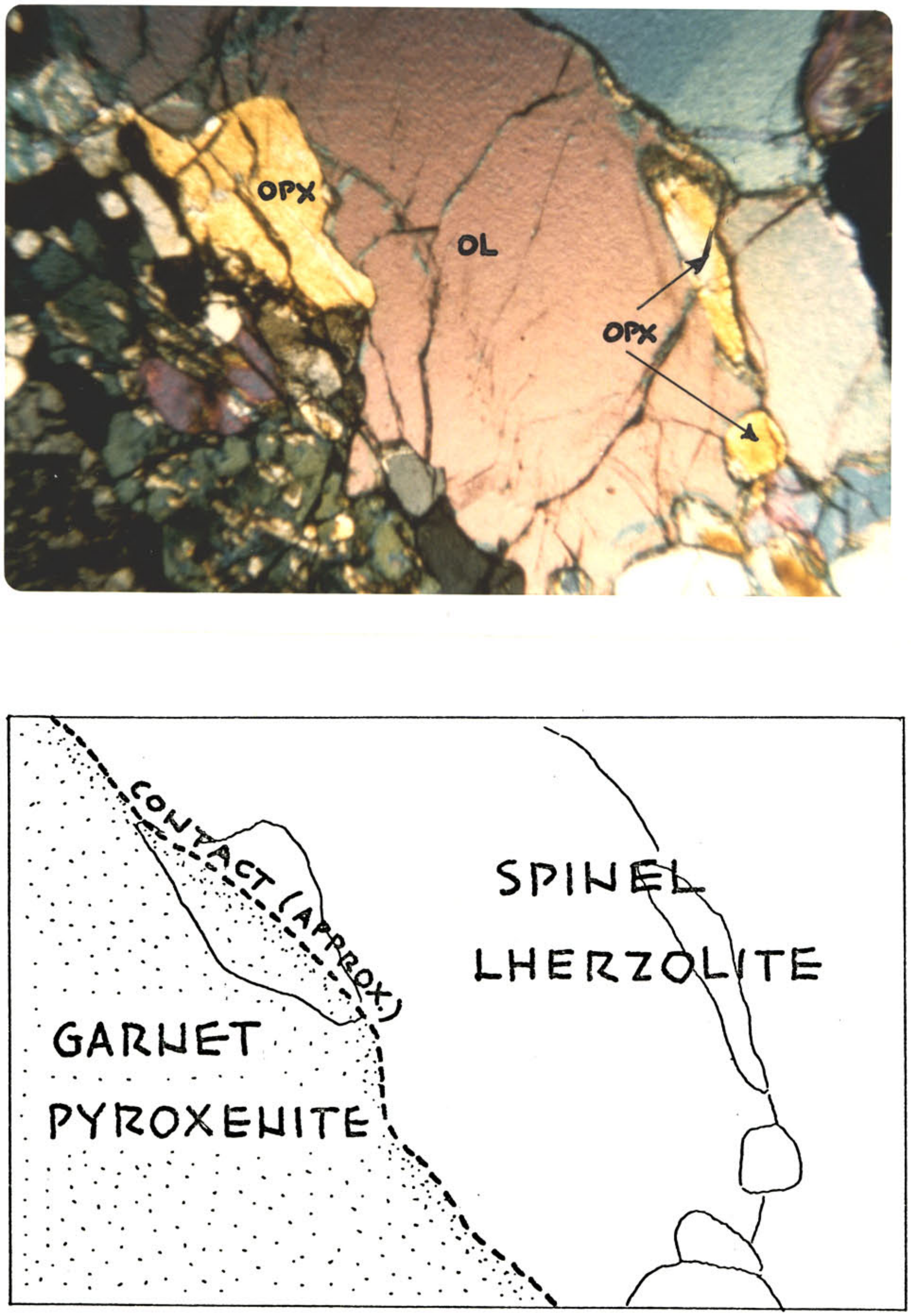


Figure II-13: Near-contact olivine grain separating three small orthopyroxene grains, which extinguish in unison (see text).

material actually analyzed by Green and Ringwood has a more nearly pyroxenitic composition (Table II-9)and as seen in Fig. II-18 other Salt Lake pyroxenites (Kuno, 1969) have rather similar compositions.

Fig. II-14 is a reproduction of the temperature-pressure diagram determined for pyrolite III by Green and Ringwood. If the same diagram applies to the pyroxenitic material from Salt Lake Crater, limitations can be placed on the pressure and temperature regimes from which these inclusions came. (The corrected pyrolite composition has a higher Cr_2O_3/Al_2O_3[0.113] than pyroxenite [0.043]. The effect of this difference [MacGregor, in press] is to move subsolidis boundaries about 1.5 kb lower for the pyroxenite, than for the pyrolite to which the diagram applies.)

None of the pyroxenites contains primary plagioclase. (In regions of advanced formation of melt from the garnet, small plagioclase laths are occasionally found in a microcrystalline aggregate along with abundant opaques. These probably formed along with the more common glass at or near the surface.) At pressures less than P_1 (Fig. II-14) plagioclase is a primary phase, and hence a lower pressure limit can be set for the pyroxenites of about 10 kb. None of the pyroxenites of this study contains what is unquestionably primary garnet - all the garnet having formed either by direct exsolution from the pyroxenes or as coro-

TABLE II-9. A comparison of major element compositions for Pyrolite III, Pyrolite III-40% olivine*, and Salt Lake Crater garnet pyroxenite R7444.

	Pyrolite III	*Pyrolite III—* *40% olivine*	*R7444*
SiO_2	45.20	48.23	48.46
Al_2O_3	3.54	5.73	8.96
Fe_2O_3	0.48	0.80	9.34**
FeO	8.04	7.20	
MgO	37.48	19.50	22.97
CaO	3.08	5.05	7.39
TiO_2	0.71	1.18	0.45
Cr_2O_3	0.43	0.65	0.38
Na_2O	0.57	0.95	0.90

* Olivine composition used is from Ross, Foster and Myers (1954)(olivine #4, p. 707).

** Total iron as FeO.

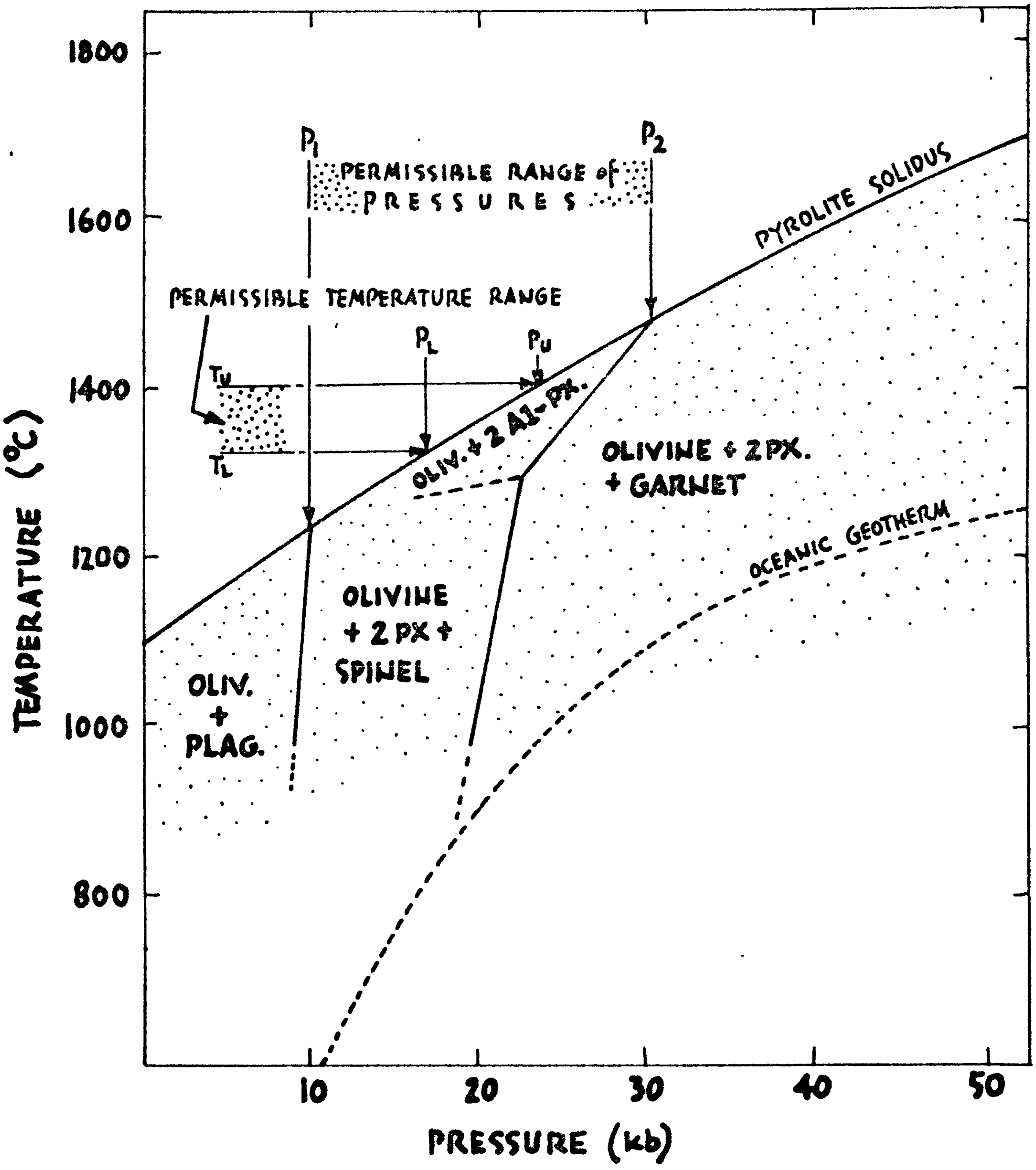


Figure II-14: Subsolidus phase relations for pyrolite III (after Green and Ringwood, 1967). See text for explanation of permissible temperature and pressure ranges.

nas around spinels (Fig. II-5). Above P_2, garnet is a primary phase, indicating that the pyroxenites have originated at pressures between 10 and 30 kb. (Beeson and Jackson report Salt Lake pyroxenites with some apparently primary garnet, but all these xenoliths also contain garnet formed from spinel, so that the upper pressure limit for their xenoliths cannot be much greater than 30 kb.) When the reconstructed equilibration temperatures of the clinopyroxenes determined by Beeson and Jackson are plotted on the diagram of Green and Ringwood, it is found that the lower and upper temperature limits intersect the solidus at pressures (P_L and P_U) lying entirely within the range set by the mineralogy (Fig. II-14). If the permissible pressure range is considered to be from P_L to P_U (about 18 to 22 kb) it is interesting to compare the actual textures and mineralogy of the pyroxenites with a predicted set of textures which might result from a cooling history from 1300-1400°C to about 1100°C. (The effect of higher Cr_2O_3/Al_2O_3 in the corrected pyrolite, than in the pyroxenite is to move subsolidus boundaries (Fig. II-14) to slightly lower pressures. The effect of Fe on the En-Di solvus is to lower the temperatures of initial equilibration for pyroxenite clinopyroxenes determined from the solvus in the Fe-free system. However, the two effects do not alter the conclusions reached, other than to lower by about 5 km the estimate of the depths from which the xenoliths may have come.) The initial assemblage would be dominated by two aluminous pyroxenes, with small amounts of olivine, and, in the cases of high Al_2O_3/SiO_2, small amounts of spinel. Depending on the pressure, isobaric cooling will cause one of two phase boundaries to be crossed. Spinel (lower pressures) and garnet (higher pressures) would be expected to form at the expense of aluminous pyroxene, perhaps by exsolution. Continued cooling in the lower pressure range would cause garnet to form at the expense of spinel and pyroxene by a reaction of the general form:

Al-pyroxene + spinel = garnet ± olivine ± orthopyroxene

(MacGregor, 1964; Green, 1966; Ito and Kennedy, 1967; Green and Ringwood, 1967; Jackson and Wright, 1970). The anticipated assemblage would be two pyroxenes, garnet, olivine, and perhaps some remnant spinel within regions of garnet. The observed textures in the Salt Lake pyroxenites closely fit this expected mineralogy. Spinel and garnet are found as exsolution from both pyroxenes, and garnet coronas everywhere surround primary spinel grains.

Hence, the Salt Lake pyroxenite suite originated at depths of 60 to 75 km in the upper mantle, where they underwent an extended cooling history from about 1300°C to 1100°C prior to being brought to the surface in the Salt Lake eruption.

2. The Relative Ages of Pyroxenite and Lherzolite

It will be helpful to summarize the textural observations:

[1] The pyroxenites formed at temperatures about 1300°C and underwent extensive cooling at depth, while the lherzolites formed at about 950°-1000°C and show evidence of very little pre-eruptive cooling.

[2] Single contiguous spinel grains at the L-P contact show wide compositional range. The portion in the pyroxenite is green, low Cr/Al, and shows garnet coronas. The lherzolite portion is black, high Cr/Al and shows no garnet corona, but is mantled by and contains chrome diopside grains (Fig. II-8).

[3] Spinels within wholly lherzolite xenoliths are also intergrown with chrome diopside. These spinels also show wide range in composition colinear with the variation found at the L-P contact (R7397, Table II-4; Fig. II-9).

[4] Type I pyroxene grains astride the L-P contact are turbid clinopyroxene with abundant orthopyroxene exsolution in the pyroxenite portion (Figs. II-10, 11). The portion of the same grain lying in the lherzolite is clearer exsolution-free orthopyroxene continuous crystallographically with the orthopyroxene exsolution in the pyroxenite part of the grain. Deformation banding is present in the lherzolite orthopyroxene, not in the pyroxenite clinopyroxene.

[5] Wholly orthopyroxene grains (Type II) astride the L-P contact show turbidity, clinopyroxene and spinel exsolution lamellae in the pyroxenite portion, while parts of the same grains in the lherzolite portion are clearer orthopyroxene, showing little or no exsolution and containing rounded olivine blebs and embayments (Fig. II-12). Orthopyroxenes well within the pyroxenite, and others well within the lherzolite, have appearances similar to their respective portions of grains astride the L-P contact.

[6] At the L-P contact and within lherzolite portions, single olivines commonly separate small orthopyroxene grains which extinguish in unison, appearing to be parts of formerly larger contiguous grains (Fig. II-13).

[7] Clinopyroxenes in the lherzolite portions become less turbid, contain less orthopyroxene exsolution, and become more green (Cr-rich) with distance from the L-P contact. Their appearance near the L-P contact is very similar to that within regions of pyroxenite. The modal abundance of clinopyroxene decreases with distance from the L-P contact (in the lherzolite).

[8] Trains of glass droplets abound in lherzolite near the L-P contact, but no alteration of the host silicates is microscopically observable. Far less intergranular material is found in lherzolite far from the L-P contact.

[9] Olivines become richer in Mg relative to Fe with distance from the L-P contact (Table II-5).

[10] Garnet has formed around the aluminous spinels in pyroxenite portions, and has melted to varying degrees (Figs. II-6, 7); no garnet nor evidence of melting is found in the lherzolite.

[11] Lherzolite minerals, both at the contact and within the lherzolite, show more compositional variation than minerals from attached portions of pyroxenite.

[12] Lherzolites show deformation textures; pyroxenites do not.

Simultaneous Origin

The 300-400°C discrepancy between the calculated temperatures of initial equilibration of the clinopyroxenes from lherzolite and pyroxenite (Fig. II-4) is sufficiently difficult to reconcile with the notion that the two materials formed simultaneously, that this possibility is not considered further.

Lherzolite Predates the Pyroxenites

Kuno (1969) reports major element compositions for a pyroxenite ("olivine eclogite" HK61082601c) and an enclosed region of lherzolite (HK61082601c-B) occurring in the same hand specimen. He interprets the lherzolite to be a fragment of the pervasive upper mantle peridotite zone incorporated in a basaltic liquid which was trapped and crystallized at depth.

Under this supposition, it is interesting to reconsider the textures, particularly those at the L-P contact. Textures in lherzolite having no attached pyroxenite, must be considered inherent properties of the upper mantle peridotite zone. On the other hand, the presence of similar textures at the L-P contact might be thought to have resulted from an interaction between the fragment and the melt. If so, the similarities are coincidental. For example, spinels in wholly lherzolite xenoliths the world over show extreme compositional variation (Ross, Foster and Myers, 1954, Carter, 1969), despite the relatively uniform silicates associated with them. Spinel grains astride the L-P contact (Fig. II-8) could be thought to result if a heterogeneous Cr-rich lherzolite spinel on the fragment's exterior acted as the nucleation for the crystallization of spinel from the more aluminous liquid. The colinearity of compositional trends in spinels at the L-P contact, and those within lherzolites must, in this view, be coincidental (see items 2 and 3 in the textural summary above). The presence of chrome diopside in and around spinel might also have been explained as the interaction between the liquid and the fragment's spinel, were it not for the similar mantled spinels throughout lherzolites.

The presence of garnet around the aluminous portions of contact-straddling spinels, and its absence around the Cr-rich lherzolite portions of the same grains presents some serious difficulty for the view that pyroxenites are entrapped melts in host lherzolite. As shown earlier, garnet forms from the solid state recrystallization of spinel and clinopyroxene. MacGregor (in press) demonstrated that increases in spinel Cr/Al ratios increase the pressure (for constant temperature) at which the garnet reaction occurs. The absence of garnet from the Cr-rich spinels might, therefore, imply that the pressure was sufficient to form garnet around aluminous spinels, but not around the Cr-rich spinels. However, garnet formation takes place *after* spinel crystallization and requires ion diffusion in the solid state. The first garnet forms where spinel and clinopyroxene are in contact. Very little ion migration is needed. However, garnet coronas may be 1 mm wide, and hence additional garnet at this stage can only form if ion diffusion in clinopyroxene, garnet and spinel can take place. If ion diffusion in spinel must occur to form garnet, it is not reasonable that the spinel in Fig. II-8 could have maintained its compositional gradient (no ion migration) during garnet formation. (One end has a Cr_2O_3 content about 2.4%. The other end of a grain no more than two mm long, as about 20% Cr_2O_3.) It seems, therefore, that the development of the spinel heterogeneity postdates the formation of garnet. This would avoid the additional problem, that maintenance of a high compositional gradient is unlikely at temperatures of 1300°C (the initial crystallization temperatures for the pyroxenite)..

