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The Genesis of Basaltic Magmas

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Contents

Abstract. This paper reports the results of a detailed experimental investigation of fractionation of natural basaltic compositions under conditions of high pressure and high temperature. A single stage, piston-cylinder apparatus has been used in the pressure range up to 27 kb and at temperatures up to 1500° C to study the melting behaviour of several basaltic compositions. The compositions chosen are olivine-rich (20% or more normative olivine) and include olivine tholeiite (12% normative hypersthene), olivine basalt (1% normative hypersthene) alkali olivine basalt (2% normative nepheline) and picrite (3% normative hypersthene). The liquidus phases of the olivine tholeiite and olivine basalt are olivine at 1 Atmosphere, 4.5 kb and 9 kb, orthopyroxene at 13.5 and 18 kb, clinopyroxene at 22.5 kb and garnet at 27 kb. In the alkali olivine basalt composition, the liquidus phases are olivine at 1 Atmosphere and 9 kb, orthopyroxene with clinopyroxene at 13.5 kb, clinopyroxene at 18 kb and garnet at 27 kb. The sequence of appearance of phases below the liquidus has also been studied in detail. The electron probe micro-analyser has been used to make partial quantitative analyses of olivines, orthopyroxenes, clinopyroxenes and garnets which have crystallized at high pressure.

These experimental and analytical results are used to determine the directions of fractionation of basaltic magmas during crystallization over a wide range of pressures. At pressures corresponding to depths of 35—70 km separation of aluminous enstatite from olivine tholeiite magma produces a direct fractionation trend from olivine tholeiites through olivine basalts to alkali olivine basalts. Co-precipitation of sub-calcic, aluminous clinopyroxene with the orthopyroxene in the more undersaturated compositions of this sequence produces derivative liquids of basanite type. Magmas of alkali olivine basalt and basanite type represent the lower temperature liquids derived by approximately 30% crystallization of olivine-rich tholeiite at 35—70 km depth. At depths of about 30 km, fractionation of olivine-rich tholeiite with separation of both olivine and low-alumina enstatite, joined at lower temperatures by sub-calcic clinopyroxene, leads to derivative liquids with relatively constant SiO_2 (48 to 50%) increasingly high Al_2O_3 (15—17%) contents and retaining olivine + hypersthene normative chemistry (5—15% normative olivine). These have the composition of typical high-alumina olivine tholeiites. The effects of low pressure fractionation may be superimposed on magma compositions derived from various depths within the mantle. These lead to divergence of the alkali olivine basalt and tholeiitic series but convergence of both the low-alumina and high-alumina tholeiites towards quartz tholeiite derivative liquids.

The general problem of derivation of basaltic magmas from a mantle of peridotitic composition is discussed in some detail. Magmas are considered to be a consequence of partial melting but the composition of a magma is determined not by the depth of partial melting but by the depth at which magma segregation from residual crystals occurs. Magma generation from parental peridotite (pyrolite) at depths up to 100 km involves liquid-crystal equilibria between basaltic liquids and olivine + aluminous pyroxenes and does not involve garnet. At 35—70 km depth, basaltic liquids segregating from a pyrolite mantle will be of alkali olivine basalt type with about 20% partial melting but with increasing degrees of partial melting, liquids will change to olivine-rich tholeiite type with about 30% melting. If the depth of magma segregation is about 30 km, then magmas produced by 20—25% partial melting will be of high-alumina olivine tholeiite type, similar to the "oceanic tholeiites" occurring on the sea floor along the mid-oceanic ridges.

Hypotheses of magma fractionation and generation by partial melting are considered in relation to the abundances and ratios of trace elements and in relation to isotopic abundance data on natural basalts. It is shown that there is a group of elements (including K, Ti, P, U, Th, Ba, Rb, Sr, Cs, Zr, Hf and the rare-earth elements) which show enrichment factors in alkali olivine basalts and in some tholeiites, which are inconsistent with simple crystal fractionation relationships between the magma types. This group of elements has been called "incompatible elements" referring to their inability to substitute to any appreciable extent in the major minerals of the upper mantle (olivine, aluminous pyroxenes). Because of the lack of temperature contrast between magma and wall-rock for a body of magma near to its depth of segregation in the mantle, cooling of the magma involves complementary processes of reaction with the wall-rock, including selective melting and extraction of the lowest melting fraction. The "incompatible elements" are probably highly concentrated in the lowest melting fraction of the pyrolite. The production of large overall enrichments in "incompatible elements" in a magma by reaction with and highly selective sampling of large volumes of mantle wall-rock during slow ascent of a magma is considered to be a normal, complementary process to crystal fractionation in the mantle. This process has been called "wall-rock reaction". Magma generation in the mantle is rarely a simple, closed-system partial melting process and the isotopic abundances and "incompatible element" abundances of a basalt as observed at the earth's surface may be largely determined by the degree of reaction with the mantle or lower crustal wall-rocks and bear little relation to the abundances and ratios of the original parental mantle material (pyrolite).

Occurrences of cognate xenoliths and xenocrysts in basalts are considered in relation to the experimental data on liquid-crystal equilibria at high pressure. It is inferred that the lherzolite nodules largely represent residual material after extraction of alkali olivine basalt from mantle pyrolite or pyrolite which has been selectively depleted in "incompatible elements" by wall-rock reaction processes. Lherzolite nodules included in tholeiitic magmas would melt to a relatively large extent and disintegrate, but would have a largely refractory character if included in alkali olivine basalt magma. Other examples of xenocrystal material in basalts are shown to be probable liquidus crystals or accumulates at high pressure from basaltic magma and provide a useful link between the experimental study and natural processes.

Introduction

Basaltic rocks, by virtue of their widespread occurence in both oceanic and continental regions, occupy a unique position in igneous petrogeny. It is widely believed that basaltic magmas are derived from the earth's mantle. Basalts accordingly contain a vast reservoir of information, the correct interpretation of which should throw a great deal of light on the chemical and mineralogical nature of the mantle. Petrologists have long recognized the existence of a number of different types of basalt characterized by distinctive chemical and mineralogical properties, e.g. alkali basalts, olivine tholeiites, high-alumina basalts and quartz tholeiites. The main theme of this paper is a detailed experimental study of the relationships between the principal magma types, their fractionation trends at various pressures and their derivation by partial melting processes in the peridotitic mantle.