Two types of pyroxene grains are found astride the L-P contact. Type I shows clinopyroxene with abundant orthopyroxene exsolution in the pyroxenite portions, and clear exsolution-free orthopyroxene in the lherzolite (Figs. II-10, 11). The clear orthopyroxene is crystallographically continuous with the exsolved orthopyroxene in the pyroxenite portions (Fig. II-12). The clinopyroxene appears similar to clinopyroxene well within the pyroxenite. As in Type I, Type II grains show clear orthopyroxene in their lherzolite portions. However, the parts of Type II grains in the pyroxenite have the appearance of pyroxenite *orthopyroxenes*. These observations are difficult to reconcile with the idea that lherzolites are xenoliths in melts trapped at depth. Ostensibly, the turbid, exsolution-rich portions of such grains of both types might have resulted from an interaction between the liquid and peripheral orthopyroxenes of the lherzolite fragment. If so, it is unreasonable that some grains would acquire the appearance of pyroxenite *clinopyroxenes*, while others assumed the appearance of pyroxenite *orthopyroxenes*. Moreover, because of the pyroxenite's initial temperatures of about 1300°C, it is unlikely that the pyroxenes would have been affected to depths of less than 1 mm, or that the compositional gradients could have been maintained.

Also inconsistent with the hypothesis that pyroxenite postdates the lherzolite is the observation that none of the lherzolite minerals shows any evidence of melting. Trains of glass droplets are indeed found in lherzolite minerals, but they show little alteration of the host grains. Kushiro *et al.* (1968) determined the solidus for a Salt Lake Crater lherzolite xenolith under anhydrous consitions and saturated with water (Fig. II-15). The wet solidus crosses the 10-30 kb range at about 1000°C. If the supposed melt from which the pyroxenite crystallized contained appreciable water, that melt could have been as much as 300°C above the solidus of the lherzolite fragment being incorporated. The absence of melting evidence is very difficult to explain. Even if the lherzolite fragment had not undergone partial melting, its reaching a temperature of 1300°C was unavoidable. The increased solubility of enstatite in diopside at elevated temperature should have caused the homogenization of clino- and orthopyroxenes at mutual contacts in

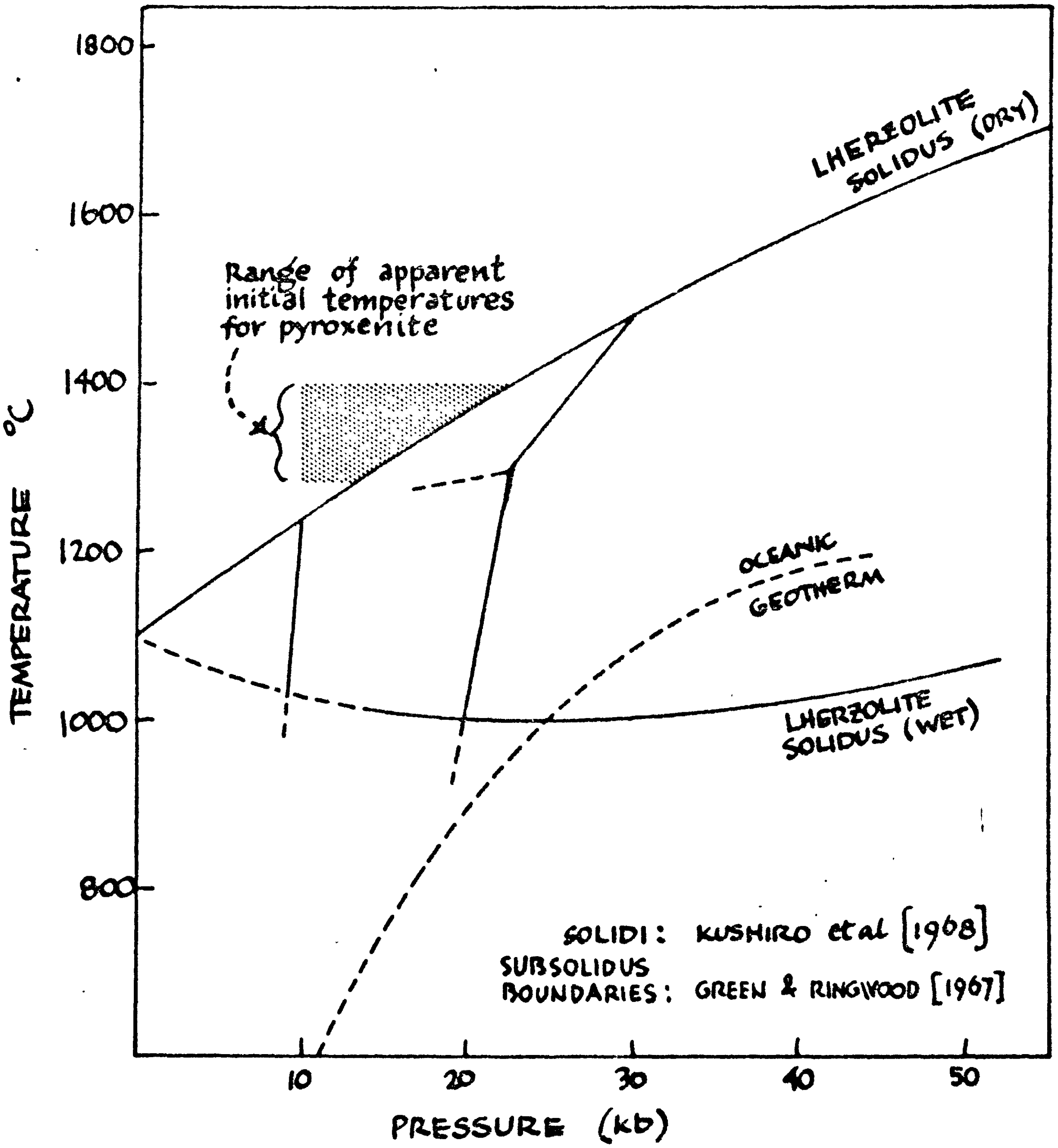


Figure II-15: Melting relations for a natural spinel lherzolite from Salt Lake Crater under anhydrous and water-saturated conditions.

the lherzolite. Subsequent cooling with the pyroxenite should then have produced exsolution in lherzolite clinopyroxenes more nearly comparable to that in pyroxenite clinopyroxenes. Striking features of lherzolite pyroxenes are their high Ca/(Ca + Mg) ratios and the small amounts of exsolution they display.

Single olivine grains along the L-P contact often show small orthopyroxene grains around them extinguishing in unison (Fig. II-13). It is difficult to imagine these orthopyroxenes randomly acquiring identical crystallographic orientation; hence, it is suggested that the orthopyroxene grains are parts of a formerly larger contiguous grain. The consumption by a lherzolite mineral (olivine) of pyroxenite orthopyroxene requires that lherzolite postdates the pyroxenite.

In light of the preceeding discussion, the hypothesis that lherzolites in contact with pyroxenite are fragments of a lherzolite mantle intruded by melts trapped at depth is considered very unlikely.

Pyroxenite Predates the Lherzolite

Eliminating the first two possible age relationships leaves the hypothesis that the lherzolite formed at a time later than the pyroxenite. Several feasible models can be devised for which the remaining hypothesis would be true. (1) the pyroxenite could have been the floor of a magma chamber onto which the lherzolite settled as a crystal cumulate. (2) the two materials could have been juxtaposed in faulting. (3) the lherzolite formed from the pyroxenite.

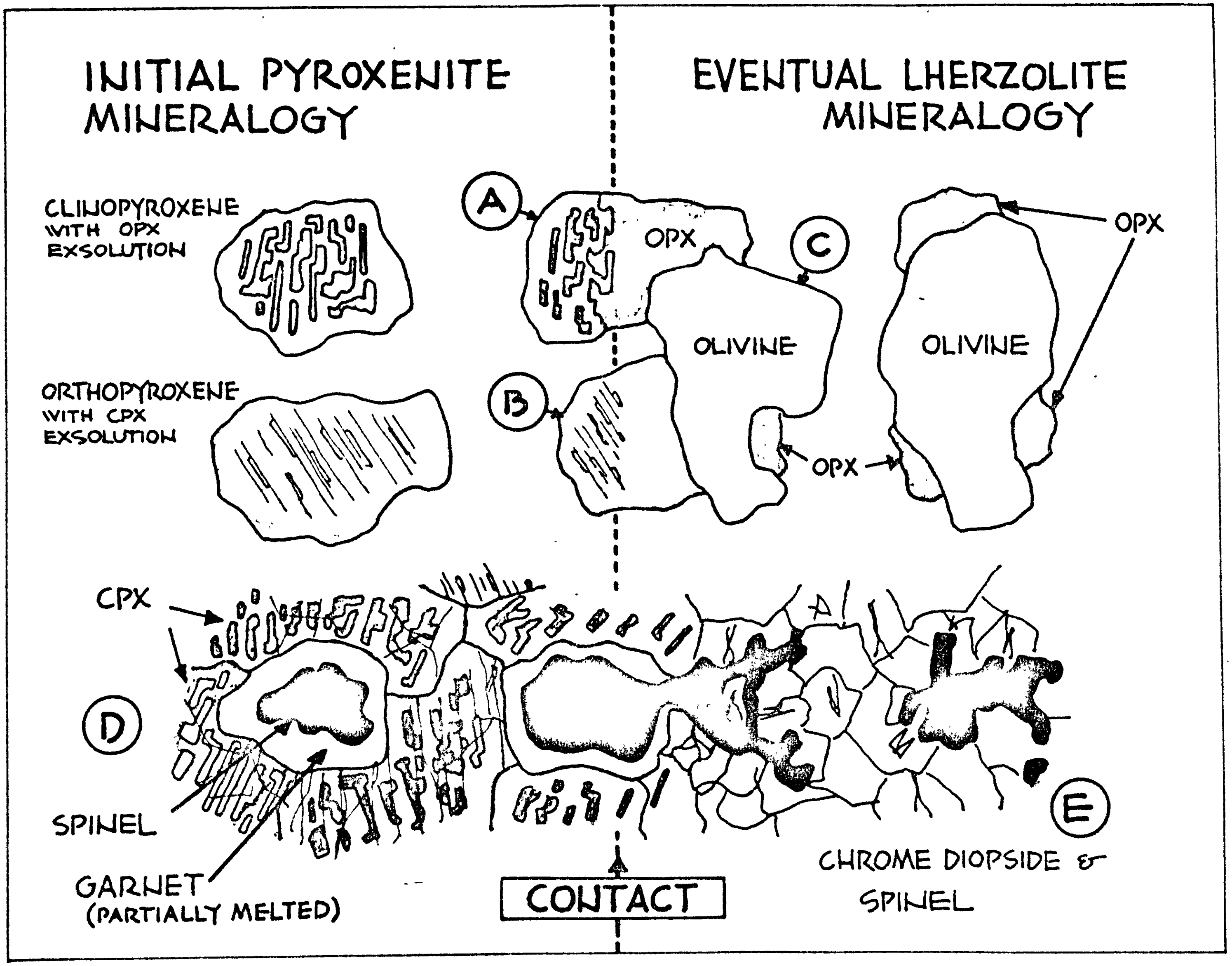
White (1966) stressed the unlikelihood that lherzolite formed as a cumulate from a basaltic magma. His reasoning is summarized in Chapter I. Further evidence against this hypothesis is the observation that in R7444-SLC48 lherzolite occurs as small irregular disconnected patches each entirely enclosed by pyroxenite. This observation also eliminates the possibility that the two materials are in fault contact, as does the complexly intergrown nature of the L-P contact.

The following discussion deals with the possibility that the lherzolite has formed at the expense of preexisting pyroxenite. A number of difficulties encountered by the Kuno hypothesis (lherzolite as xenoliths in trapped melts at depth) are quickly eliminated. Most notable is that of the high initial temperature of the pyroxenite. Lherzolite clinopyroxenes show equilibration temperatures (about 950-1000°C) comparable to those of the re-equilibrated pyroxenite clinopyroxenes (1050-1100°C). If lherzolite has formed from pyroxenite, and the process took place after the cooling and subsolidus recrystallization of the pyroxenite, the high temperature difficulties are entirely avoided.

Four textures at the L-P contact support the idea that lherzolite is a constant-volume replacement product of the pre-existing pyroxenite (See Fig. II-16 and III-5). In each of these textures, a pyroxenite mineral is found in transition to a lherzolite mineral. The result is that the four textures jointly account for the formation of the entire lherzolite mineralogy. Three of the four textures involve the appearance of pyroxenes astride the L-P contact (items 4-6 in the textural summary). Pyroxenite clinopyroxene is transformed to lherzolite orthopyroxene (A). Orthopyroxene is also transformed to lherzolite orthopyroxene (B). Olivine appears to grow at the expense of orthopyroxene of both types (C). Taken together, the three textures account for the formation of lherzolite orthopyroxene and olivine from pyroxenite clino- and orthopyroxene. The formation of the remainder of the lherzolite mineralogy can be interpreted from the spinel grains straddling the L-P contact (Figs. II-8; 16, (D) and (E)). Related garnet, spinel and clinopyroxene grains are transformed to chrome-diopside-spinel intergrowths. Garnet appears to be eradicated by melting and by re-entering solid solution in the clinopyroxene. In the process, the combined volume of

Figure II-16

Schematic appearance of textures at the lherzolite-pyroxenite contact. Encircled letters refer to descriptions in the text.


INITIAL PYROXENITE MINERALOGY
EVENTUAL LHERZOLITE MINERALOGY
CLINOPYROXENE WITH OPX EXSOLUTION
ORTHOPYROXENE WITH CPX EXSOLUTION
A
OPX
C
OPX
OLIVINE
OLIVINE
B
OPX
CPX
D
SPINEL
GARNET (PARTIALLY MELTED)
CONTACT
E
CHROME DIOPSIDE & SPINEL

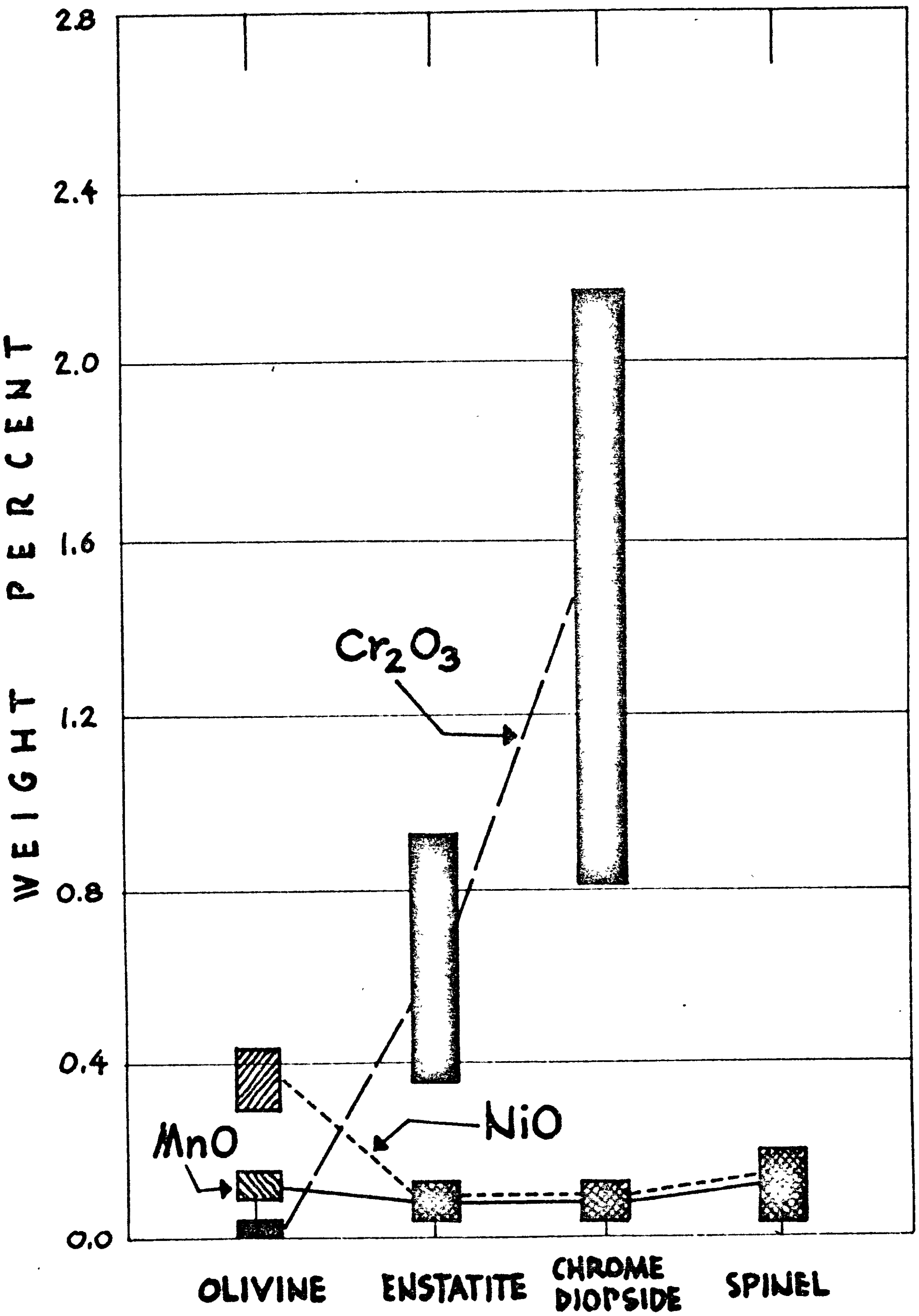


Figure II-17: Minor element variations in lherzolite silicates (after Ross, Foster and Myers, 1954, p.721).

clinopyroxene and spinel is considerably reduced. (In R7444-SLC48, spinel constitutes 8%, and clinopyroxene 38% of the pyroxenite. Chrome diopside constitutes 9% and spinel less than 1% of the associated lherzolite.) Both lherzolite minerals are left enriched to widely varying degrees in Cr_2O_3. Heterogeneity results from an inability to reach equilibrium on the scale of the whole rock. Ross, Foster and Myers (1954) analysed lherzolite chrome diopsides for their minor elements, and in contrast to very uniform Mn and Ni contents, Cr in the clinopyroxenes shows a much larger variation (Fig. II-17). As noted in the petrographic description, small orthopyroxenes are generally associated with chrome diopsides around spinel. These orthopyroxenes should also show variable Cr_2O_3, and indeed, they do. As additional support for the idea that lherzolite spinels are remnants of pyroxenite spinels, compositional trends in lherzolite spinels from Kilbourne Hole, New Mexico, and Salt Lake Crater define rather straight lines with pyroxenite spinels as one end member (Fig. II-9).