Recent workers have attempted to systematise the classification and nomenclature of basaltic rocks and the reader is referred particularly to the excellent summaries of previous and present views on basalt classification by Yoder and Tilley (1962, p. 346—348, 349—356), Macdonald and Katsura (1964), Coombs (1963), Kuno (1960). We shall adhere to a classification of basaltic rocks similar to that accepted by Yoder and Tilley (1962) and Macdonald and Katsura

(1964). Since our main concern is with chemistry rather than the modal mineralogy of basaltic rocks, it is convenient to accept a classification directly based on CIPW normative mineralogy (i.e. an indirect chemical classification). The "basalt tetrahedron" of YODER and TILLEY (1962; Fig. 1) is a convenient and simple illustration of this method of classification of basaltic rocks. The principal classes of basalts are clearly represented as primary phase volumes in this figure. The following terms are used to define these principal magma types:

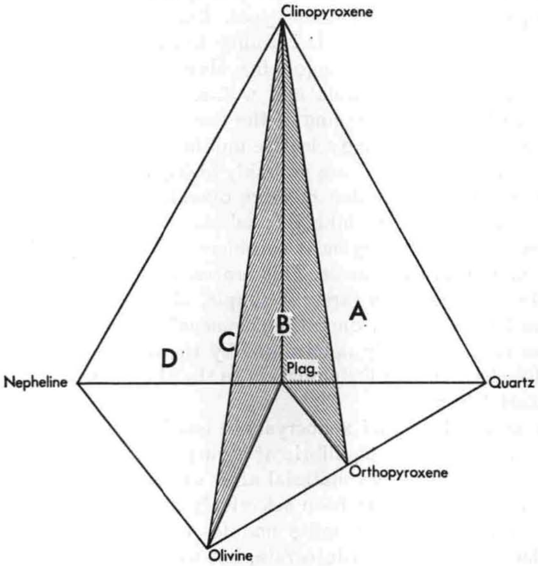

Fig. 1. Diagrammatic representation of the major mineralogy of basalts using the "basalt tetrahedron" (YODER and TILLEY, 1962). The plane olivine—clinopyroxene—plagioclase is referred to as the "critical plane of undersaturation". *A* field of quartz tholeiites; *B* field of olivine tholeiites; *C* field of alkali olivine basalts; *D* field of olivine basanites

Tholeiite or tholeiitic suite: basalts characterised by the presence of normative hypersthene.

Quartz Tholeiite: — basalts with normative quartz and hypersthene.

Olivine Tholeiite: — basalts with normative olivine and hypersthene. *Olivine basalt* is used for brevity and convenience in nomenclature for basalts falling within this group but having low (0—3%) normative hypersthene content. This convention is similar to the use of the term by YODER and TILLEY (1962).

Alkali[1] *Olivine Basalt:* — basalts characterised by the presence of normative olivine and nepheline but with normative nepheline less than 5%.

Basanite: — basalts characterized by normative olivine and nepheline and with more than 5% nepheline.

Olivine Nephelinite: — basalt-like rocks without normative albite and typically with olivine, diopside and nepheline as the major normative minerals.

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[1] The term alkali olivine basalt is preferred to the more grammatically correct "alkaline olivine basalt" or "alkalic olivine basalt" (MACDONALD and KATSURA, 1964, p. 89) for reasons of common usage and euphony.

The use of a classification based on the calculation of the CIPW norm enables objective grouping of analyzed basalts but it creates an artificial clarity of classification which is absent from the natural rocks. Apart from uncertainties introduced by the quality of chemical analyses and by post-magmatic or late magmatic changes in oxidation state, it is clear that the "critical plane of undersaturation" (Fig. 1) dividing olivine + nepheline normative basalts from olivine + hypersthene normative basalts is a finite zone of indeterminacy in basalt classification. Thus, POLDERVAART (1962, 1964) argued that a group of hypersthene-normative basalts were nevertheless petrographically and genetically related to the alkali olivine basaltic suite. These hypersthene-normative alkali olivine basalts characteristically have $Ab - 2\,En_{Hy} - 1.5\,Fs_{Hy} > 0$, while olivine tholeiites have values $Ab - 2\,En_{Hy} - 1.5\,Fs_{Hy} < 0$ (Ab, En_{Hy}, Fs_{Hy} are the normative albite, enstatite in hypersthene, and ferrosilite in hypersthene, respectively). YODER and TILLEY (1962, p. 404) expressed similar uncertainty about the affinities of olivine basalts with low normative hypersthene contents. The data complied by COOMBS (1963) and his definition of an "indicator ratio" to classify the basalt magma types are in good agreement with POLDERVAART's criteria for separation of the two suites. On the other hand, MACDONALD and KATSURA (1964) use a simple total alkalies ($Na_2O + K_2O$ in weight per cent) vs silica plot to distinguish magmas of alkali olivine basalt type from the tholeiitic suite.

The test of any chemical method of classification is in comparison with the modal mineralogy, i.e. (a) the presence of phenocrystal and groundmass olivine in the alkali olivine basalt and absence of calcium-poor pyroxene; (b) the presence of calcium-poor pyroxene and partially resorbed or mantled phenocrystal olivine in the olivine tholeiites. However, such petrographic classification is frequently nondefinitive because of uncertainty in identification of groundmass and because of the presence of common glass within the rock. In general, these methods of basalt classification work equally well for rocks removed from the transition zone between alkali olivine basalts and olivine tholeiites but all are subjective and may be mutually inconsistent when applied to rocks within the transition zone. Nevertheless a fundamental point is that all the above attempts at classification recognise the existence of a continuous spectrum of basalt compositions between alkali olivine basalt magmas and olivine tholeiites.

Previous Investigations

The general question of possible fractionation trends of alkali olivine basalt and tholeiites at low pressure has been fully discussed by YODER and TILLEY (1962). These authors point out that the "critical plane of silica undersaturation" $Cpx - Ol - Plag$ (Fig. 1) separating the alkali basalts from the olivine tholeiites represents a "thermal barrier" for fractionation at low pressures and that under normal low pressure conditions, fractionation trends in basaltic magmas move *away* from this plane on either side. The recognition of different basalt magma types and of a continuous spectrum of compositions between alkali basalts and tholeiites has accordingly posed a major problem in petrogenesis. One of the earliest hypotheses aimed at solving this problem involves the separation of hypersthene in place of olivine from olivine tholeiite magmas, leading to derivative liquids of alkali basalt type (HOLMES and HARWOOD, 1932; POWERS, 1935, p. 65; LARSEN, 1940, p. 926; MACDONALD, 1949, p. 1576; and TILLEY, 1950, p. 45.).

An important step in the study of basalt fractionation was the initiation of detailed experimental work at high pressures on natural basalts (YODER and TILLEY, 1962; GREEN and RINGWOOD, 1964; TILLEY and YODER, 1964). YODER and TILLEY (1962, p. 507) appealed to the removal of garnet from a parental liquid at high pressures to produce an alkali basaltic liquid while, in contrast, removal of omphacite and effective enrichment of garnet in the liquid would produce tholeiite-type liquids. Calculations of the effects of extracting either garnet or omphacite from actual basaltic compositions as extruded at the surface show that the residual liquids do not possess typical basaltic compositions and do not follow the trend suggested (pages 153—156). Accordingly this hypothesis must be rejected in its simplest form. Nevertheless, YODER and TILLEY's results are important since they demonstrated a new control of fractionation in basalt magmas at high pressure and that the low pressure "thermal divide" between alkali basalts and their derivatives and olivine tholeiites and their derivatives was absent at high pressure.

GREEN and RINGWOOD (1964) demonstrated experimentally that orthopyroxene replaced olivine as the liquidus phase in an olivine-rich tholeiite at pressures from 13 to 18 kb. The orthopyroxene was shown to be rich in Al_2O_3 and the extraction of aluminous orthopyroxene gave a very direct fractionation trend from olivine tholeiites to derivative alkali olivine basalt magmas. These experiments, although of a preliminary and reconnaissance nature, strongly supported earlier hypotheses on the role of hypersthene crystallization and demonstrated a fundamental change in the fractionation of olivine-rich basaltic magmas at modest pressures. The data provided a simple and very direct fractionation trend from olivine and hypersthene normative olivine tholeiites to olivine and nepheline normative alkali olivine basalt by separation of a single phase.

Present Investigation

This paper presents the results of some 120 experimental runs over the period 1963—1966 investigating the melting relations of selected olivine-rich basaltic rocks. The study was aimed at extending and confirming the preliminary data on the role of aluminous orthopyroxene in basalt fractionation and obtaining further information on the possible direction of fractionation of basalts and effects of partial melting in the mantle at moderate to high pressures. Our approach to the problem has been to select an olivine tholeiite composition considered for petrographic reasons to be a possible "primitive" composition. The nature, compositions and proportions of phases which crystallize between the liquidus and solidus of the olivine tholeiite have been determined over a wide range of pressures. This information, together with knowledge of the original composition, allows the fractionation trends and compositions of residual liquids to be established for different load pressure conditions. Within a limited range of pressures (between 12 and 18 kb) it was found by these means that the residual liquid was fractionating directly *towards* the alkali olivine basalt field. Accordingly, a new basalt composition lying on the general path of fractionation of the olivine tholeiite, and intermediate in composition between the olivine tholeiite and a typical alkali olivine basalt was prepared. The composition of this basalt would be obtained by a comparatively small degree of crystallization of the parental olivine tholeiite within the specified pressure interval. This second basalt had a composition lying close to the critically undersaturated plane (Fig. 1), possessing 1 % of normative hypersthene. A detailed study of the crystallization behaviour of this second basalt, called "olivine basalt" for convenience (YODER and TILLEY, 1962) was then carried out. It was found that within the above pressure range, the residual liquids resulting from crystallization of the olivine basalt followed a fractionation trend directly across the critically undersaturated plane and into the alkali olivine basalt field. The fractionation trend was followed further into the critically undersaturated field in an exactly analogous way by the detailed study of a typical alkali olivine basalt with a composition just within the critically undersaturated field.

The objective, in studying the crystallization behaviour of a series of related basaltic compositions linked by a common fractionation trend between 12 and 18 kb, was to established the nature and limitations of this fractionation trend in the most unambiguous way possible. Assumptions and extrapolations concerning the proportion of phases which crystallize, their compositions, and changes in oxidation states are minimised by this procedure. Some doubts had previously been expressed regarding the implications of GREEN and RINGWOOD's (1964) preliminary results and interpretation on the fractionation of the parental olivine tholeiite (O'HARA, 1965; TILLEY and YODER, 1964) because of the degree of crystallization which was assumed, and because of possible changes in oxidation state. The present methods of investigation should remove these doubts.

These methods also yield results directly applicable to problems of partial melting and magma segregation in the mantle. If the liquidus phases of the basaltic compositions are also stable in the sub-solidus mineral assemblage of a "parental peridotite" then the partial melting process can be regarded essentially as the reverse of the fractionation process. In this way we are able to apply our results to problems of partial melting deep within the mantle and attempt deductions on the nature and proportions of the basaltic liquids derived by partial melting.

a) "Primary" Basaltic Liquids

In a project designed to study the fractionation trends of basaltic magmas at high pressure it is obviously desirable to select those compositions which exist as liquids at depth and are not the end result of processes of crytallization and fractionation, or of contamination, at upper crustal levels. The liquids formed by processes of partial melting or complete melting of mantle rock are commonly referred to as" primary magmas" or "primary liquids" and these liquids may be considerably different in chemistry from the major basaltic magma types which we see extruded in bulk at the earth's surface. O'HARA (1965, p. 19—27) has given a thoughtful review and analysis of the arguments on the nature of primary basaltic liquids and concludes that the composition of extrusive basalt magmas is the product of the interplay of two processes (p. 27):

"(i) Partial melting of the mantle, yielding primary magmas which vary in composition according to the pressure, partial pressure of volatiles and degree of overheating which accompany the process.

(ii) Continuous crystallization differentiation during the movement of the liquids to the surface."

These general conclusions indicate that the compositions of liquids or glasses observed in natural rocks may not provide a reliable guide to liquid compositions at depth.

Studies of the currently active Kilauea and Mauna Loa volcanoes of Hawaii demonstrate most convincingly the role of fractionation at shallow levels in producing a series of bulk compositions ranging from picrite with over 30% normative olivine to tholeiites with variable normative quartz (MUIR and TILLEY, 1963; MACDONALD and KATSURA, 1961; MURATA and RICHTER, 1966a, b). The low-pressure fractionation in these Hawaiian tholeiites is dominated by settling of olivine. The most undersaturated basaltic glass from Kilauea Iki (1959 eruption) contains only 6.5% normative olivine (MACDONALD and KATSURA, 1961) but MACDONALD and KATSURA clearly recognize the role of olivine settling in making their estimate of the *average magma* composition for the 1959 Kilauea Iki lava lake. These authors note the tendency for the bulk composition of the erupted magma to become more mafic as the eruption proceeds and their estimated composition for the lava lake contains 18% normative olivine. The olivine tholeiite composition chosen for this study matches very closely with MACDONALD and KATSURA's estimated 1959 average Kilauea Iki lava lake composition[2]. The chosen olivine tholeiite composition also compares closely with a weighted average composition for the first eruptive phase of Kilauea Iki in 1959 (MURATA and RICHTER, 1966b) although the Al_2O_3 content is higher (13.1% compared to 11.4%) in our olivine tholeiite composition. There is, of course, no compelling evidence that these estimates represent the bulk composition of a Kilauean liquid at 30—60 km. However, the relatively rapid ascent of the Kilauea Iki magma (EATON and MURATA, 1960) deduced from seismic evidence suggests that, particularly in the most vigorous extrusive phase or slightly thereafter, the bulk composition of erupted magma may be virtually unmodified from the liquid composition existing at considerable depth (20—60 kms).

The estimation of liquid compositions at depth from the evidence of natural extrusive magmas is thus somewhat subjective and requires detailed investigations of the type which have been undertaken at Hawaii. Limitations may be deduced by restricting attention to extrusive bulk compositions with high $Mg/Mg + Fe^{++}$ values, as the generalization that the liquidus ferro-magnesian silicates are enriched in Mg relative to Fe, appears well founded. This limitation directs attention towards olivine-rich, picritic rocks but the distinction between possible primary liquids and shallow level accumulative picrites is currently impractical.

An alternative, more direct approach to this problems is to seek the composition of liquids forming at the minimum melting temperatures under high pressure in peridotitic compositions such as are believed to be present in the upper mantle. O'HARA (1963a, p. 71—77; 1963b, p. 116—118), DAVIS (1964, p. 165—171) and DAVIS and SCHAIRER (1965) have presented data relevant to this problem. O'HARA (1963a, 1963b) made a preliminary study of melting relations in various

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[2] MACDONALD and KATSURA (1961) list $Al_2O_3 = 12.9\%$ for the "parent magma" for the 1959 lava lake samples but this is derived from the average of columns 2 and 3, Table 1, and should be 12.1%. This error was followed in our own work with the Al_2O_3 content of the prepared olivine tholeiite being 13.1%.

mixes of natural minerals separated from a garnet peridotite and an eclogite. He reported lower solidus temperatures in olivine + garnet + pyroxene mixes than in simple garnet + pyroxene mixtures and, comparing his data with that of Yoder and Tilley (1962) on quartz-bearing eclogites, concluded that the whole range of garnet + pyroxene assemblages must be considered a thermal barrier at 30 kb between lower temperature liquids in either olivine-bearing or quartz-bearing eclogitic rocks. If this conclusion is valid then a quartz-bearing eclogite, or a liquid of that composition, cannot be derived by partial melting of a garnet peridotite at 30 kb.