In light of the preceeding discussion, the lherzolite mineralogy can be fully accounted for by transformations of the pre-existing pyroxenite mineralogy. Three lherzolite minerals, chrome diopside, orthopyroxene and spinel are present largely as modified relict minerals from the pyroxenite. Very little to none of the lherzolite olivine has pyroxenite ancestry. Overall, the process transforms an assemblage of clinopyroxene, orthopyroxene, spinel, garnet and perhaps small amounts of olivine into the lherzolite assemblage: olivine, orthopyroxene, chrome diopside and spinel. Melting and dissolution in clinopyroxene apparently eliminate all garnet.

The presence of deformation textures in lherzolites from Salt Lake Crater, and their absence in associated pyroxenites has been used as evidence that the lherzolite pre-dates the pyroxenite in contact with it (Beeson and Jackson, in press). However, transitional Type I pyroxene grains found at the L-P contact often show deformation banding in their lherzolite portions, and not in their pyroxenite portions (Fig. II-11). In the proposed model, if lherzolite minerals form by recrystallization under conditions of stress, as might be expected in regions of magma generation and movement, the recrystallized minerals might be expected to show deformation textures.

Olivine in Salt Lake Crater lherzolites shows considerably more deformation banding than enstatite. Chrome diopside shows very little deformation. In the envisioned transformation of pyroxenite to lherzolite, olivine crystallizes for the first time at the contact. Most of the enstatite and chrome diopside in lherzolite has ancestry in the pyroxenite pyroxenes. This distinction may explain the abundance of deformation textures in the olivine, and their paucity in lherzolite pyroxenes. Regardless, the presence of deformation in lherzolite and its absence in pyroxenite does not necessitate that the lherzolite predates the pyroxenite.

Some constraints can be placed on the transformation process. None of the transformations infered from textures at the L-P contact is isochemical. An inspection of the major element compositions of co-existing pyroxenite and lherzolite (Table II-10), shows that if the process is at constant volume, the original pyroxenite has been greatly enriched in MgO, while strongly depleted in CaO, Al_2O_3, Na_2O, TiO_2 and SiO_2. Studies of R7444-SLC48 (Chapter III, Fig. 3) show the pyroxenite strongly depleted in REE by the process, leaving the residual lherzolite with a total REE content less than 1/5 that of the original material.

Both from a textural and compositional standpoint, it seems likely that a liquid has been involved. Abundant glass droplet trains and intergranular material occur in the transitional region. Qualitative electron microprobe analyses of this material show it to be very heterogeneous. This observation is consistent with the variable lherzolite mineral chemistry, if the envisioned process is responsible for their formation.

TABLE II-10. Partial major element compositions for the lherzolite and pyroxenite portions of R7444.

	R7444P	*R7444L*	*R7444P minus R7444L*
SiO_2	48.46	43.84	+ 4.52
Al_2O_3	8.96	1.26	+ 7.70
FeO	9.34	11.68	- 2.34
MgO	22.97	39.13	-16.16
CaO	7.39	1.67	+ 5.72
Cr_2O_3	0.38	0.29	+ 0.09
TiO_2	0.45	0.16	+ 0.29
Na_2O	0.90	0.16	+ 0.74
TOTAL	98.85	98.19	+ 0.66

Initially the process might appear to be a closed-system partial melting of the pyroxenite leaving the lherzolite as refractory residue. Glass formed from melted garnet is present in all pyroxenites, and might constitute the partial melt. Such glass is never more abundant modally than about 20%, and is often no more than 5%. Mass balance calculations can be attempted for a situation where parental pyroxenite (MgO = 20%) melts to a liquid (MgO = 44%). Such attemptes fail for the reason that small degrees of partial melting imply large percentages of residue. In R7444-SLC48, for example, the pyroxenite portion contains about 18% glass and would under closed-system partial melting retain 82% as residual lherzolite. One hundred grams of R7444 pyroxenite as starting material can supply 22.97 gm of MgO (Table II-10) whereas the production of 82 gm of lherzolite requires 32.09 gm of MgO. Only negative concentrations of MgO in the associated liquid can satisfy mass balance. For degrees of melting smaller than 18%, the discrepancy is larger. However, for degrees of melting around 80%, a mass balance can be written resulting in a calcium-rich "basaltic melt" and lherzolite residue. Such high degrees of melting would necessitate the consumption of both garnet and clinopyroxene in melt formation, and supporting textural evidence in L-P xenoliths is lacking. Some Salt Lake pyroxenites, without attached lherzolite, show microcrystalline material formed at clinopyroxene's expense, as well as glass formed from garnet. The degree of melting in these rocks is higher than 20%, and they could have been involved in the generation of liquid. These rocks show no incipient lherzolite formation, however.

The pyroxenite-lherzolite transformation, must then, have been open, with respect to material transport. The process envisioned is the attack of the pyroxenite by a permeating melt. The melt removes the labile components and leaves barren residual lherzolite in its place. The liquid becomes enriched in Ca, Al, Ti, Si, alkalis, REE and presumably other trace elements, and is depleted in Mg. Fe, Ni, and Mn and Cr apparently are neither enriched nor depleted (Fig. II-18). If lherzolite now occupies a volume equal to that of the original pyroxenite, a comparison of the major element compositions of the lherzolite and associated pyroxenite gives an indication of the compositional changes experienced by the liquid, but cannot directly predict the degree to which the formation of a given mass of lherzolite changes the composition of the permeating liquid. The following algebraic relation can be written for each element:

$$m_{M_1} + m_P = m_{M_2} + m_L$$

where m is the mass of the element; M_1 refers to the melt prior to incursion, and M_2 to the melt after performing the transformation of a unit quantity of pyroxenite (*P*) to lherzolite (*L*). If large relative quantities of liquid are necessary, the transformation of a given amount of pyroxenite changes the liquid composition to a minor degree. (The same permeating liquid, however, could presumably transform more pyroxenite to lherzolite as its composition was being changed.) Some reasonable limits can be placed on the minimum quantity of liquid involved per unit lherzolite formed. The formation of 100 gm lherzolite involves the deposition from the liquid of about 20 gm MgO (the difference between pyroxenite and lherzolite MgO contents), and the removal of about 8 gm each Al_2O_3 and CaO. Clearly equal proportions of liquid and pyroxenite are not reasonable estimates. The liquid's MgO content would be reduced by 20 weight percent. Almost no basalts have MgO contents greater than 20%, nor less than about 5%. It seems, therefore, that the volumes of liquids involved must be at least several times greater than the volume of lherzolite formed.

A Comparison of the Proposed Model with Data of Previous Studies

It is interesting to consider the major element data of Kuno (1969) in light of the model for the formation of lherzolite. Figure II-18 is a representation of these data. Kuno notes that the variation in lherzolite compositions is distinct from that in pyroxenite compositions. He does not mention, however, the observation that these two variations taken together form smooth and often linear trends. He interprets the garnet-bearing pyroxenites as pockets of trapped melts at depth, but feels the lherzolite compositional variation is related to their formation either as crystal cumulates, or as material depleted to different degrees in the production of basaltic magma. (Kuno admits that neither of these explanations is without difficulty. In both cases, the hypothetical magma is unlike Hawaiian basalts.) The smooth trends of lherzolite and pyroxenite compositional variation would, in his interpretation, be coincidental.

Under the hypothesis that the pyroxenites are entrapped melts in a lherzolitic upper mantle, one might imagine that the variation is lherzolite composition is due to contamination of some initial magnesian lherzolite by the entrapped melts (the pyroxenites). In this case, the compositions of lherzolites and pyroxenites taken together would fall on straight lines between the initial lherzolite composition and the pyroxenite composition. For certain elements (Mg, Fe, Si, Ti, and Mn) straight trends do exist. Straight trends *do not* result for Al_2O_3 nor particularly for CaO. Figure II-19 is an enlargement of Kuno's CaO diagram, with the addition of pyroxenite and lherzolite whole-rock compositions from Jackson and Wright (1970), and this

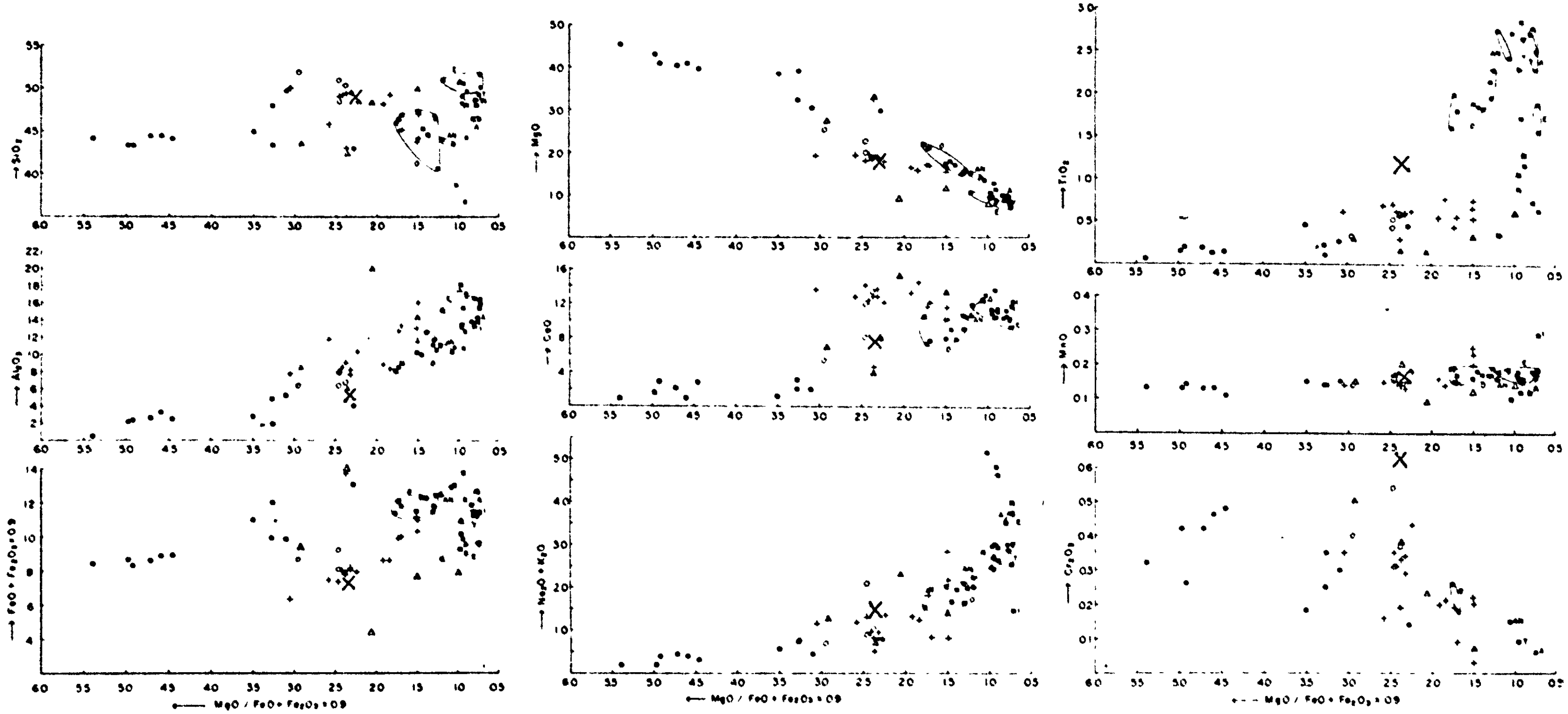
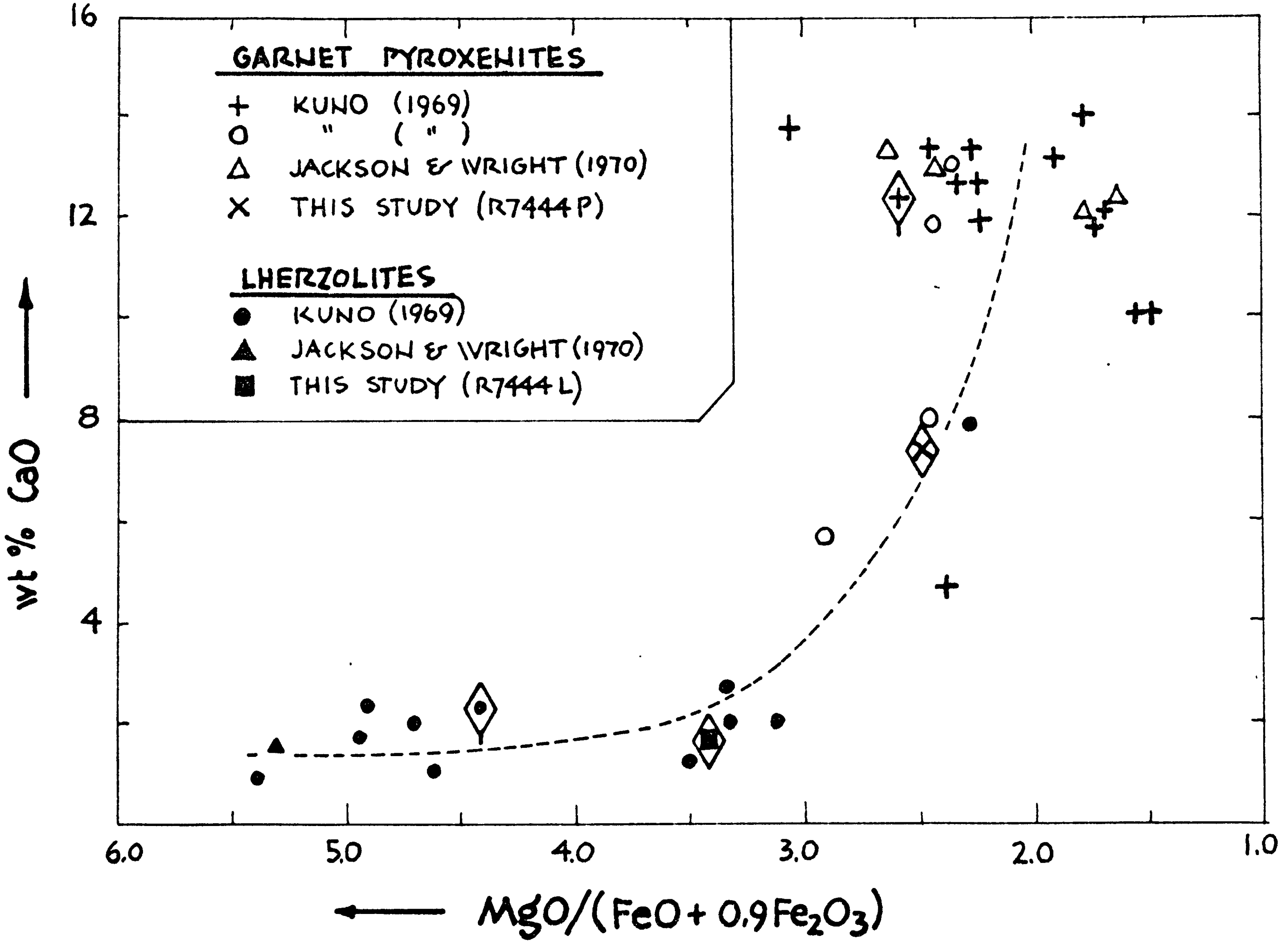


Figure II-18: Major element variations in lherzolites, pyroxenites and basalts from Hawaii (from Kuno, 1969, p. 216). ● = Salt Lake lherzolites; + & ○ = Salt Lake garnet pyroxenites; ■ = Hawaiian basalts. X is the composition of pyrolite III (Green & Ringwood, 1967), -40% olivine.

Figure II-19

Enlarged view of the CaO variation diagram of Kuno (1969, p. 216). Data are added for lherzolite and garnet pyroxenites from Jackson and Wright (1970) and for coexisting lherzolite and garnet pyroxenite (R7444) from this study. Diamonds ◊, ◊ enclose data points for lherzolite and pyroxenite from the same xenolith (R7444 and HK61082601c, respectively).


GARNET PYROXENITES
+ KUNO (1969)
o " (")
△ JACKSON & WRIGHT (1970)
× THIS STUDY (R7444P)
LHERZOLITES
● KUNO (1969)
▲ JACKSON & WRIGHT (1970)
■ THIS STUDY (R7444L)
16
12
8
4
wt % CaO
6.0
5.0
4.0
3.0
2.0
1.0
MgO/(FeO + 0.9 Fe2O3)

study (R7444P and R7444L). All lherzolites, even those with low MgO/ΣFeO ratios, are strongly depleted in CaO relative to the pyroxenites. A mixing model is not applicable to the Al_2O_3 and CaO data, and hence cannot be the explanation of any of the smooth compositional trends.

The combined variation in lherzolite and pyroxenite compositions is consistent with the model proposed in this study for the origin of lherzolite. If, as textures suggest, most of the pyroxenite clinopyroxene is transformed to lherzolite orthopyroxene in the immediate vicinity of the contact (Fig. II-16, [A]), lherzolites will be strongly depleted in clinopyroxene from their earliest formation. Hence, near-contact lherzolite will be strongly depleted in CaO relative to the pyroxenites, since clinopyroxene is the only phase containing large amounts of CaO. Much of the Al_2O_3 in the pyroxenite is contained in clinopyroxene as well. It is therefore consistent that all lherzolites, even those near the L-P contact, are also strongly depleted in Al_2O_3 relative to the pyroxenites (Fig. II-18). Thus, the proposed model - forming lherzolite as the residue when a permeating melt leaches basaltic components from the pyroxenite - can explain both the smooth linear and non-linear trends in composition displayed by Salt Lake Crater lherzolites and pyroxenites.

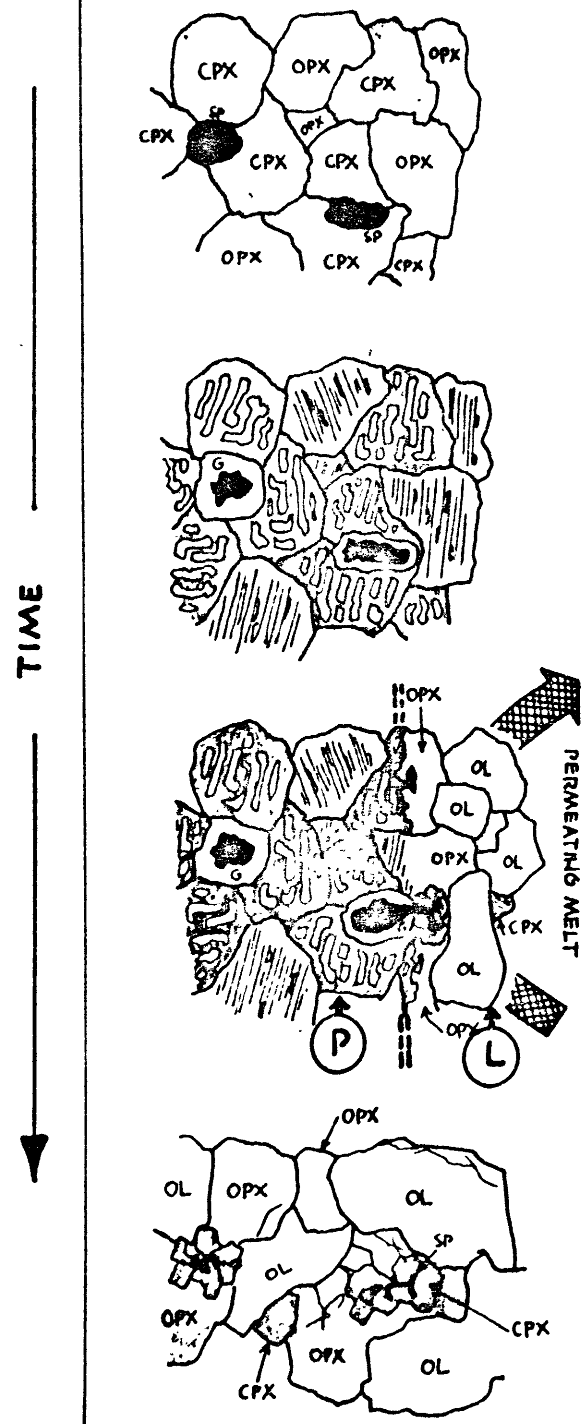
Summary

Figure II-20 is a schematic historical summary of a pyroxenite from the time of its initial crystallization to its transformation to spinel lherzolite. The pyroxenites initially crystallized as coarse-grained assemblages dominated by aluminous clinopyroxene, with smaller amounts of aluminous orthopyroxene, Al-rich spinel, and in some cases a few percent olivine. Upon subsequent cooling, large amounts of orthopyorxene have exsolved from clinopyroxene, and smaller amounts of clinopyroxene have unmixed from orthopyroxene. Pyroxenes generally display ordered patterns of euhedral spinel flakes, which are also an exsolution product on cooling. Garnet has formed both by direct exsolution from pyroxene (mainly clinopyroxene), and as coronas around primary spinel grains resulting from a reaction between spinel and the enclosing clinopyroxene. Pyroxenites generally contain between 5% and 15% glass, which is not material absorbed from the host basalt (Chapter III, Fig. 1) but is the product of a melt formed from garnet.

The transformation of pyroxenite to lherzolite apparently entails the removal of basaltic components from the pyroxenite by a permeating melt, leaving lherzolite as a barren residue. Several textures occurring at the lherzolite-pyroxenite contact indicate that the lherzolite min-

Figure II-20

Schematic history of a pyroxenite from the time of its initial crystallization as a garnet-free assemblage to its transformation to spinel-lherzolite.



1. initial assemblage: aluminous clinopyroxene, aluminous orthopyroxene, Al-rich spinel.

2. upon cooling, clinopyroxene exsolves orthopyroxene, and vice versa. garnet unmixes from pyroxene, and forms coronas around spinels.

3. a melt attacks the pyroxenite, leaching basaltic components, leaving barren residual lherzolite. Lherzolite opx,cpx and spinel are remnants of the same pyroxenite minerals. Olivine forms from pre-existing orthopyroxene.

4. eventual lherzolite assemblage: olivine, orthopyroxene, chrome diopside, and spinel. The latter two occupy much less volume than their pyroxenite counterparts.

eralogy originates through transformations of pyroxenite minerals. Most of the pyroxenite clinopyroxene is transformed to lherzolite orthopyroxene (Fig. II-16, [A]). As a result, the lherzolite is left with a strong depletion in CaO relative to the pyroxenite within a short distance from the contact. Exsolution-bearing pyroxenite orthopyroxene is also transformed to exsolution-free lherzolite orthopyroxene (B). Olivine is found as embayments in these near-contact lherzolite orthopyroxenes, and apparently originates when the permeating liquid leaches silica from the orthopyroxene (C). Associated grains of clinopyroxene, garnet and spinel in the pyroxenite are transformed to chrome diopside-spinel intergrowths in the lherzolite (D) and (E). The process involves the reduction of the volume of both clinopyroxene and spinel, and the eradication of the garnet. Apparently, the garnet disappears by undergoing melting and also by re-entering solid solution in the clinopyroxene. During their volume reduction, both clinopyroxene and spinel become more chrome-rich (Tables II-7, II-10). Enrichment in Cr_2O_3 in these two minerals varies widely within small regions of near-contact lherzolite. This heterogeneity in spinel and chrome diopside Cr_2O_3 contents is apparently maintained, as wholly lherzolite inclusions in several studies show wide ranges in the composition of these minerals with respect to Cr_2O_3/

Al_2O_3 (Ross *et al.*, 1954; Carter, 1969; this study). The overall transformation results in the four phase assemblage: olivine, orthopyroxene, chrome diopside and spinel, the latter three of which have ancestry in pyroxenite minerals.

In summary, the textural and major element data are inconsistent with the idea that the Salt Lake Crater pyroxenites are liquids which have intruded pre-existing lherzolite and solidified at depth to their present assemblages. Rather, the data favor the idea that lherzolite is the secondary material, having formed as a barren residue, when a permeating melt removed basaltic components from pre-existing pyroxenite.

The following chapter discusses the distributions of rare earth elements in whole-rock and separated mineral samples from near-contact lherzolite and garnet pyroxenite, to further elucidate the contact relationship.

CHAPTER III

RARE EARTH DISTRIBUTIONS IN LHERZOLITE AND GARNET PYROXENITE AND THE CONSTITUTION OF THE UPPER MANTLE

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ABSTRACT

Spinel lherzolite occurs at Salt Lake Crater, Hawaii intimately associated in single xenoliths with garnet pyroxenite. Investigations of the rare earth element (REE) distributions in these two-assemblage rocks show (1) pyroxenite mineral REE patterns sum very nearly to their whole-rock pattern, (2) chondrite-normalized REE patterns for pyroxenites are gently curved and convex-upward; they have maxima in the range Pr to Gd, with total REE contents 4 to 10 times average chondrites, (3) lherzolites are depleted in REE relative to associated pyroxenites, but show a greater light REE enrichment, (4) lherzolite chrome diopsides and associated pyroxenite clinopyroxenes have very similar REE distributions.

The data suggest the following interpretations: (1) Xenoliths are not contaminated by the host-rock. (2) Solid/liquid distribution coefficients applied to pyroxenite clinopyroxenes give liquids with REE patterns unlike any Hawaiian basalt. If the pyroxenites crystallized from liquids no longer present, their formation is probably unrelated to the Hawaiian vulcanism. The pyroxenites may be trapped liquids occurring as components in tholeiites, or may be parental to tholeiites. (3) Similarity in REE pattern between associated lherzolite and pyroxenite clinopyroxenes is inconsistent with the usual interpretation of

these bimodal xenoliths - that the pyroxenite is a liquid which intruded upper mantle lherzolite and crystallized at depth. REE data, supported by textural observations, suggest that lherzolite has formed as residue when basaltic components were removed from original pyroxenite by a permeating melt. Relative to lherzolite, pyroxenite is the more primitive upper mantle material.

INTRODUCTION

Inclusions of spinel lherzolite (olivine, enstatite, chrome diopside and spinel) have been found in a large number of basaltic localities, and they are generally the most abundant ultramafic rock in a given xenolith suite (Forbes and Kuno, 1967). The inclusions show remarkably uniform mineral chemistry (Ross, Foster and Myers, 1954; Wilshire and Binns, 1967; White, 1966) despite being incorporated in host rocks of varied composition. The ubiquity and uniformity of lherzolites has been used to support the peridotite upper mantle model.

Jackson (1966, 1968) has shown that the xenolith suite at Salt Lake Crater, Hawaii, has a bimodal distribution. One group of inclusions is typical spinel lherzolite. Salt Lake Crater lherzolites were analyzed in the studies of Ross *et al.*, and White. Occurring in nearly equal volume are xenoliths of garnet pyroxenite composed dominantly of garnet and clinopyroxene, with smaller amounts of orthopyroxene, olivine, spinel and in some cases, phlogopite. As noted by Green (1966), and by Beeson and Jackson (in press), subsolidus recrystallization textures indicate a cooling history from about 1300-1400°C to about 1100°C at pressures on the order of 20 kb. Attached lherzolites, however, show evidence of having formed at 1000-1100°C, and show no record of a high temperature past.

Several Salt Lake Crater xenoliths containing regions of lherzolite in contact with garnet pyroxenite have been studied. Inclusions of this type were mentioned by Jackson. Kuno (1969) reports major element compositions for the associated lherzolite and pyroxenite of such a xenolith. He interprets the lherzolite to be a fragment of the pervasive upper mantle peridotite zone incorporated in a basaltic melt which crystallized at depth to the pyroxenite assemblage. Another interpretation has been proposed. From a detailed petrographic and electron microprobe study of the lherzolite-pyroxenite contact in several xenoliths, Reid (1970) has concluded that the lherzolite has formed from pre-existing pyroxenite, as the residue left after the extraction of basaltic components from the pyroxenite by a permeating melt. The two studies lead to different interpretations as to which of the two rock types is the more primitive upper mantle material.

Because of their coherent and systematic geochemistry, the REE are particularly well suited to the study of a contact relationship of this sort. Considerable effort has been made in recent years to determine the distribution of the REE between mafic rock-forming minerals and their coexisting liquids (Schnetzler and Philpotts, 1968, 1970; Onuma *et al.*, 1969; Masuda and Kushiro, 1969). Clinopyroxene/liquid distribution coefficients determined by these several groups of workers differ by as much as a factor of three in absolute magnitude. All results, however, indicate the tendency of clinopyroxene to accomodate the heavy REE equally well, and to increasingly discriminate against the light REE. This consistency in relative distribution coefficients justifies their use in the study of systems like Salt Lake Crater xenoliths where clinopyroxene is the dominant REE-bearing phase.

EXPERIMENTAL

Analyses were made using the neutron activation procedures of Haskin *et al.* (1968). Xenoliths were trimmed of host basalt skins, and coarse crushed in an acid-cleaned steel percussion mortar. Fine crushing was performed in a boron carbide mortar. Powders were leached in warm acetic acid, or dilute HCl under agitation in an ultrasonic cleaner to remove deuteric $CaCO_3$. Separates of clinopyroxene from lherzolite and pyroxenite xenoliths were readily made using the Frantz isodynamic separator alone. The separation of garnet required the use of Clerici solution. Mineral separates were cleaned in warm HCl with the use of the ultrasonic vibrator. The purity of minerals to be analyzed was checked under the binocular microscope, and contaminating grains were picked out by hand. Powders weighing between 0.25 and 0.85 grams were irradiated, the amount being determined by the anticipated REE levels and the sample supply.

REE distributions are displayed as comparison diagrams (Masuda, 1962; Coryell *et al.*, 1963) normalized to a composite of nine chondritic meteorites analyzed by Haskin *et al.* (1968). Duplicate analyses of pyroxenite R7444 indicate an average precision of about ± 7%. Accuracy is believed equal to the precision. The results are shown in Table 1 and Figures 1 through 6.

RESULTS and DISCUSSION

The Question of Contamination from the Host Basalt

Recent studies of trace element distributions in mantle-derived xenoliths have demonstrated the need for caution in applying results to the discussion of processes taking place in the upper mantle. Erlank (1969) has shown that the garnets in eclogites from Roberts-Victor Mine, South Africa, contain alteration rims and crack-filling material rich in alkali elements. Often impossible to remove, such contamination drastically alters the indigenous trace element patterns of the xenoliths. Since the xenoliths at Salt Lake Crater have been brought to the surface in a basalt with unusually high trace element contents (Schilling and Winchester, 1969), it is necessary to determine whether the xenoliths' trace element contents are dominated by material absorbed from the host basalt. Relevant data have been obtained from garnet pyroxenite, R7401.

Clinopyroxene and garnet together comprise 83% of the rock and contain virtually all of the REE. Olivine, orthopyroxene and spinel, each with much lower REE contents make up the remainder. When the REE patterns of the carefully cleaned garnet and clinopyroxene are added together in their modal proportions, their sum very closely matches the whole-rock pattern (Figure 1).

This result indicates that the alteration free clinopyroxene and garnet are the important host phases for REE in the pyroxenite. The relative enrichment of the garnet in heavy REE is expected from crystal chemical limitations. This is evidence against the importance of minute inclusions rich in REE. Additional evidence against contamination is that the garnet La abundance is 0.019 ppm compared to a typical Honolulu Series nephelinite La abundance of 68 ppm (Schilling and Winchester, 1969). These data justify the use of REE data in Salt Lake Crater pyroxenites as indicators of processes taking place in the upper mantle.

There is some indication that the pyroxenites have been closed systems since before the formation of garnet. None of the garnet-bearing pyroxenites examined contains what is unquestionably primary garnet (R7401 and R7444 contain no primary garnet; at least 90% of the garnet in R7629 is definitely secondary). All garnet has formed during sub-

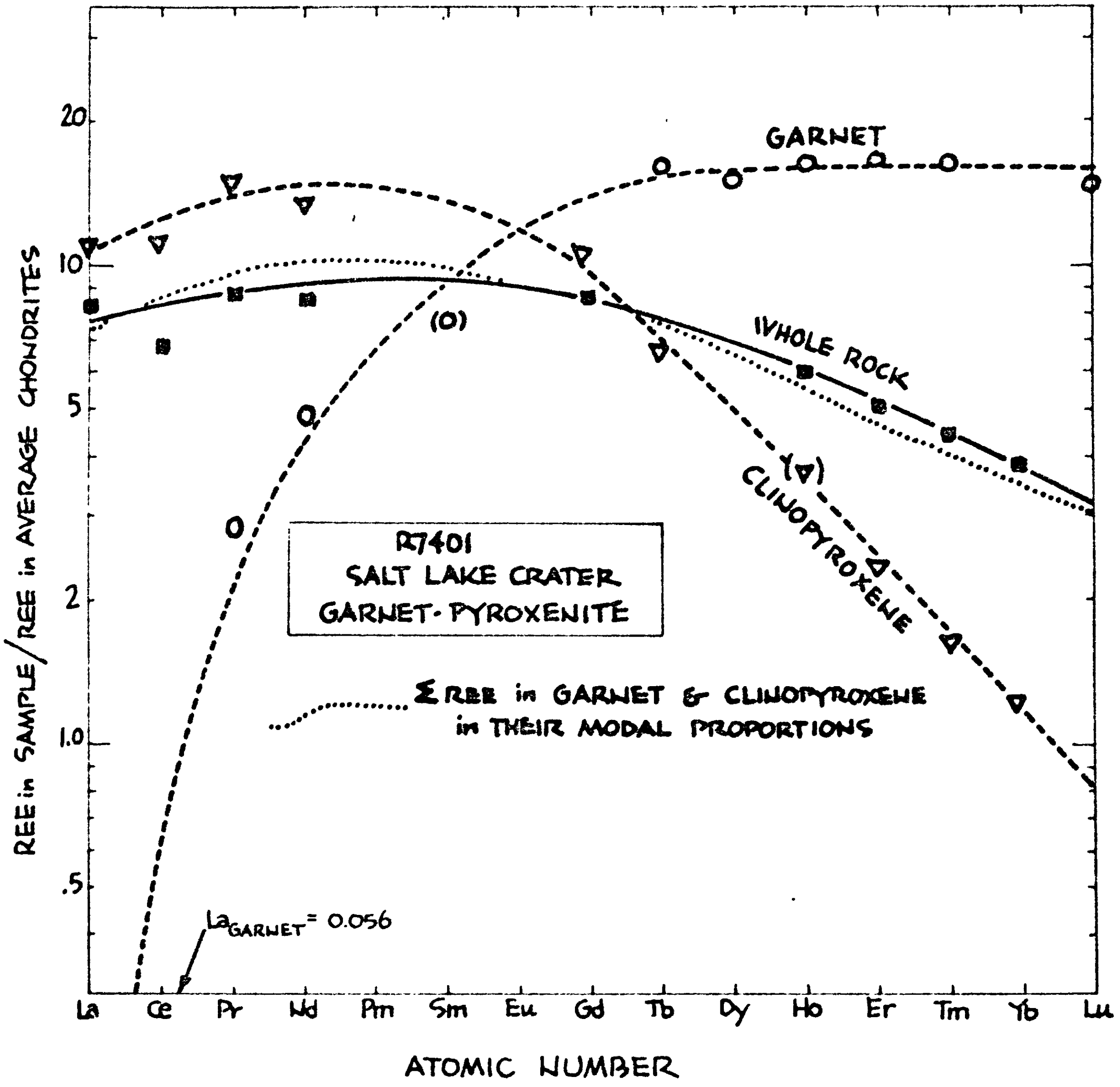


Figure 1: REE distributions in whole rock and separated garnet and clinopyroxene samples of R7401 garnet pyroxenite. Dotted line is the sum of REE patterns of the clinopyorxene and garnet added in their modal proportions (65% clinopyroxene, 18% garnet).