Davis (1964) and Davis and Schairer (1965) have reported results on the system diopside—forsterite—pyrope at 40 kb. The minimum liquidus temperature in this system is at 1670 ± 10° C and at a composition $Di_{47}\,Py_{47}\,Fo_6$. Davis and Schairer deduce that the primary liquid for melting of the Ol + Ga + Cpx + Opx assemblage in the $CaO\text{-}MgO\text{-}Al_2O_3\text{-}SiO_2$ system will have a composition of approximately $Di_{47}\,Py_{47}\,Ens_3\,Fo_3$. Expressed in the CIPW norm to illustrate the approximate low pressure mineralogy, this composition gives $An_{32}\,Di_{22}\,Ens_{26}\,Fo_{20}$.

Thus, the studies at both 30 kb and 40 kb suggests that the compositions of liquids derived by partial melting of peridotite would contain around 20% or more olivine if expressed in the low pressure CIPW norm. Such liquids would be similar to rocks transitional between olivine tholeiites with around 10% olivine and tholeiitic picrites with 25—30% of olivine.

b) *Basaltic Compositions Chosen for Study*

In the previous section, evidence from natural volcanic sequences and from experimental studies at high pressure has been summarized, leading to the conclusion that liquids derived from partial melting of peridotite will have 20% or more normative olivine. In the initial selection of an olivine tholeiite composition for experimental study we used the estimate by Macdonald and Katsura (1961) of a "parental magma" composition of the Kilauea Iki 1959 lava lake. The composition used contains 20% olivine in the CIPW norm and, when expressed in an ideal eclogite norm, contains 46% garnet, 50% pyroxene, 4% ilmenite or alternatively 46% garnet, 49% pyroxene (Di + Jd + Ens) 3% olivine, 2% rutile depending on the assumed role of TiO_2. This compares closely in terms of mineral proportions, though not in mineral compositions, with Davis and Schairer's (1965) estimate of the reaction point Ol + Ga + Cpx + Opx + *Liq* at 40 kb in garnet, two-pyroxene, olivine assemblage, i.e. $Pyrope_{47}\,Diopside_{47}\,Ens_3\,Fo_3$. It is considered therefore that the olivine tholeiite composition is significant and relevant in studies of fractionation of basalts at high pressure. Conversely tholeiitic basalts with less than about 12% normative olivine cannot be considered "primary" basaltic liquids in the sense of being derived without fractionation from depths of more than 40 kms within the earth and are not relevant compositions for study of basalt fractionation at pressure in excess of 10 kb. Picritic compositions with more than 20% normative olivine (CIPW norm) may also be important as possible liquids or liquid + crystal derivatives from parental mantle peridotite.

The choice of an alkali basalt composition for experimental study was influenced both by the observed and inferred fractionation trends of the olivine tholeiite towards an alkali olivine basalt between 12 and 18 kb, and by petrological considerations of a more general nature. Thus most alkali olivine basalt liquids, as observed in extrusive or chilled rocks, have probably progressed some way along a low pressure fractionation trend controlled by olivine separation. The average "alkali basalt" (mostly olivine-bearing) of the Hawaiian Islands listed by MACDONALD and KATSURA (1964) contains 14% normative olivine and neither hypersthene nor nepheline in the norm. The average Hawaiian alkali olivine basalt listed by KUNO et al. (1957) contains 18% normative olivine and the average from Japan (KUNO, 1960) contains 17% normative olivine. The composition chosen for study contains high normative olivine (26%), low normative nepheline (2%) but

Table 1. *Chemical compositions, CIPW norms and "eclogite" norms of compositions used experimentally*

	Olivine tholeiite	Olivine basalt	Alkali olivine basalt	Picrite
SiO_2	46.95	47.05	45.39	45.51
TiO_2	2.02	2.31	2.52	1.93
Al_2O_3	13.10	14.17	14.69	12.44
Fe_2O_3	1.02	0.42	1.87	0.92
FeO	10.07	10.64	12.42	8.67
MnO	0.15	0.16	0.18	0.15
MgO	14.55	12.73	10.37	18.79
CaO	10.16	9.87	9.14	9.67
Na_2O	1.73	2.21	2.62	1.64
K_2O	0.08	0.44	0.78	0.08
P_2O_5	0.21	—	0.02	0.20
	100.04	100.00	100.00	100.00
$\frac{100\ Mg}{Mg + Fe^{++}}$ (atomic ratio)	72.0	68.1	59.8	79.4
CIPW Norms				
Or	0.6	2.7	4.5	0.5
Ab	14.7	18.9	18.0	13.9
Ne	—	—	2.2	—
An	27.6	27.3	26.2	26.3
Di	17.0	17.6	15.7	16.5
Hy	12.3	1.3	—	2.8
Ol	21.9	27.2	25.8	34.6
Ilm	3.8	4.4	4.8	3.7
Mt	1.4	0.6	2.9	1.3
Ap	0.5	—	—	0.4
"*Eclogite*" *Norm* — assuming TiO_2 in ilmenite, small "K_2O" solubility in acmite-omphacite. Garnet is $Ca_{0.5}$ (Mg, Fe)$_{2.5}$ $Al_2Si_3O_{12}$ in all compositions.				
Pyroxene	50	52	47	40
Garnet	46	44	46	43
Olivine	—	—	2	13
Quartz	tr	tr	—	—
Ilmenite	4	4	5	4

is appreciably more iron-rich (100 $Mg/Mg + Fe^{++} = 60$) than the olivine tholeiite composition (100 $Mg/Mg + Fe = 72$). Analyses of alkali olivine basalts and basanites containing peridotite inclusions with the moderately high pressure mineralogy of olivine + aluminous pyroxenes + spinel show that these have 100 $Mg/Mg + Fe^{++}$ ratios between 60 and 70. These basalts, because of their transported xenolithic material, may be assumed to be derived, without fractionation of olivine, from depths greater than 20 km (BULTITUDE and GREEN, unpublished). Such basalts also have high normative olivine contents ($>15\%$). It is considered that the chosen alkali olivine basalt composition is consistent with the available geological evidence on the estimation of compositions of critically undersaturated basaltic liquids derived without fractionation from depths of 20 km or more.

The olivine basalt composition was calculated as the arithmetic mean of the olivine and alkali olivine basalt compositions. As was previously discussed, it was also chosen to lie along the fractionation trend of the olivine tholeiite between 12 and 18 kb.

The compositions and CIPW norms of the glasses as used in the experiments are given in Table 1. Also given are "eclogite norms" based on calculation of garnets of composition $Ca_{0.5}(Mg, Fe)_{2.5}Al_2Si_3O_{12}$ and assuming solid solution of the minor orthopyroxene in some compositions in the calcium-rich pyroxene. The "eclogite norms" contain olivine in all four compositions if TiO_2 is calculated as rutile but only the picrite contains significant olivine if TiO_2 is calculated as ilmenite. Both the olivine tholeiite and olivine basalt lie essentially on the garnet + pyroxene join with neither quartz nor olivine, but the alkali olivine basalt contains 2% olivine and the picrite contains 13% olivine in eclogite nineralogy. All three basaltic compositions lie close to, but on the olivine-poor side of, the suggested minimum melting composition at 40 kb, using the simple Di-Fo-Pyrope system as a guide.