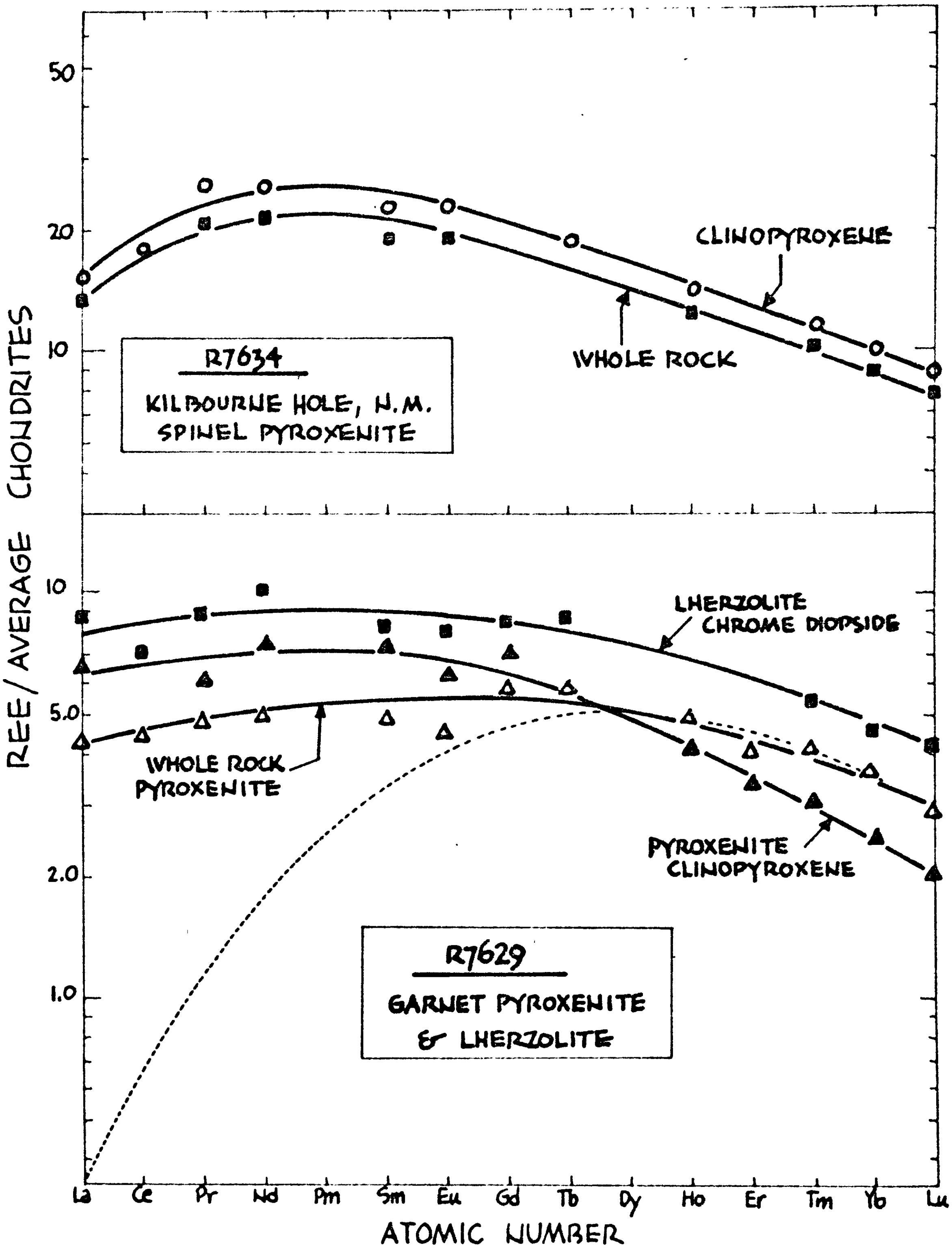
solidus cooling either by direct exsolution from clinopyroxene, or as the product of a reaction between spinel and clinopyroxene. On the other hand, no Salt Lake pyroxenite has been found which has not undergone some degree of recrystallization to form garnet. Spinel pyroxenite inclusions found at Kilbourne Hole, N.M., however, are texturally like the Salt Lake Crater pyroxenites, except that no garnet has formed. REE data for the Kilbourne Hole pyroxenite (R7634) and its separated clinopyroxene indicate that the whole-rock REE pattern is strictly dominated by the clinopyroxene (Figure 2). Clinopyroxene constitutes about 80% of the rock, with spinel and olivine making up the remainder. The shapes of the REE patterns for R7634 and R7401 are quite similar (Figures 1 and 2). This similarity indicates that garnet formation in R7401 has caused a simple closed-system redistribution of REE from the original clinopyroxene to the eventual garnet-clinopyroxene pair. The crystallization of garnet, which preferentially accomodates the heavy REE has left the clinopyroxene with a depletion of heavy REE relative to the original clinopyroxene pattern.

REE Distributions in the Pyroxenites

The pyroxenites have chondrite-normalized REE distributions which are gently curved convex-upward. Enrichment of the light REE is uniformly low (La/Yb ranges from

Figure 2

REE distributions in whole rock and separated clinopyroxene samples from R7634 (Kilbourne Hole, N.M.) spinel pyroxenite; R7629 garnet pyroxenite; R7629 lherzolite. Lightly dashed curve is the REE distribution for a clinopyroxene in equilibrium with a liquid having the REE pattern of R7629 whole rock garnet pyroxenite.


50
20
10
CLINOPYROXENE
WHOLE ROCK
R7634
KILBOURNE HOLE, N.M.
SPINEL PYROXENITE
10
5.0
2.0
1.0
LHERZOLITE
CHROME DIOPSIDE
WHOLE ROCK
PYROXENITE
PYROXENITE
CLINOPYROXENE
R7629
GARNET PYROXENITE
& LHERZOLITE
REE / AVERAGE CHONDRITES
La
Ce
Pr
Nd
Pm
Sm
Eu
Gd
Tb
Dy
Ho
Er
Tm
Yb
Lu
ATOMIC NUMBER

1.2 to 2.0). Except for R7444 (Figure 3), all show maxima in the range Pr to Gd. (The REE pattern for R7444 pyroxenite is straight, and nearly flat. Its separated clinopyroxene, however, shows a maximum at Nd [Figure 3]. The simple combination of this clinopyroxene pattern with a typical garnet pattern [R7401, Figure 1] yields a whole-rock pattern which is curved, not straight. A third phase, not present in appreciable amounts in R7401, must be present and must contribute the light REE necessary to create the straight pattern. The nature and significance of this phase is considered in a later section).

An indication of the conditions of crystallization can be obtained by applying solid/liquid distribution coefficients to the REE patterns of the *original* pyroxenite clinopyroxenes. The present clinopyroxenes have REE patterns altered by the subsolidus formation of heavy REE-enriched garnet, and hence will not give useful information about the initial pre-garnet crystallization of the pyroxenites. Howover, clinopyroxene alone is the dominant REE bearing phase in the garnet-free assemblage (R7634, Figure 2). Hence the present whole-rock and the initial clinopyroxene have REE distributions with the same shape, though differing REE concentrations. The approximate initial (pre-garnet) clinopyroxene REE abundances can be obtained by multiplying the present whole-rock pattern by the inverse of the modal proportion of clinopyroxene in the rock. Since major element

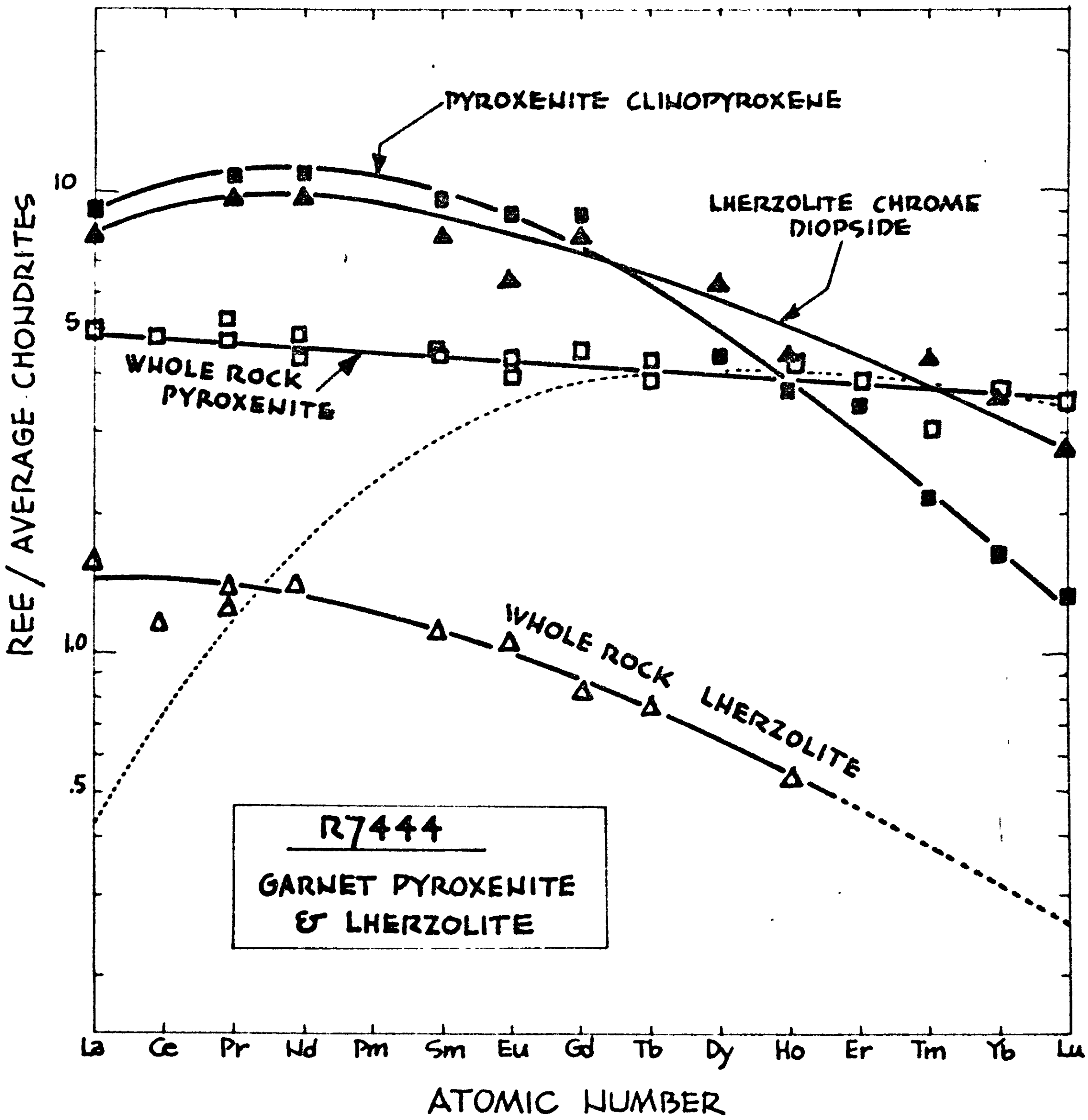


Figure 3: REE distributions in whole rock and separated clinopyroxene samples from R7444 garnet pyroxenite and R7444 lherzolite. Lightly dashed curve is the REE distribution for a clinopyroxene in equilibrium with a liquid having the REE pattern of R7444 whole rock garnet pyroxenite.

compositions of the clinopyroxene used in the study of Onuma *et al.* (1969) most closely matche pyroxenite clinopyroxenes, their distribution coefficients have been used. (Other sets of coefficients lead to the same conclusions). When these coefficients are applied to the REE patterns of pre-garnet clinopyroxenes in R7401 and R7629, hypothetical liquids are obtained which do not resemble any Hawaiian basalt (Figure 4). All Hawaiian basalts analyzed by Schilling and Winchester have convex-upward patterns. The hypothetical liquids are clearly concave-upward.

This dissimilarity can be interpreted in a number of ways. (1) If the pyroxenites crystallized from liquids no longer present, their formation is probably unrelated to the vulcanism creating the Hawaiian island chain. The conclusion of Green (1966) that the Salt Lake Crater pyroxenites are early crystallizations from alkali basalt magma is not supported by this data. (2) The pyroxenites may be the complete crystallizations of entrapped liquids, whose REE patterns were retained by the pyroxenites. Pyroxenite REE patterns are not equivalent to any basalts, though some tholeiites have patterns which are more REE-enriched but have similar shapes (Figure 4). Though similar to tholeiites in major element composition, Salt Lake Crater pyroxenites have been shown by Jackson and Wright (1970) to have considerably higher CaO contents and lower alkali element and TiO_2 levels

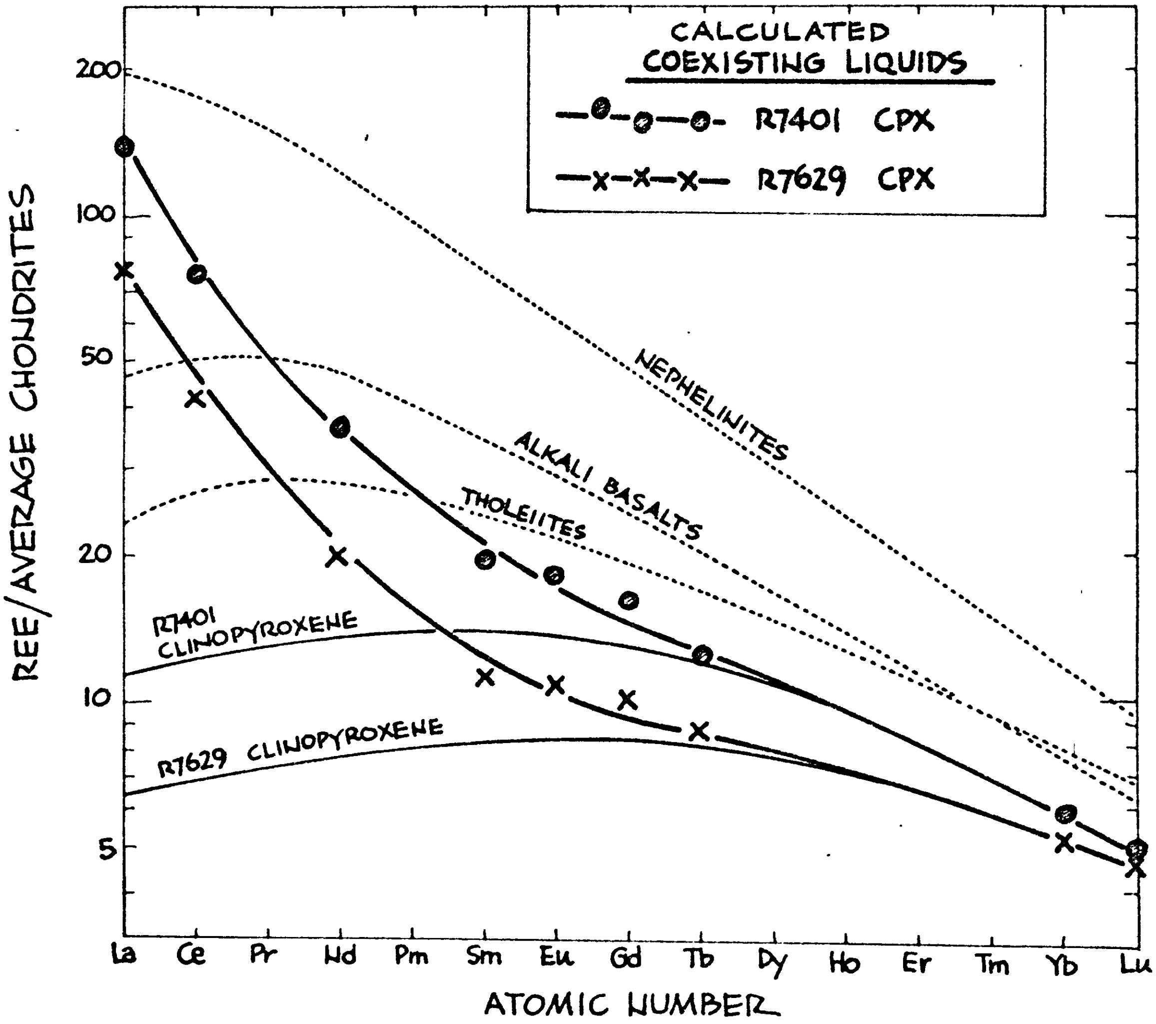


Figure 4: REE Distributions in liquids coexisting with "pre-garnet" clinopyroxenes from garnet pyroxenites R7401 and R7629. Distribution coefficients used are from Onuma *et al.*, 1969. Lightly dashed curves are Hawaiian lavas (Schilling and Winchester, 1969).

than Hawaiian tholeiites. Hence, if pyroxenites are entrapped liquids, no equivalent liquid is found at the surface. They may be crystallizations of liquids occurring only as components of tholeiitic basalts. (3) Pyroxenites may have similar major element and REE chemistry by being parental to tholeiites. Partial melting of a rock composed of garnet, clinopyroxene, orthopyroxene, olivine and spinel would result in initial liquids dominated by clinopyroxene and garnet (Ito and Kennedy, 1967). Pyroxenite REE patterns would be inherited by the liquids if all garnet and clinopyroxene underwent melting. (Ito and Kennedy have shown that garnet and clinopyroxene are both entirely melted within 50°C of the solidus for a natural garnet peridotite composition at 20 kb.) Raising the REE level in the liquid from pyroxenite concentrations to those found in Hawaiian basalts would require either fractional crystallization of phases with low REE abundances, or perhaps an enrichment process like the one proposed in this paper for the origin of lherzolite.

Contribution of REE data to the understanding of the Lherzolite-Pyroxenite contact relationship

Textural studies of these two-assemblage xenoliths suggest two alternative explanations for their paragenesis. As proposed by Kuno (1969), one possibility is that the pyroxenite is a basaltic melt which has intruded the pervasive upper mantle lherzolite zone and crystallized at depth. The other interpretation is that the lherzolite is the younger material, having formed as residue when a permeating melt removed labile components from the pre-existing pyroxenite (Reid, 1970). The possibility of the two materials having formed simultaneously is small. Initial crystallization temperatures based on clinopyroxene exsolution data (Beeson and Jackson, in press; Reid, 1970) differ by 300-400°C for the two materials.

1. Pyroxenite has intruded lherzolite

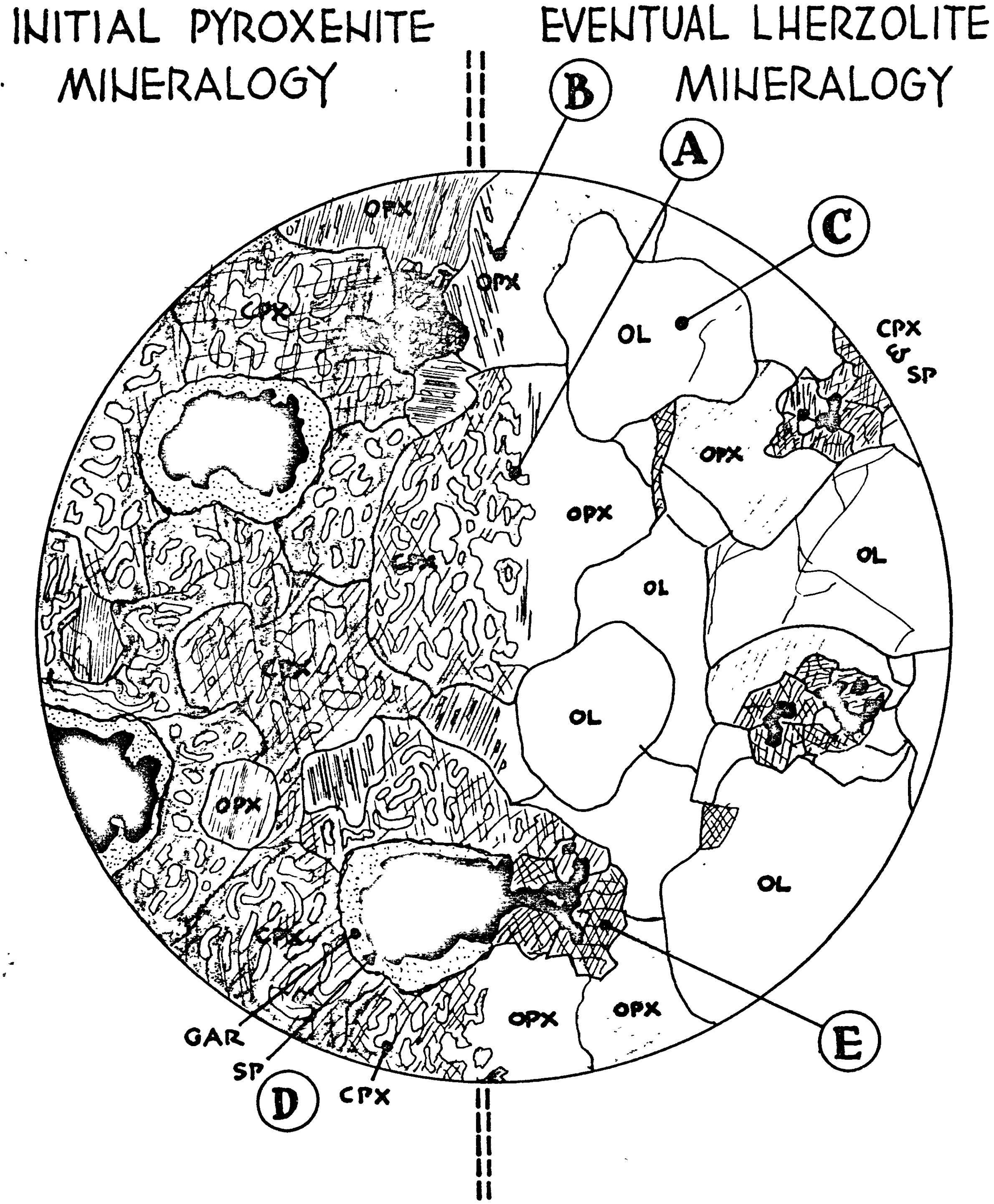
If the pyroxenite is a basaltic melt which intruded deep-seated lherzolite, and if one assumes that the melt and the intruded lherzolite came to equilibrium within several centimeters of the contact, the REE distribution in the lherzolite clinopyroxene can be predicted. (The assumption of equilibrium is justified on the basis of the high initial temperatures of the pyroxenites - 1300-1400°C - and the knowledge that the analyzed chrome diopside was separated from material within 2 cm. of the contact). A clinopyroxene equilibrating with a basaltic liquid with the REE distribution of either R7444 or R7629 whole-rock will have a REE distribution increasingly depleted toward the light REE relative to the pyroxenite (the supposed liquid). The dotted lines in Figures 2 and 3 are these clinopyroxene patterns. Actually analyzed lherzolite clinopyroxenes from R7444 and R7629 have distributions quite different from these hypothetical clinopyroxene patterns. Rather than depleted, they are enriched in the light REE, and contain higher absolute REE contents of the light REE than the associated pyroxenite. Despite the descrepancy in the absolute value of the clinopyroxene/liquid distribution coefficients (Schnetzler and Philpotts, 1967, 1970; Onuma *et al.*, 1969; Masuda and Kushiro, 1969) none show any tendency for clinopyroxene to accomodate more light REE than its associated liquid. The lherzolite clinopyroxene REE distributions, hence, cannot be reconciled with the idea that the pyroxenite is a melt which has intruded the lherzolite and solidified at depth.

2. Lherzolite formed from pyroxenite

In bimodal inclusions, lherzolite is invariably fresher. Pyroxenite pyroxenes are considerably more turbid than lherzolite pyroxenes stemming from included trains of glass beadlets, and exsolution patterns of tiny euhedral spinel flakes in the former. Glass apparently formed from melted garnet constitutes up to 15% of some pyroxenites. Lherzolite mineral grains have very clean contacts, and show no sign of melting. (Experimental studies of lherzolite melting at high pressures and high water pressures

Figure 5

Schematic appearance of the lherzolite-pyroxenite contact. Letters (A) through (E) refer to statements in the text.


INITIAL PYROXENITE MINERALOGY
EVENTUAL LHERZOLITE MINERALOGY
B
A
C
OPX
OPX
CPX
OL
CPX & SP
OPX
OPX
CPX
OL
OL
CPX
OL
OPX
OL
CPX
OPX
OPX
GAR
SP
D
CPX
E

[Kushiro *et al.*, 1968] indicate that the lherzolite should be partially melted if intruded by a pyroxenite melt at 1300-1400°C.) There is also no evidence of invasion of lherzolite by pyroxenite. Close petrographic inspection of the contact region in a number of inclusions has revealed transitional grains of several types which suggest that the lherzolite mineralogy originates from the pyroxenite (Figure 5). Lherzolite orthopyroxene originates by transformation of both pyroxenite pyroxenes (A) and (B). Lherzolite olivine grows at the expense of orthopyroxene (C). Clinopyroxene-spinel-garnet intergrwoths in the pyroxenite are transitional to chrome diopside-spinel intergrowths in the lherzolite (D) and (E). The chrome diopside-spinel pair are enriched to widely varying degrees in Cr_2O_3 within a given xenolith. The enrichment results from the reduction in volume of chrome diopside and spinel relative to the corresponding pyroxenite minerals, and perhaps by the removal of Cr_2O_3 from the permeating liquid. Olivine and orthopyroxene near the contact show less range in composition, but are variable in MgO/FeO. Pyroxenite minerals by contrast are homogeneous.

If the lherzolite has originated by reaction of a melt with the pyroxenite, as proposed, the REE contents of near-contact lherzolite chrome diopsides may reflect this lack of equilibrium, and hence, the chrome diopsides may still retain some of the REE characteristics of the pyroxenite clinopyroxenes from which they formed. Lherzolite chrome diopside and associated pyroxenite clinopyroxene have been analyzed from two Salt Lake Crater garnet pyroxenites. In each case (R7444, Figure 3; and R7629, Figure 2), the associated clinopyroxenes have similar REE distributions. (The agreement is poorer for R7629, Figure 2, than for R7444, Figure 3, perhaps because of a higher content of exsolved orthopyroxene in R7629 pyroxenite clinopyroxene [which acts as a diluent], than in the corresponding separate from R7444.) In each rock, the associated lherzolite and pyroxenite clinopyroxenes show less similarity in their heavy REE distributions, than in the light REE. Lherzolite clinopyroxene in each case has greater relative concentrations of the heavy REE than its pyroxenite counterpart. This difference may be related to the disappearance of garnet at the contact. If garnet is eradicated in part by re-entering solid solution in the clinopyroxene, as textural observations indicate, the high heavy REE concentrations in the garnet would account for the enrichment in heavy REE in the lherzolite chrome diopside relative to its associated pyroxenite clinopyroxene.

REE data from pyroxenite R7444 suggest the possible presence of the liquid responsible for the transformation of pyroxenite to lherzolite. Although the modal sum of the REE patterns for garnet and clinopyroxene add precisely to the whole-rock pattern in the case of R7401 (Figure 1), the whole-rock pattern for R7444 cannot be accounted for entirely by these two minerals. Garnet with very low concentrations of the light REE, and the curved REE distribution of R7444 pyroxenite clinopyroxene (Figure 3) cannot yield the observed straight-line whole-rock distribution. A small amount of a phase rich in the light REE must also be present. Contamination from the host nephelinite basalt is a possibility (Figure 4 shows a nephelinite REE distribution). However, calculations show that the whole-rock pattern in R7444 lherzolite is more nearly matched by its separated mineral REE contents than is the whole-rock pattern for R7444 pyroxenite. That is, more of the light REE enriched phase is present in the pyroxenite than in the lherzolite. If the source of this mineral were the host basalt, both materials should have comparable contamination and the lherzolite REE pattern should be affected to the greater degree, since it contains about 5 times lower REE concentrations than the pyroxenite. Garnet in R7444 pyroxenite has undergone nearly complete melting; R7401 garnet shows relatively little glass. In regions of advanced melting of garnet in R7444, electron microprobe analyses have revealed the presence of excess alkalis relative to the garnet from which the glass has formed. The source of these alkalis and the excess light REE may be the liquid performing the leaching process which leaves residual lherzolite in the place of original garnet pyroxenite.

An earlier REE study of lherzolites may be consistent with this model. Nagasawa *et al.* (1969) have reported REE distributions for whole-rock and separated chrome diopside samples from several Salt Lake Crater lherzolites. When distribution coefficients are applied to the REE patterns of chrome diopsides, the associated liquids have chondrite-normalized distributions which are concave-upward, like the hypothetical liquids associated with pyroxenite clinopyroxenes of this study. In fact, a lherzolite chrome diopside analyzed by Nagasawa *et al.* (Sample no. 2) has a REE pattern nearly identical to the R7401 *pyroxenite* clinopyroxene reported here. In terms of the model proposed in this paper, this chrome diopside came from near-contact lherzolitic material, and retained the REE characteristics of the pyroxenite clinopyroxene from which it formed.

The REE distribution has been determined for a chrome diopside separated from a Salt Lake Crater lherzolite having no attached pyroxenite. Its pattern is unlike any pyroxenite clinopyroxene (Figure 6). Application of distribution coefficients to this pattern indicates that the liquid last in equilibrium with this pyroxene had a distribution similar to the more alkali-rich, silica-undersaturated Hawaiian basalts. In the proposed model, this lherzolite may have been brought from a region in the mantle further removed from the lherzolite-pyroxenite contact, and may represent a more nearly equilibrated assemblage. The liquid may be all or a component of the per-

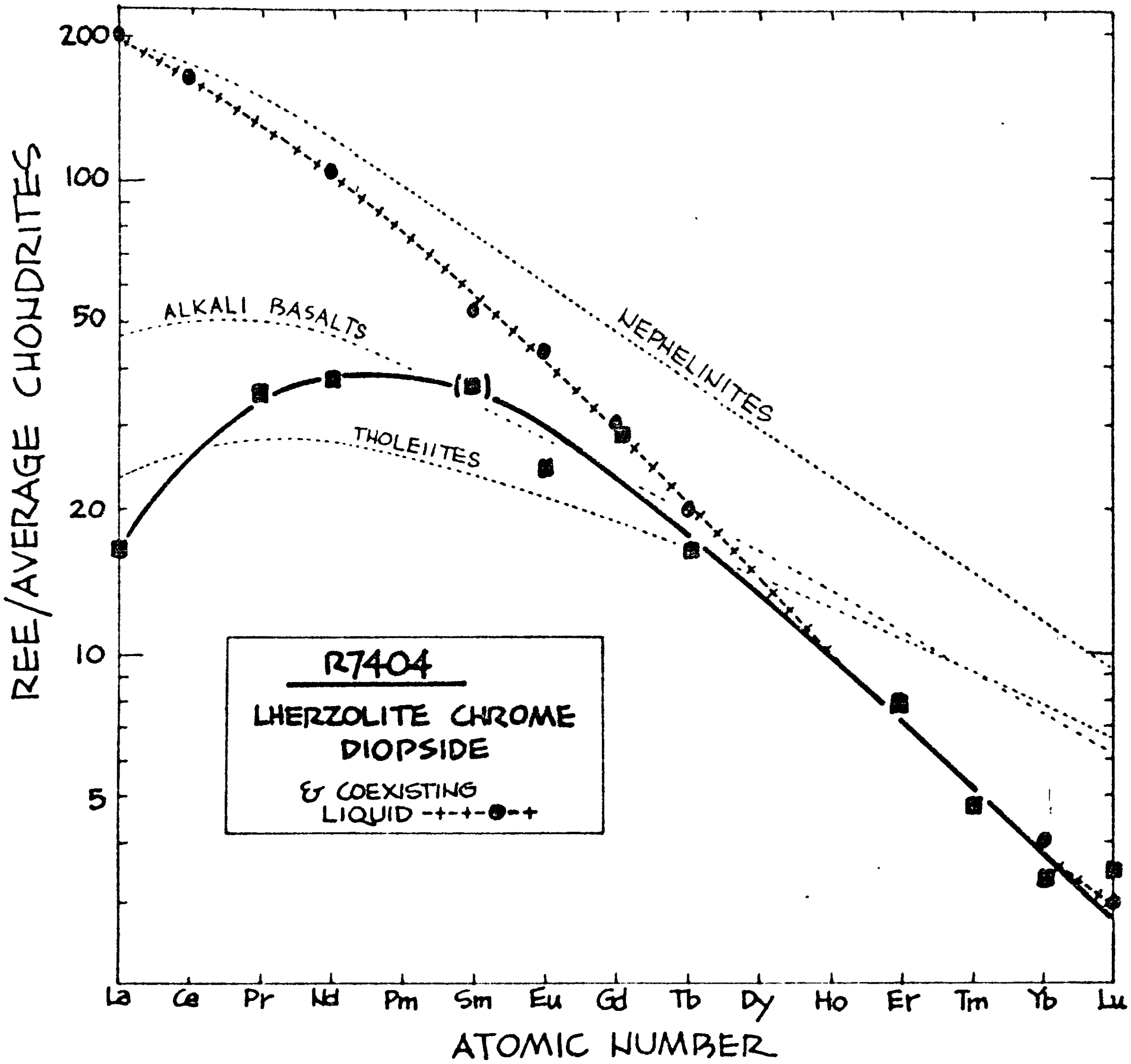


Figure 6: REE distributions in chrome diopside from lherzolite showing no attached pyroxenite. The coexisting liquid pattern was determined using distribution coefficients of Onuma *et al.*, 1969. Lightly dashed curves are Hawaiian lavas (Schilling and Winchester, 1969).

meating melt involved in the hypothetical leaching process.