Experimental Procedure

a) Preparation of Basaltic Glasses

The glasses of olivine tholeiite, tholeiitic picrite and alkali olivine basalt were prepared from natural analyzed Hawaiian basalts with added MgO and $Fe + Fe_2O_3$ to obtain the desired compositions. The olivine tholeiite and tholeiitic picrite were prepared from a quartz tholeiite of the Pololu Series, Hawaii, and, as a result, have the low K_2O and Na_2O contents typical of these particular flows. After adding the MgO, Fe and Fe_2O_3 in calculated proportions, the components were intimately mixed and then fused to a glass. The alkali olivine basalt was similarly prepared from an analyzed Hawaiian alkali olivine basalt. The olivine basalt composition was prepared entirely by reacting mixtures of A. R. chemicals, in the required proportions, then adding Fe in the required amount and fusing the mixture to a glass.

Fusion of the basaltic compositions was carried out in either platinum tubes or crucibles at temperatures of 1,400—1,500° C. Oxidation was prevented by fusing in an argon atmosphere in contact with graphite at about 1,500° C. Loss of iron due to solution in the platinum containers during fusion was monitored in several control experiments and calculated amounts of Fe were added to compensate for this effect. After checking for homogeneity by examination in refractive index liquids, the glasses were finely ground and then analyzed in duplicate for FeO and Fe_2O_3 content. The compositions were calculated to 100% using the figures for FeO and Fe_2O_3 determined by analysis. The compositions obtained are listed in Table 1. In all cases the initial material was entirely glass — even the picrite was quenched without the formation of quench olivine.

b) Apparatus and Methods

The experimental runs were carried out using a single-stage, piston-cylinder apparatus similar to that described by BOYD and ENGLAND (1960a, b, 1963). The techniques used have been described previously (GREEN and RINGWOOD, 1967). In the melting runs reported in this paper, exclusion of water from the sample is essential. This was attained by enclosing the sample in a platinum tube of 0.006″ wall thickness with both ends sealed by welding. In addition the furnace assembly used contained an inner boron nitride sleeve between the talc and graphite cylinders and the boron nitride, graphite and all ceramic components were dried for 1 hour at 1,000° C in an argon atmosphere. To minimise the loss of iron to the platinum capsule, the periods of runs were limited to 1 hour at temperatures less than 1,400° C and to shorter times at temperatures greater than 1,400° C. Temperatures were measured by a Pt-Pt10%Rh thermocouple in contact with the charge. Variation of recorded temperature during a run was smaller than $\pm 5°$ C whilst the temperature gradient across the sample was smaller than 10° C. No correction was made for the effect of pressure on the thermocouple EMF. Ignoring this latter factor, the maximum error in determination of charge temperature was 15° C. In practice however, the relative errors in series of runs were much smaller than this. Thus, experimental runs spaced 5° apart and carried out with identical procedures were internally consistent and yielded reproducible results. Sample pressures were calculated from the load pressure on the 3″ ram, assuming a -10% pressure correction for friction and imperfect pressure transmission in the furnace assembly (GREEN, RINGWOOD and MAJOR, 1966). The absolute accuracy of pressure measurement is estimated at $\pm 3\%$.

c) Examination of Sample

Small portions of the sample were crushed and examined optically in refractive index liquids. This usually served to identify the phases present and the degree of melting. X-ray powder diffraction methods were used particularly to confirm the identification of olivine in the presence of clinopyroxene. Provided there was no access of water to the charge, little difficulty was found in quenching liquids to glass at pressures of 18 kb or lower. In some runs, particularly at 27 kb, there was patchy development of quench clinopyroxene (in part nucleating on primary crystals) but this was usually readily distinguishable from primary crystals by its feathery or fine-aggregate texture, by uneven extinction and by higher refractive indices than any primary clinopyroxene present.

Several fragments of most of the partial melting runs were mounted in epoxy resin and polished for use with the electron microprobe. These polished surfaces were excellent in showing the form and distribution of the primary phases, the quench clinopyroxene and the glass. Slight differences in hardness and in reflectivity were adequate to enable the phases to be clearly distinguished in reflected light.

d) Microprobe Methods and Calibration

An electron probe X-ray micro-analyser (Applied Research Laboratories model E.M.X.) was used for quantitative analysis for selected elements in olivine, enstatite, clinopyroxene and garnet in many of the experimental runs, particularly those near the liquidus. The electron beam of the microprobe was focussed to about 1 micron diameter, giving a volume of analysis (i.e. the volume from which characteristic X-rays were detectable) of 2—3 microns diameter and similar depth. An accelerating voltage of 12 kV, emission current of 60—100 mA and a specimen current of 0.04 micro-amps were used. Spot analyses were made at intervals of 1 to 10 microns with an integration time of 70 seconds. Analyses for Fe, Ca and Al were made simultaneously using K_α radiation. Corrections were made for specimen current fluctuations and for background but inter-element and matrix absorption and fluorescence effects were avoided by using calibration curves from glasses of known pyroxene composition. These glasses, prepared specifically for this project from carefully weighed A. R. chemicals, are enstatite ($100\ Mg/Mg + Fe = 90$) with 2.0% CaO to which were added Al_2O_3 contents from 0—20% and a second series of aluminous hypersthene ($100\ Mg/Mg + Fe = 75$; $Al_2O_3 = 8.0\%$) with variable CaO content from 0—8%. After fusion at 1,500—1,600° C and quenching to glass, these compositions were checked for iron content by analysis. The calibration curve for Fe was also based on analyzed olivines. Calibration curves for higher CaO contents in clinopyroxenes were based on a series of glasses of variable SiO_2, Al_2O_3, TiO_2, CaO, MgO, Na_2O and K_2O contents prepared by T. H. GREEN. During analysis of crystals in the high pressure runs, frequent re-analysis of several standards provided a means of monitoring machine drift. The determinations of Fe, Ca and Al (reported as oxides in accordance with normal petrographic practice) are considered accurate to ± 0.2 weight percent on large homogeneous samples. The limit of detection for Ca and Al with the methods used was 0.1% Ca and 0.2% Al_2O_3 respectively. These estimates of accuracy and limits of detection are derived from the reproducibility of results on standards and from the linearity of the calibration curves. Slightly greater uncertainties derive from the nature of the crystals in the high pressure runs — these will be discussed in later sections.

Since the analyses of crystals are only partial analyses, the compositions of pyroxenes, olivines and garnets must be calculated assuming stoichiometry of the crystals and considering the major solid solution end members only i.e. Ca, Mg, Fe silicates or aluminosilicates. Components such as Mn or Ni silicates were neglected. In the clinopyroxenes, TiO_2 and Na_2O may be present in minor amounts but are not considered in the present study. Olivine was calculated as forsterite, fayalite and larnite solid solution, orthopyroxene and clinopyroxene as enstatite, ferrosilite, wollastonite and Al_2O_3 solid solution, and garnet as grossular, almandine and pyrope solid solution. The latter calculation gives a calculated Al_2O_3 content which compares quite closely with the measured Al_2O_3, particularly since Fe^{+++} is assumed to be zero.

e) *Changes in Fe Content and Oxidation State*

The use of platinum capsules for the experimental runs means that loss of Fe to solid solution in the platinum and resulting oxidation of remaining FeO, may be of sufficient importance to materially change the chemistry and thus the mineralogy of the basalts. To evaluate this effect, semi-micro chemical analyses for FeO and Fe_2O_3 were carried out on 2—8 mg samples from some of the experimental runs. The results of these analyses are given in Table 2. The loss of iron is greater than from the subsolidus runs previously reported (GREEN and RINGWOOD, 1967) and averages about 20—25% of the total iron present. In most cases there is slight reduction of the iron rather than oxidation and this may be attributed to the presence of hydrogen and a reducing environment within the graphite furnace sleeve.