CONCLUSIONS

This investigation of Salt Lake Crater xenoliths having lherzolite and garnet pyroxenite in mutual contact, suggests it is unlikely that the lherzolite represents primitive mantle material intruded by a liquid which has crystallized to garnet pyroxenite at depth (as suggested by Kuno, 1969). Textural, REE, and major element data indicate that lherzolite is the younger material, having formed from the pre-existing garnet pyroxenite. Depletion of REE and basaltic major element components (Ca, Al, Na, K, Ti) in the lherzolite relative to the pyroxenite (Table 2) suggests that the transformation involves the removal of these materials from the pyroxenite, leaving lherzolite as a barren Mg-rich residue. If true, our view of a predominantly peridotitic upper mantle may be inaccurate. If lherzolite forms from pyroxenite, lherzolite is not primitive upper mantle material, and would occur only in regions where basalt generation has taken place. The preponderance of lherzolite as a xenolith type in basalts may be more a function of its local abundance in the region through which the basalt was erupted, and of its greater stability in the presence of basaltic melt (Kutolin, 1970), than of its ubiquity in the upper mantle in general.

In the model presented here, garnet pyroxenite is more primitive upper mantle material than spinel lherzolite. A comparison of the major element compositions of typical garnet pyroxenite and spinel from Salt Lake Crater lherzolite with that of "pyrolite III" (Green and Ringwood, 1967) appears in Table 2. The densities of pyrolite III and typical spinel lherzolite from Salt Lake Crater are comparable. Density for the pyroxenites is variable, and depends on the extent to which recrystallization to form garnet has progressed. Xenoliths with about 5% garnet have densities near 3.38; those with garnet contents above 30% have densities at least 3.50 gms/cm^3. Press (1969) has shown that the lithoshere may have densities greater than those predicted by the pyrolite model. Salt Lake Crater garnet pyroxenites have densities consistent with the results of Press, and show textural evidence of having been derived from depths of about 60-80 km (Green, 1966). The uniformity of spinel lherzolite inclusions the worldover (Ross *et al.*, 1954) suggests that other lherzolites may have originated from pyroxenite in events similar to that preserved in the Salt Lake Crater xenoliths. If so, garnet pyroxenite may constitute a large proportion of the upper mantle.

ACKNOWLEDGEMENTS

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TABLE 1. Concentrations of the REE in whole-rock and separated mineral samples from xenoliths from Salt Lake Crater, Hawaii and Kilbourne Hole, New Mexico (ppm).*

	R7401 WR	R7401 CPX	R7401 Gt	R7404 CPX	R7444P WR	R7444L WR
La	2.57	3.65	0.018	5.30	1.56	0.49
Ce	5.81	9.56	...	...	4.26	0.98
Pr	0.97	1.70	0.29	3.97	0.56	0.14
Nd	4.97	7.93	2.87	22.18	2.50	0.85
Sm	...	...	(0.76)	((3.97))	0.76	0.20
Eu	0.64	1.27	0.56	1.60	0.27	0.067
Gd	2.26	2.55	4.95	6.73	1.05	0.21
Tb	...	...	0.73	0.70	0.20	0.037
Y	10.21	5.84	28.30	10.19	6.17	0.92
Ho	0.39	0.25	1.07	0.32	2.74	0.038
Er	1.00	0.53	3.31	1.45	0.87	...
Tm	0.14	0.062	0.48	0.13	0.089	...
Yb	0.73	0.25	...	0.66	0.65	...
Lu	...	...	0.48	0.12	0.40	...

* Missing data points are due variously to (1) incomplete separation of adjacent REE in the ion exchange procedure, (2) laboratory mishaps and (3) allowing insufficient time for the oxalate precipitations in early runs. Bracketed values correspond to elements for which the standard was lost in analysis. Double bracketed data are assumed values against which single bracketed data are compared.

TABLE 1. (Cont'd.)

R7444P	R7444L	R7634	R7634	R7629P	R7629P	R7629
CPX	CPX	WR	CPX	WR	CPX	CPX
2.88	2.67	4.30	4.80	1.45	2.17	2.90
...	...	...	15.42	3.97	...	6.20
1.18	1.08	2.28	2.88	0.56	0.67	1.00
6.37	5.60	12.97	15.41	3.02	4.54	5.34
1.66	1.43	3.31	3.94	0.89	1.33	1.54
0.60	0.44	1.28	1.42	0.31	0.43	0.56
2.21	2.09	5.63	6.62	1.46	1.78	2.10
(0.29)	(0.25)	0.55	0.85	0.28	((0.26))	0.41
(8.25)	(12.9)	16.3	19.2	7.17	((9.41))	5.59
0.25	0.31	0.84	0.95	0.35	0.28	...
0.67	...	1.71	2.02	0.80	0.68	...
0.074	0.12	0.30	0.34	0.12	0.087	0.16
0.32	0.70	1.76	1.87	0.74	0.50	0.90
0.044	0.094	0.26	0.30	0.096	0.068	0.14

TABLE 2. A comparison of major element compositions of pyrolite III, with typical Salt Lake Crater lherzolite and garnet pyroxenite.

	Pyrolite III		Typical Lherzolite*		Typical Garnet Pyroxenite**	
SiO_2	45.20		43.40		48.46	
Al_2O_3	3.54		2.29		7.96	
Fe_2O_3	0.48	8.47†	1.00	8.64†	3.55	8.12†
FeO	8.04		7.74		4.92	
MgO	37.48		42.98		19.96	
CaO	3.08		1.60		12.00	
Na_2O	0.57		0.17		1.90	
K_2O	0.13		<0.03		0.19	
TiO_2	0.71		0.15		0.43	
Cr_2O_3	0.43		0.42		0.54	
TOTAL	99.66		99.78		99.91	
MgO/("FeO")	4.43		4.97		2.46	

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* Lherzolite #2, Kuno, 1969, p. 194

**Pyroxenite #16, Kuno, 1969, p. 195

† Total iron as FeO ("FeO")

CHAPTER IV

STRONTIUM ISOTOPIC RELATIONSHIPS IN XENOLITHS AND THEIR HOST BASALT FROM SALT LAKE CRATER, HAWAII

Two goals motivated this strontium investigation: first, a desire to determine the presence or absence of a genetic relationship between the Salt Lake Crater xenoliths and their host basalt. Secondly, radiometric age data was obtained from the inclusions themselves, in an effort to find fragments of old and possibly primitive upper mantle material.

Are the xenoliths and the host basalt genetically related?

The host basalt at Salt Lake Crater is too young (Pleistocene to Recent, [Winchell, 1947; Jackson and Wright, 1970]) to have undergone a measurable increase in its Sr^{87}/Sr^{86} ratio since eruption. The xenoliths, with lower Rb/Sr than the basalts, are even less able to change their Sr^{87}/Sr^{86} in the same time interval. Hence xenoliths and host basalts showing the same isotopic compositions today, were in isotopic equilibrium at the time of eruption. Conversely, xenoliths now showing Sr^{87}/Sr^{86} ratios different from the host rock values were not in isotopic equilibrium at the time of eruption. There are three possible relationships between Sr^{87}/Sr^{86} ratios in the basalt and the xenoliths.

(1) Basalt and inclusions showing the same isotopic composition are possibly but not necessarily genetically related. (2) If the basalt $^{87}/^{86}$ ratio is higher than the xenolith values, it is possible but unlikely, that the two are cogenetic. (If a disequilibrium melting process selectively attacked minerals rich in alkalis [with high Rb/Sr] the resulting liquid might have a higher $^{87}/^{86}$ ratio than the residue. This requires that the parent material was not homogeneous with respect to Sr^{87}/Sr^{86}.) (3) Lower $^{87}/^{86}$ in the basalt than in the inclusions is strong evidence for the lack of a genetic relationship. The xenoliths, and particularly their clinopyroxenes, have very low Rb/Sr ratios. Since $^{87}/^{86}$ ratios lower than the initial ratio cannot be obtained in a closed system, a basalt with an $^{87}/^{86}$ ratio lower than its low-Rb/Sr xenoliths cannot be related in a simple genetic manner to those inclusions.

O'Neill *et al.* (1970) reported Sr isotopic compositions in some xenoliths and basalts from Oahu. They found no appreciable differences in Sr^{87}/Sr^{86} between a group of four Salt Lake Crater pyroxenites, and two nephelinitic basalts from vents other than Salt Lake Crater in the Honolulu Series group. From the similarity in these isotopic compositions, the authors conclude the possibility of a genetic relationship between the two groups of material.

Although sufficient data has not been taken to be conclusive, the Sr data of the present study suggests a different conclusion. Good samples of host basalt at Salt Lake Crater are rare. The xenoliths occur in an agglomerate tuff composed primarily of fragmental tholeiite and coral reef material. This material is clearly not representative of the host rock. Some xenoliths, on the other hand, show partial skins of dense fine grained nephelinite, which taper to zero thickness at their edges (Fig. IV-1). The suggestion is that these skins have crystallized against the inclusions prior to their most violent abrasion, and hence represent part of a liquid/inclusion system some distance below the surface. These basalt skins are considered the best samples of the composition of the Salt Lake Crater host basalt. Fig. IV-2 is a histogram of the Sr isotopic results of this study. Xenolithic minerals and whole-rock material are cross-hatched and show no overlap with the basalt values. The data suggest that the Salt Lake Crater xenoliths are not genetically related to the basalt which brought them to the surface.

Lower $^{87/86}$ ratios in the basalt than in the inclusions are consistent with the idea that the melt came from a greater depth in the upper mantle incorporating the xenolithic material in its path of ascent. Since Rb is more strongly fractionated than Sr in partial melting processes, the upper mantle should become depleted in Rb relative to Sr with depth and time. A reduction in Rb/Sr lowers the

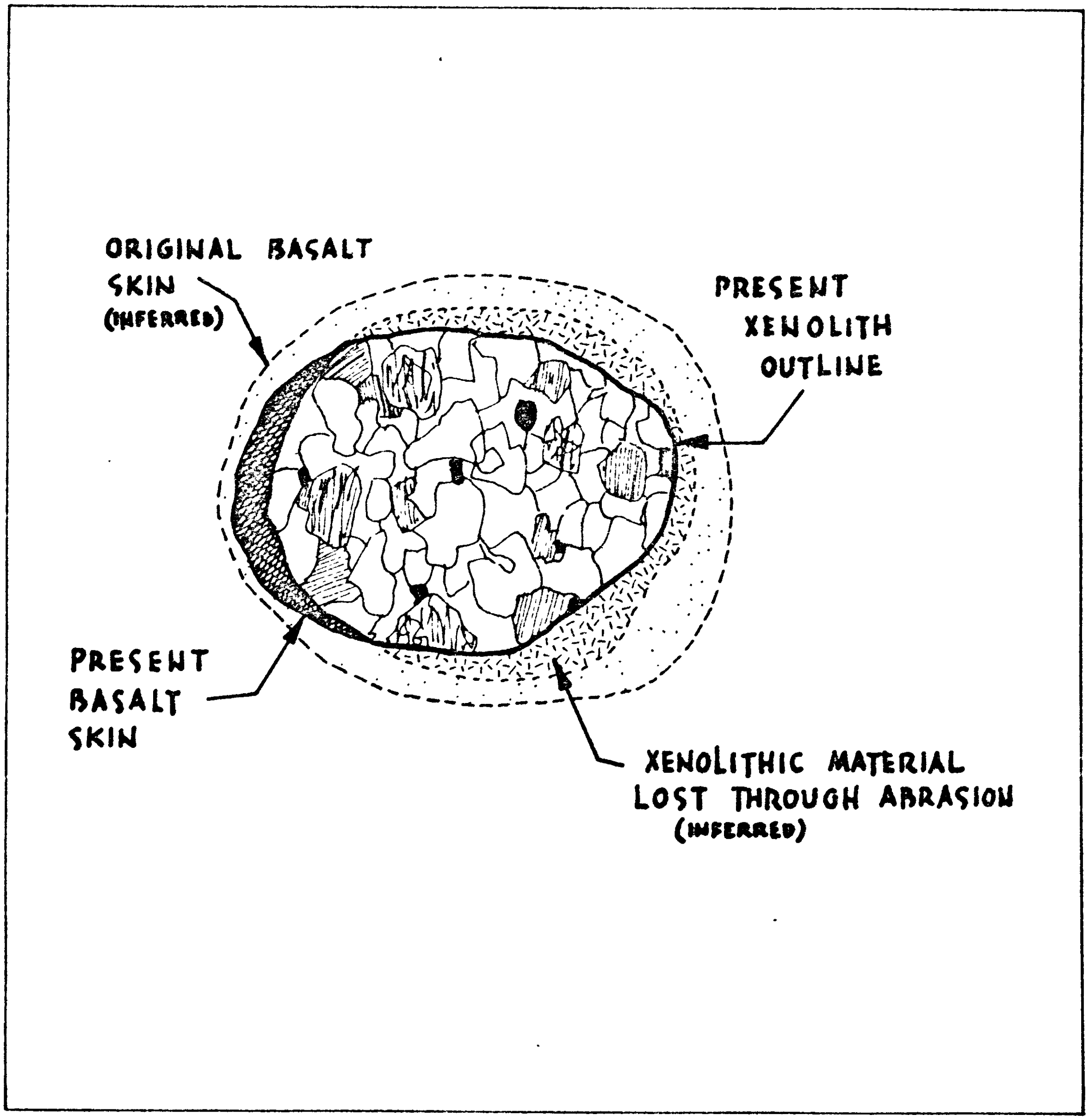

Figure IV-1: Schematic appearance of Salt Lake Crater xenoliths with partial basalt skins, considered the best samples of the Salt Lake Crater host basalt.

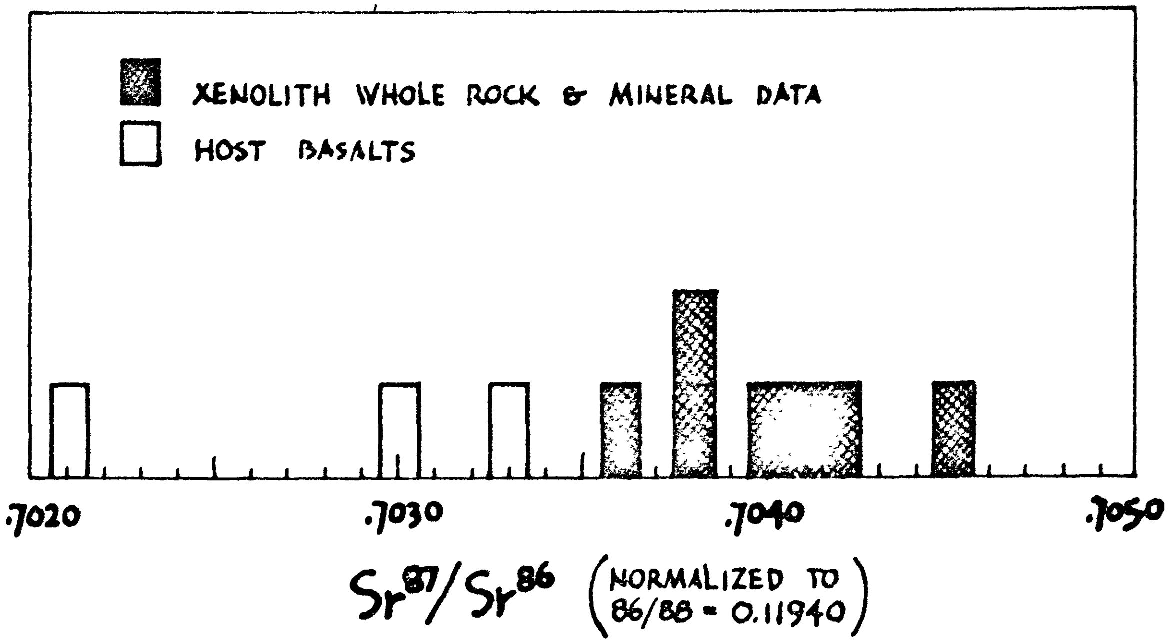

Figure IV-2: Sr^{87}/Sr^{86} ratios in xenolithic material and the host basalt from Salt Lake Crater, Hawaii.

ability to grow Sr^{87} relative to Sr^{86}, and with time the upper mantle should also show decreasing Sr^{87}/Sr^{86} with depth. Basalts derived from different levels should reflect this fractionation, with those generated at the greatest depths showing the lowest Sr^{87}/Sr^{86} ratios.

Based on major element compositions of some Salt Lake garnet pyroxenite xenoliths (the same samples analyzed by O'Neill *et al.*) Jackson and Wright demonstrated that certain pairs of pyroxenites have chemistry consistent with their being parent and residue respectively to the production of the Honolulu Series basalts. The Sr data of this study do not support this conclusion. Textural evidence in the pyroxenite suite is also inconsistent with the conclusions of Jackson and Wright. The garnet pyroxenites proposed as residues (68-SAL6, 68-SAL24) contain textures which are evidence of extensive cooling at very nearly isobaric conditions. These textures, discussed in Chapter II, involve the abundant exsolution from clinopyroxene of orthopyroxene and garnet, and the formation of additional garnet by the subsolidus recrystallization of spinel and clinopyroxene. A cooling from 1300 or 1400°C to about 1100°C, at about 20 kb has been inferred. It is difficult to envision an isobaric process at a depth of 70 km involving the rapid cooling of material through 200-300°C. Hence it is unreasonable that rocks showing evidence of such extended cooling could have been residual in the formation of the basalt in which they were carried to the surface, particularly in light of the highly explosive nature of the Salt Lake Crater eruption.

Possible fragments of ancient upper mantle material

A second goal of the strontium investigation has been the search for old fragments of the upper mantle. Again the data reported are too few to be conclusive, but there is indication that such a search would not go unrewarded.

Caution must be exercised in the interpretation of strontium isotopic data from ultramafic inclusions of supposed mantle origin. Erlank (1969) has shown that the garnets from xenoliths from Roberts-Victor Mine, South Africa, contain alkali-rich alteration rims, and crack-filling material. In some cases removal of this material is impossible, rendering trace element data on the present mineral systems useless as information about the upper mantle. Contamination from crustal material is also a danger in the study of xenoliths brought to the surface in continental basalts. The search for old mantle fragments in xenolith suites is a search for high Sr^{87}/Sr^{86} ratios. Continental crustal material has relatively high Rb/Sr ratios and high Rb and Sr contents. Through time, large amounts of radiogenic Sr^{87} are developed, and through contamination, can cause large changes in the Sr^{87}/Sr^{86} ratios of xenoliths with low Sr contents. Bence (1966) found no Hawaiian basalt with Sr^{87}/Sr^{86} outside the range from about 0.703 to 0.707. Hence xenolithic material from the Hawaiian Islands which shows Sr^{87}/Sr^{86} ratios above 0.710 or so, must be a fragment of old mantle material, since contamination from the host rocks could raise the xenolith ratio no higher than about 0.707.

A garnet- and phlogopite-bearing pyroxenite xenolith (R7399-SLC6) from Salt Lake Crater has given some promising results. The Rb and Sr data for it are shown in Table 1. Sr^{87}/Sr^{86} determined for the clinopyroxene, phlogopite and whole-rock fall in a narrow range from 0.7036 to 0.7041. The very low Rb/Sr ratio in the clinopyroxene allows its Sr^{87}/Sr^{86} ratio to be considered the initial ratio for the rock. The near-equivalence of the Rb-rich phlogopite and whole-rock values with the clinopyroxene ratio suggest either that the rock has recently formed or that the phlogopite has continually equilibrated with the clinopyroxene through time. The high Sr^{87}/Sr^{86} values in the garnet indicate the likelihood of the second alternative.

It is important to establish that the garnet $^{87/86}$ values are real properties of the mineral itself, especially in light of the low Sr concentrations it contains. Several factors might produce apparently high Sr^{87}/Sr^{86} ratios in samples with low Sr contents and otherwise low Sr^{87}/Sr^{86}

TABLE IV-1. Rb, Sr and Sr isotopic results for R7399.

	Rb(ppm)	Sr(ppm)	Sr^{87}/Sr^{86}
Gar #1	...	...	0.718
Gar #2	0.037	0.514	0.740
Whole-Rock	...	...	0.7041
Phlog	...	...	0.7036
Gar #3	...	...	0.7105
Cpx	0.100	130.4	0.7038
Cpx	...	...	0.7038
Gar #4	...	0.866	0.715
Gar #5	...	...	0.727

ratios. (1) Memory in the mass spectrometer of spiked Rb runs could raise the measured $^{87}/^{86}$ ratio, by contributing Rb^{87} to the ^{87} peak despite showing a low Rb^{85} peak, which indicates a small normal Rb correction. The fact that the Rb at the outset of these runs was normal, and that the measured $^{87}/^{86}$ ratio was quite constant over the course of the runs, suggests that this is not the explanation of the high garnet $^{87}/^{86}$ ratios. (2) Double ionization of ytterbium, whose isotopes at masses 174 and 172 are in the proper relative abundance to cause an increase in the measured $^{87}/^{86}$ ratio, is a possibility. Calculations show, however, that if doubly ionized Yb were present in the beam in sufficient amount to cause a run whose true $^{87}/^{86}$ ratio was 0.704 to appear to be 0.727, for example, the measured $^{86}/^{88}$ ratio would be 0.1160. No run has shown a $^{86}/^{88}$ ratio further than 0.0007 from the normalization value of 0.1194. Hence it is unlikely that the garnets show high $^{87}/^{86}$ ratios as a result of rare earth interferences.

A third possibility is that of contamination introduced in the chemical procedure. In the course of Sr analyses of lunar material, Hurley (personal communication, 1970) has found that most of the Sr blank level is introduced in the ion exchange procedure. Although whole-process blanks for Rb and Sr have shown measured levels of 7 nanograms Rb and 16 nanograms Sr in the present study, the runs on which these values were determined were very badly overspiked, and hence are unreliable. The measured Sr contents of the garnet are between 500 and 800 nanograms/gm, and contamination could have been an appreciable fraction of the analyzed Sr. Sr^{87}/Sr^{86} values of 0.715 to 0.740 might have been attributed to contamination from the ion exchange columns if those columns had been used in the analyses of Precambrian continental rocks whose $^{87}/^{86}$ values often exceed 1.0. However, the columns themselves, and the resin, were new at the start of these analyses. No Sr samples other than those from ultramafic xenoliths and their host basalts were eluted through the columns, so that contamination from the columns is not a likely explanation of the high $^{87}/^{86}$ ratios displayed by the garnet, though it is a likely source of the variability the garnet $^{87}/^{86}$ ratios show. Laboratory dust, with high $^{87}/^{86}$ values, could give the observed garnet numbers, but if it is responsible, the relative contributions to the blank of the various parts of the procedure observed by Hurley would have to have been different in this work.

The measured garnet ratios therefore are probably real. Taken together, the Rb/Sr and Sr isotopic data for R7399-SLC6 do not form an isochron. The results suggest that garnet and clinopyroxene have acted as sinks for the radiogenic Sr^{87} produced in the phlogopite, which it easily loses because of the loose fit of the Sr^{87} atom in the site formerly occupied by its larger radioactive parent, Rb^{87}. The isotopic disequilibrium between garnet and clinopyroxene requires that temperatures were not high enough to allow exchange of Sr^{87} between these minerals. Recent results indicate the persistence of isotopic disequilibrium at these conditions is possible. Peterman *et al.* (1970) report data for a lherzolite from Dish Hill, California, showing clear isotopic disequilibrium between the chrome diopside, olivine, and orthopyroxene. Over short periods of time (on the order of tens of millions of years), the addition of radiogenic Sr^{87} from the mica to the Sr-rich clinopyroxene would cause no observable change in the pyroxene $^{87}/^{86}$ ratio. Similar additions to the very Sr-poor garnet over the same times would measurably alter its isotopic composition.

An estimate of the minimum age of the formation of garnet in R7399-SLC6 can be made by assuming that at that time the garnet's $^{87}/^{86}$ ratio fell on the mantle growth curve, and that it and the clinopyroxene absorbed radiogenic Sr from the mica in proportion to their modal abundances. (It is more likely that each would absorb Sr in proportion to its Sr content, but the present assumption will give a lower limit on the age of garnet formation.) Garnet with 0.5 ppm Sr requires the addition of 0.0012 ppm radiogenic Sr^{87} to change its Sr^{87}/Sr^{86} from 0.704 to 0.727. Proportionate absorbtion of Sr^{87} by pyroxene requires a total of about 0.014 μgm Sr^{87} to have been produced for each gram of garnet, or by each 0.2 gm of phlogopite, whose modal abundance in R7399-SLC6 is about 1/5 that of garnet. Phlogopite with 150 ppm Rb can produce this Sr^{87} in about 35 million years. If garnet and clinopyroxene absorb radiogenic Sr^{87} in proportion to their Sr concentrations, the age of garnet formation would be on the order of 1.5 billion years. This is an estimate of the age of an event which took place in the rock at some time after its formation as a garnet-free assemblage. The time necessary for that garnet-free assemblage to have cooled to the point where the garnet forming reaction could take place is unknown. The rock itself would be older than 1.5 billion years. Textural data (Chapter II) and REE data (Chapter III) indicate that the spinel-lherzolites occurring world-wide as inclusions in basalt have formed from pre-existing garnet pyroxenites like R7399-SLC6. An old age for pyroxenite material is consistent with this view.

The variability in $^{87}/^{86}$ ratio in the garnet from R7399-SLC6 is probably due to three factors. (1) Garnet and phlogopite are unevenly distributed with respect to each other in the rock. Migration of Sr^{87} from the mica might be expected to affect garnet near the mica more than more distant garnet. (2) Garnet varies in grain size and in the degree to which it is fractured. $CaCO_3$ is present along most fractures in the rock, and although care was taken to leach such carbonate from the garnet separates using an ultrasonic cleaner, remnant carbonate would contaminate Sr^{87}/Sr^{86} ratios in the garnet in varying degrees toward about 0.708. (3) Common Sr from the ion exchange columns may have altered the garnet ratios.

The petrographic and REE study of the xenolith suite from Salt Lake Crater has suggested worldwide importance for this material in the composition of the upper mantle. In the course of the rare earth part of this study, a pyroxenite xenolith (similar to R7399-SLC6) and its constituent garnet and clinopyroxene were analyzed for their REE patterns (Chapter III, Figure 1). When the garnet and clinopyroxene REE patterns are added together in their modal proportions, the fit to the whole-rock pattern is very close. The data indicate that, unlike the Roberts-Victor inclusions (Erlank, 1969), the Salt Lake Crater xenoliths have *not* been contaminated by the host basalt, and hence studies of their trace element distributions constitute valuable aids in understanding the upper mantle. The strontium isotopic part of this study is not sufficiently complete to warrant conclusions of a general sort. It has, however, shown that the xenoliths at Salt Lake Crater deserve a closer look, and with more detailed strontium isotopic work, they very likely would reveal new information about the age of the upper mantle and the processes by which crustal material is generated from it.

CHAPTER V

THE ROLES OF LHERZOLITE AND GARNET PYROXENITE IN THE CONSTITUTION OF THE UPPER MANTLE

Earth scientists commonly assume an upper mantle with a peridotitic composition. The assumption stems from models based on comparison of the earth with meteoritic, solar and stellar element abundances, and models based on terrestrial rocks thought to have been derived from the upper mantle.

The assumption that the earth's non-volatile element abundances are similar to those in carbonaceous chondrites is the basis of a calculation by Ringwood (1966) which allows an estimate of the composition of the mantle as a whole. Under the additional assumptions that the FeO/(FeO+MgO) in the mantle is 0.12, and that the earth's silicate/metal ratio is 69/31, he has shown that the chondrite model leads to composition for the whole mantle which compares closely with the pyrolite composition. This similarity is considered evidence for relatively little fractionation of the major rock-forming elements throughout the whole mantle. It would seem more appropriate if one of Ringwood's conclusions were an assumption; namely, that " if the whole mantle has not undergone appreciable fractionation of its major rock-forming elements, and if the overall earth has a chondritic composition in its non-volatile elements, then the overall mantle and pyrolite have similar compositions." The upper mantle to a depth of 150 km. constitutes only 8.2% of the volume of the whole mantle. Hence, the chondrite model could be valid, and the overall mantle could have a peridotitic composition, and yet the upper mantle may indeed have been fractionated.

The other source of evidence for a peridotitic upper mantle is based on terrestrial rocks supposed to have come from the upper mantle. Harris, Reay and White (1967) have noted the similarity in bulk composition between alpine ultramafic rocks, inclusions in basalts and inclusions in kimberlite. Although each group shows wide ranges in composition, the authors feel that "estimates of the composition of undepleted mantle derived independently from each source approach a common value" (p. 6359). The variation is largely in the rocks' contents of basaltic components, particularly CaO and Al_2O_3. They reason that rocks from each group showing the highest CaO and Al_2O_3 contents are representative of undepleted upper mantle material, and that others less rich in these elements represent residual material depleted to varying degrees in basaltic constituents. A possible alternative interpretation of this variation stems from the observation that most xenoliths are coarse-grained, and often show mineralogical banding. Xenoliths with greater-than-average contents of clinopyroxene may have originated in the same episode as nearby material richer in olivine, and yet analyses of these rocks would give the impression that the clinopyroxene-rich rock is more undepleted in CaO and Al_2O_3 and hence more primitive than the olivine-rich rock. The conclusions of Harris *et al.* hence are not necessarily indicative of peridotitic material from the mantle which shows a range of degrees of depletion in its basaltic components.

Typical spinel lherzolite occurs in xenoliths at Salt Lake Crater in contact with garnet pyroxenite. Several lines of evidence (textures, major element distributions, REE distributions, and Sr isotopic data) converge to the conclusion that garnet pyroxenite pre-dates the lherzolite, and that the lherzolite has formed from the pyroxenite. The striking similarities shared by the lherzolite inclusions of the study of Ross *et al.* (1954), White (1966), and Carter (1969) are indications that they have formed in similar events; that is, that garnet pyroxenite was present in regions of basalt production in many parts of the world. Salt Lake pyroxenites have textures which indicate that they have originated at depths of 60-75 km (Chapter II). Press (1969) has shown that the material from this region of the upper mantle may have density greater than that of lherzolite or pyrolite. Densities of separated minerals from Salt Lake garnet pyroxenites can be estimated from data given by Deer, Howie and Zussman (1963). Table V-1 shows modal analyses for the pyroxenites of the REE study, and the whole-rock densities calculated from them. The densities of Salt Lake Crater pyroxenite xenoliths fall within the cluster of curves for models found to satisfy Press' model. (Inclusions richer in garnet than R7444, R7401 or R7629 occur at Salt Lake Crater, and would have somewhat higher densities [3.50-3.55]). Hence, the pyroxenites have appropriate densities and come from the proper depths to qualify as the dominant material of the upper mantle. The uniformity and ubiquity of lherzolite xenoliths has been considered evidence favoring the idea that the upper mantle has a lherzolite or peridotite composition. However, if typical lherzolite at Salt Lake Crater has formed from more primitive pyroxenite, the uniformity and ubiquity of lherzolite may be more of an indication of the pervasive distribution of the pyroxenitic material from which they formed, rather than of the widespread occurrence of the lherzolites themselves. Lherzolite would occur only in regions of the upper mantle where basalt production and differentiation has taken place. The relative scarcity of pyroxenite as a xenolith type would result if it were largely consumed by basalt production, and because it is less likely to survive the trip to the surface in the host basalt than lherzolite which is residual in nature and hence relatively more stable in the presence of basaltic melt (Kutolin, 1970).

TABLE V-1. Modal percentages and densities for some Salt Lake Crater xenoliths.

	$\rho=3.75^*$	$\rho=3.75^*$	3.34^*	3.34^*	3.36^*	
	SPINEL	GARNET	CPX	OPX	OLIVINE	ρ
R7401	2.9	18.0%	65.0	9.0	5.1	3.43
R7444	8.0	18.0	37.5	36.5	...	3.45
R7629	0.9	9.2	66.2	23.7	tr	3.38

As noted by Jackson (1966), some pyroxenites have up to 50% garnet. (Density for 50% clinopyroxene, 50% garnet is 3.54.) Garnet shows evidence of being preferentially melted in all xenoliths observed in thin section (about 30). The xenolith population at Salt Lake Crater may be poorer in garnet than the region from which they came, if garnet-rich xenoliths were more easily disaggregated in the eruption as a result of garnet's melting.

[*] Densities for spinel, clinopyroxene, orthopyroxene and olivine estimated from data in Deer, Howie and Zussman, 1962. Garnet and spinel have nearly identical densities as seen in Clerici separations.

The entire discussion applies only to the region in the mantle from which spinel lherzolite is derived. Alpine peridotites probably come from shallower depths; the xenoliths in kimberlites from depths greater than about 100 km. The fact that Press' model cannot resolve features of the upper mantle shallower than about 50 km enables nothing to be said of this sort about the outermost upper mantle. Below about 150 km Press' model predicts densities more nearly in accord with the pyrolite model. It is therefore possible that olivine-rich compositions, perhaps material similar to the garnet peridotites from kimberlites, are representative of the lower portions of the upper mantle. It is argued here, however, that regions of the upper mantle between 50 and 150 km do not have compositions rich in forsteritic olivine, and that the garnet pyroxenites from Salt Lake Crater, Hawaii, may be representative of this zone of the upper mantle.

CHAPTER VI

SUGGESTIONS FOR FURTHER RESEARCH

The main purpose of the present study has been to determine the genetic relationship between pyroxenite and lherzolite in mutual contact. The decision as to which is the older material has a large effect on proposed upper mantle compositions. A very useful study would be the determination of the Rb-Sr ages of associated lherzolite and garnet pyroxenite. This would involve the analysis of whole-rock and separated mineral samples from the two rock types for their Sr^{87}/Sr^{86} and Rb^{87}/Sr^{86}.

More major element analyses of near-contact lherzolite minerals using the electron microprobe would be helpful in clarifying the process by which lherzolite appears to form from pyroxenite. Detailed mapping of the variation of Cr_2O_3 and Al_2O_3 in coexisting spinel and chrome diopside in the lherzolite portions of two-assemblage xenoliths is a first priority. It would also be interesting to know the precise variation of Mg/Fe as a function of distance perpendicular to the contact, and radially from the centers of olivine, orthopyroxene and chrome diopside grains. Type 1 pyroxene grains at the contact (pyroxenite clinopyroxene and lherzolite orthopyroxene in a single grain) would also be interesting to carefully map with the microprobe for their total major element compositions, particularly their Mg/Fe and CaO contents. Further microprobe analyses of the crack-filling material near the lherzolite-pyroxenite contact are also suggested by the present study.

The fission track technique is capable of showing the microscopic distribution of once trace element, uranium, relative to the textures of the rock. Maps of uranium distribution in minerals on both sides of the contact might clarify processes such as suggested in the rare earth study. Lherzolite chrome diopside and its associated pyroxenite clinopyroxene have very similar REE patterns. With the fission track technique, one could get detailed and possible elucidating information on the distribution of uranium across transitional grains of the types shown in Figure 5, Chapter III.

Additional REE analyses could be very informative. R7444 contains about 18% glass, which has a major element composition similar (except for excess alkalis) to the garnet from which it formed. Either by use of heavy liquids, or by selective dissolution, the glass could be removed and analyzed for its REE distribution. Since the glass may contain components of the liquid performing the pyroxenite-lherzolite transformation (see Chapter III), a knowledge of its REE pattern, and hence its relationship to Hawaiian lavas would be most useful.

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