The changes in iron content and oxidation state are not so large as to destroy the basaltic chemistry or even the general character of the normative mineralogy. The normative olivine and hypersthene or nepheline contents of the most altered compositions are calculated in Table 2. The olivine contents in the tholeiitic basalts are considerably decreased and hypersthene is conversely increased. In the alkali olivine basalt compositions both olivine and nepheline decrease but the compositions remain nepheline normative.

The chemical composition of the sample cannot be considered constant for the length of the experimental run and it is uncertain whether observed crystals in glass maintained equilibrium with the changing bulk chemistry throughout the length of the run. Microprobe examination of samples has demonstrated that there may be a range in $Mg/Mg+Fe$ contents of crystals, controlled by their position in the sample capsule. Crystals which have formed near the edge of the sample are more magnesian than those at the centre — this has been particularly well shown for olivine. In addition some orthopyroxene, olivine and garnet crystals have shown zoning in Fe content — this zoning is the reverse of normal igneous zoning and crystal cores are more iron-rich than crystal edges. This effect is attributed to gradual loss of Fe from the sample with the result that outer zones of crystals either maintain equilibrium with the changing liquid composition while crystal cores do not, or alternatively the outer zones of crystals have grown during the latter part of the experimental runs. Either process would produce the observed zoning of crystals.

A method of evaluating the effect of Fe-loss in relation to the compositions of crystals formed during the experimental runs is to compare the partition coefficients for liquidus crystal-magma equilibria observed in natural basalts with the partition coefficients observed in the experiments. In considering the partition of Mg and Fe between liquidus olivine and liquid, it is convenient to define a partition co-efficient

$$K_{O,L} = \left(\frac{Mg}{Mg + Fe^{++}}\right)_{\text{Olivine}} \Bigg/ \left(\frac{Mg}{Mg + Fe^{++}}\right)_{\text{Liquid}} .$$

d change in oxidation state during experimental runs

	Conditions			FeO		Fe_2O_3	
	Temp. (°C)	Press. (kb)	Time (hrs)	initial	final	initial	final
ite A	1200	9	2	10.07	8.31	1.02	0.21
ite A	1250	9	1	10.07	7.10	1.02	0.94
ite A	1370	9	1	10.07	6.92*	1.02	2.61*
ite A	1375	9	1	10.07	6.42	1.02	1.60
ite A	1440	22.5	$^1/_2$	10.07	8.33	1.02	0.88
ite A	1450	22.5	20 min	10.07	8.27	1.02	1.40
ite A	1510	27	20 min	10.07	8.07	1.02	1.11
runs	—	—	—	10.07	7.63+	1.02	1.25+
	1250	18	1	12.42	10.00	1.87	1.14
	1400	18	1	12.42	8.23**	1.87	3.90**
	1440	27	$^1/_2$	12.42	7.60	1.87	0.29
Mix 2)	1310	13.5	$^1/_2$	13.06	10.25	0.17	0.70
	1360	18	1	10.64	7.09++	0.42	1.15++
	1300	13.5	50 min	10.64	8.72	0.42	0.82
	1520	18	$^1/_2$	8.67	6.48	0.92	0.47

ompositions after experiments	*Initial compositions*
7.1% Hypersthene 13.6% Olivine	(12.3% Hy, 21.9% Ol)
4.6% Hypersthene 16.9% Olivine	(12.3% Hy, 21.9% Ol)
0.2% Nephelene 19.5% Olivine	(2.2% Ne, 25.8% Ol)
8.1% Hypersthene 17.5% Olivine	(1.3% Hy, 27.2% Ol)

From the experimental results (see pages 132—134) on the three basaltic compositions at 4.5 and 9 kb, the partition coefficients can be calculated assuming either that the "liquid" composition is that of the original mix before Fe-loss during the experiment or that the liquid composition is close to the average of the compositions analysed for FeO, Fe_2O_3 after the melting experiments (Tables 2 and 3). The examples of partition coefficients in natural basalts are the maximum values obtained and only include those with olivine determined by chemical, X-ray or refractive index measurements. The effect of olivine accumulation or of the olivine phenocrysts present representing crystal-liquid equilibrium at temperatures lower than the liquidus, will be to yield *low* values for the natural partition coefficients.

Table 3. *Partition coefficients between olivine and liquid composition in experimental and natural examples*

	Olivine in equilibrium with initial mix composition			Olivine in equilibrium with average composition after Fe loss	
	$\left(\frac{100\ Mg}{Mg+Fe}\right)_{Ol}$	$\left(\frac{100\ Mg}{Mg+Fe^{++}}\right)_{Liq}$	$K_{O,L}$	$\left(\frac{100\ Mg}{Mg+Fe^{++}}\right)_{Liq}$	$K_{O,L}$
Olivine Tholeiite	92	72	1.28	78	1.18
Olivine Basalt	90	68	1.32	73	1.23
Alkali Olivine Basalt	83	60	1.38	69	1.20
		Mean Value	1.33		

	Natural examples of maximum partition coefficients between phenocryst olivine and whole rock composition		
	$\left(\frac{100\ Mg}{Mg+Fe}\right)_{Ol}$	$\left(\frac{100\ Mg}{Mg+Fe^{++}}\right)_{Liq}$	$K_{O,L}$
1959 Kilauea 1st phase (MUIR and TILLEY, 1963)	89	66	1.35
1959 Kilauea Pele's hair (MACDONALD and KATSURA, 1961)	87.5	62.5	1.40
1887 Mauna Loa (YODER and TILLEY. 1962)	87	62.5	1.39
1868 Kilauea flow (MUIR and TILLEY 1963)	81	62	1.31
1868 Kilauea flow (MUIR and TILLEY. 1963)	88	59.5	1.48
1868 Kilauea tachylyte (MUIR and TILLEY, 1963)	81.5	53.5	1.52

In both the experimental and natural examples the partition coefficient apparently increases with decreasing values of 100 Mg/Mg $+ Fe^{++}$ for the liquid (Table 3). However within the range $60 < 100$ Mg/Mg $+ Fe^{++} < 75$ there is close similarity between the values of the natural partition coefficients and the partition coefficients for the experimental runs provided that the initial, pre-run, composition is used as the liquid composition. The comparison of partition coefficients thus suggests that the iron loss from the sample occurs after most of the crystal growth, at least in the central parts of the sample capsule.

In summary, the evidence we have obtained on the magnitude and time dependence of the loss of Fe to the platinum capsule, leads to the conclusion that this effect is not a major factor in the experiments reported herein in modifying either the major element chemical composition or the nature of the crystalline phases of the experimental runs.

Experimental Results

Detailed descriptions of the results of individual runs on the several compositions are given in Tables 4—7. Photographs of typical phases in polished sections are *ental partial melting runs on olivine tholeiite composition*

Time (mins)	Phases present*	Comments
30	Ol + Px + Pl + Glass	Low degree ($\sim$30%) of melting. Orthopyroxene r
30	Ol + Pl + ?Mt + Glass	Glass > crystals. Plagioclase approx. 10—15%. N Minor brown spinel (?magnetite_{ss}) possibly due t
30	Ol + Pl + Glass	Plagioclase definite but very weak reflections.
30	Ol + ?Mt + Glass	Common olivine ($\sim$10%), minor brown spinel (? oxidation (capsule burst during run).
20	Ol + ?Mt + Glass	Uncommon olivine, rare ?magnetite_{ss}. Near liqui
20	Ol + Glass	Very, very rare olivine, essentially liquidus run.
60	Ol + Glass	Olivine as very uncommon small euhedra. Very n
180	Ol + Px + Pl + Spinel	Plagioclase common. Olivine more common tha Spinel as moderately common small green to g Orthopyroxene not identifiable optically.
180	Ol + Opx + Cpx + Pl + Glass	Minor segregated glass, also intergranular, with small anhedra. Opx as large laths.
180	Ol + Opx + Cpx + Glass	Coarser grained and with more glass than follow fication (X-ray) uncertain.
60	Ol + Opx + Cpx + Glass	Near solidus run with fine Cpx and Ol, larger Op No definite plagioclase (X-ray).
60	Ol + Opx + Cpx + Glass	Glass moderately common but crystals > glass. large euhedra, olivine as smaller euhedra, clinop hedral or euhedral small crystals.
60	Ol + Opx + Glass	Much greater partial melting than in previous run. No definite Cpx.
60	Ol + Opx + Glass	Olivine common. Orthopyroxene uncommon. No
60	Ol + Glass	Olivine common.
60	Ol + Glass	Olivine moderately common, euhedral.
60	Ol + Glass	Olivine as uncommon, euhedral crystals. Very ne
60	Glass + quench cpx	Rare quench clinopyroxene.
60	Glass + quench cpx	Rare quench clinopyroxene.

84	10.1	1,100	120	Ol + Px + Pl + Spinel	Similar to 9 kb, 1,100°C run, olivine reflections weak.
488	11.3	1,100	150	Px + Pl + Spinel	Olivine absent, plagioclase much decreased from 10.1 kb run and spinel more abundant.
762	11.3	1,370	40	Ol + Opx + Glass	Moderately common large orthopyroxene laths, less common small olivine euhedra. Comparable crystallization to 13.5 kb, 1380°C run.
487	12.4	1,100	180	Px + Pl + Spinel	Plagioclase decreased from run 488. No definite garnet. Spinel moderately common.
469	13.5	1,100	180	Px + Pl + Ga + Amphibole	Minor plagioclase reflections, weak garnet reflections but optically distinctive as porphyroblasts. Spinel minor or absent. Minor amphibole.
461	13.5	1,180	180	Opx + Cpx + Pl + Spinel	Predominantly clinopyroxene + orthopyroxene. Plagioclase definitely present but less abundant than at 13.5 kb, 1,100°C. Spinel moderately common. Orthopyroxene is very distinctive as large porphyroblasts with both spinel and plagioclase inclusions. Clinopyroxene, in contrast, occurs as small equant granules. No glass.
456	13.5	1,220	180	Opx + Cpx + ? Pl + Spinel + Glass	Orthopyroxene and clinopyroxene similar to 13.5 kb, 1,180°C run. Opx porphyroblasts contain included spinel and rare included plagioclase. Glass as intergranular low R.I. continuum, more abundant than discrete low R.I. plagioclase grains at 13.5 kb, 1180°C. Plagioclase identification probable from X-ray data.
441	13.5	1,220	60	Opx + Cpx + ? Pl + Spinel + Glass	Similar to previous 3hr run but with less low R.I. material. Opx porphyroblasts smaller than in 3 hr run. Plagioclase probably present. Very near solidus.
782	13.5	1,250	60	Opx + Cpx + Pl + Spinel	Mainly very fine-grained clinopyroxene with uncommon small euhedral orthopyroxene porphyroblasts. Minor spinel. Moderately common disconnected amoeboid patches of low R.I. plagioclase. No definite glass. Solidus run. Plag. approx. 15%.
438	13.5	1,270	60	Opx + Cpx + Spinel + Glass	Crystals > glass. Mainly fine, subhedral clinopyroxene with uncommon orthopyroxene as larger crystals. Intergranular continuum and some segregations of glass. Minor pale green spinel.
432	13.5	1,290	60	Opx + Cpx + Spinel + Glass	Very similar to 1,270°C run. Glass more common. Minor spinel identifiable on polished surface.
425	13.5	1,310	60	Opx + Cpx + Glass	Common small equant clinopyroxene, less common large orthopyroxene laths. Crystals > glass.

Table 4 (continued)

Run No.	Pressure (kb)	Temp. (°C)	Time (mins)	Phases present*	Comments
428	13.5	1,350	60	Opx + Cpx + Glass + quench cpx	Glass > crystals. Opx > Cpx. Large euhedral orthopyroxenes and very uncommon large clinopyroxene crystals. Clinopyroxene clearly primary and in part intergrown with othopyroxene.
424	13.5	1,380	60	Opx + Glass + quench cpx	Common (10—15%) (Plate IB) euhedral orthopyroxene. Rare quench clinopyroxene, in part nucleating on othopyroxene margins.
143	13.5	1,400	60	Opx + Glass	Uncommon large orthopyroxene crystals in glass. Some crystal settling possible. Near-liquidus run.
773	13.5	1,360	5	Cpx + ? Spinel + Glass + quench cpx	Poorly crystallized fine granular clinopyroxene, minor probable spinel and intergranular glass.
772	13.5	1,380	5	Opx + Cpx + Glass + quench cpx	Mainly poorly crystallized granular and feathery clinopyroxene. Rare orthopyroxene cores.
781	13.5	1,400	5	Opx + ? Cpx + Glass + quench cpx	Mainly feathery, granular and some rather clear clinopyroxene crystals + glass. Uncommon but distinctive orthopyroxene commonly with clinopyroxene rims or patchy "intergrowth" of clinopyroxene.
774	13.5	1,420	5	Glass + quench cpx	Glass with rare, feathery clinopyroxene.
761	13.5	1,380	40	Opx + Glass + quench cpx + residual Ga	The initial material for this run was crystallized to garnet + clinopyroxene at 30 kb, 1,100° C, 4 hrs. The run consisted of abundant glass, moderately common clear, medium to large orthopyroxene laths with borders and outgrowths of feathery quench clinopyroxene. No primary clinopyroxene. Some glass areas contain small, perfectly spherical, garnet residual crystals, some with rutile inclusions.
113	18.0	1,300	60	Cpx + Ga + ? Glass	Very near to solidus, small percentage of low R.I., amoeboid grains may be glass or plagioclase.
131	18.0	1,325	60	Cpx + Ga + Glass	Garnet less than in previous run. Glass intergranular and in small segregations.
125	18.0	1,350	60	Cpx + Ga + Glass	Minor garnet (euhedral), mainly fine clinopyroxene with intergranular glass. Crystals > glass.

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121	18.0	1,370	60	Cpx + Glass	Garnet absent, orthopyroxene not identifiable. Fine-grained clinopyroxene + common glass (~40%).
122	18.0	1,375	60	Opx + Cpx + Glass	Well crystallized orthopyroxene and clinopyroxene crystals, some in parallel intergrowth. Abundant glass, and some possible quench clinopyroxene. Cpx > Opx.
118	18.0	1,380	60	Opx + Cpx + Glass + quench cpx	Large primary orthopyroxene and clinopyroxene as both separate and intergrown crystals. Cpx > Opx. Common quench clinopyroxene.
114	18.0	1,400	60	Opx + Cpx + Glass + quench cpx	Large, clear, primary orthopyroxene and clinopyroxene set in glass or glass + feathery quench clinopyroxene. Opx > Cpx.
454	18.0	1,420	30	Opx + Glass + quench cpx	Rare, large euhedral crystals of orthopyroxene in glass + minor quench clinopyroxene. Very near liquidus.
123	18.0	1,425	60	Opx + Glass + quench cpx	Rare, large euhedral crystals of orthopyroxene in glass + minor quench clinopyroxene. Very near liquidus.
448	22.5	1,410	20	Cpx + Ga + Glass	Common large euhedral garnet. Major phase is rather fine-grained, subhedral clinopyroxene. Minor glass generally intergranular, rarely as small segregations.
449	22.5	1,430	20	Cpx + Glass	Abundant fine-grained subhedral clinopyroxene in glass. Crystals > glass.
451	22.5	1,440	30	Cpx + Glass + quench cpx	Glass + common quench clinopyroxene. Some clinopyroxene is probably primary but difficult to distinguish from quench.
447	22.5	1,450	20	Glass + quench cpx	Mainly glass with patches of quench clinopyroxene.
783	27.0	1,430	30	Cpx + Ga	Medium-sized subhedral garnet and fine grained clinopyroxene. Glass absent or as very rare and very small patches. Solidus run.
442	27.0	1,450	20	Cpx + Ga + Glass	Common euhedral garnet and fine-grained clinopyroxene. Minor intergranular glass. Near-solidus run.
444	27.0	1,490	20	?Cpx + Ga + Glass + quench cpx	Moderately common euhedral garnet. Very common quench clinopyroxene, some clinopyroxene *may* be primary.
445	27.0	1,510	20	Glass + quench cpx	Mainly glass but some clinopyroxene considered to be of quench origin. Above liquidus.

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* Abbreviations used are as follows: Ol — olivine, Opx — orthopyroxene, Cpx — clinopyroxene, Ga — garnet, Pl — plagioclase.

Table 5. *Details of partial melting experiments on olivine basalt composition*

Run No.	Pressure (kb)	Temp. (° C	Time (mins)	Phases present	Comments
784	9.0	1,240	60	Ol + Cpx + Glass	Abundant very fine anhedral clinopyroxene. Minor olivine (slightly larger, subhedral). About 40% glass.
787	9.0	1,260	60	Ol + Opx + Cpx + Glass	Abundant small anhedral clinopyroxene. Moderately common olivine, rare orthopyroxene laths. About 60% glass.
418	9.0	1,280	60	Ol + Glass	Uncommon, small euhedral olivine in glass.
764	13.5	1,290	50	Opx + Cpx + Spinel + Glass	Rare orthopyroxene, abundant fine clinopyroxene, probable minor spinel with intergranular glass locally expanding into small segregations. Estimated glass (30%).
767	13.5	1,310	50	Opx + Cpx + Glass + quench cpx	Large orthopyroxene laths and some large clinopyroxene crystals. Opx > Cpx. Quench clinopyroxene common.
419	13.5	1,320	60	Opx + ? Cpx + Glass	Large orthopyroxene laths with rims and some parallel growth of clinopyroxene, in glass. No definite primary clinopyroxene and none identified with microprobe (Plate I c).
770	13.5	1,330	50	Opx + Glass + quench cpx	Uncommon, well formed orthopyroxene, some quench clinopyroxene but no definite primary clinopyroxene.
397	18.0	1,320	60	Cpx + Ga + Glass	Moderately common large, subhedral garnet with abundant fine grained clinopyroxene. Glass intergranular with some small segregations.
400	18.0	1,330	60	Cpx + Ga + Glass	Uncommon garnet, small subhedral clinopyroxene with some outgrowth of quench cpx. No orthopyroxene identifiable. Less garnet and more glass than previous run. Crystals > Glass.
405	18.0	1,335	60	Opx + Cpx + Glass + quench cpx	Large orthopyroxene laths and large clinopyroxene crystals, in some cases in parallel growth, also as distinct crystals. Common fine, anhedral and feathery quench clinopyroxene. Glass > Primary crystals.
398	18.0	1,340	60	Opx + Glass + quench cpx	Rare large orthopyroxene crystals, rimmed by feathery clinopyroxene but this is not in parallel growth as in previous run. Very near liquidus.
402	18.0	1,340	60	Opx + Glass + quench cpx	Large orthopyroxene crystals, slightly more common than in previous run, in fine-grained equant and feathery clinopyroxene and glass. No clinopyroxene in parallel growth with orthopyroxene.
396	18.0	1,360	60	Glass + quench cpx	Glass with patchy quench clinopyroxene including both feathery and anhedral forms.

404	27.0	1,420	60	Cpx + Ga + Glass	Abundant garnet and clinopyroxene, in similar proportions. Intergranular glass — approx. 20—30% melting.
407	27.0	1,450	60	Ga + Glass + quench cpx	Garnet as medium-sized and large euhedral crystals either in feathery, turbid quench clinopyroxene or in clear glass (Fig. 2D). No definite primary clinopyroxene.
415	27.0	1,460	20	Ga + Glass + quench cpx	Similar to previous run but garnet smaller and less abundant. Patchy development of quench clinopyroxene and other areas of glass + garnet. No evidence for primary clinopyroxene.
410	27.0	1,470	60	Ga + Glass	Rare garnet is spherical or amoeboid showing clear resolution effects. Glass is colourless and has grossly lost Fe to the platinum capsule. Garnet interpreted as early crystallizing liquidus phase, later being redissolved in the liquid as the composition changed due to Fe loss.

given in Fig. 2. A number of generalizations on the habits and appearance of the various crystalline phases apply to all the compositions. Olivine, when appearing on or near the liquidus, occurs as euhedra of short prismatic or pseudohexagonal habit (crystal size 5—25 microns) — the latter form, with strong development of the hkl or h0l, 0kl faces, is particularly diagnostic. Prismatic crystals show straight extinction and moderate to high birefringence, pseudohexagonal or rhombic sections show symmetrical extinction. When present in assemblages near the solidus with abundant clinopyroxene, olivine is distinguished from the latter by more euhedral and slightly larger crystals and by higher birefringence.

Orthopyroxene is very distinctive in the experimental runs with a greater tendency towards euhedral form and large, inclusion-free crystals than any other mineral present. In near-liquidus runs the orthopyroxene crystallizes as elongate prisms terminated by the basal pinacoid, in some cases with minor development of the hkl, 0kl or h0l faces. Crystals are commonly 40—400 microns long and 10—50 microns thick (Fig. 2B, C). The orthopyroxene is optically distinctive in having straight extinction with low birefringence in prismatic sections. In runs well below the solidus (1,100° C) the orthopyroxene remains fine-grained and cannot be distinguished optically from clinopyroxene. In runs close to the solidus, and particularly in the runs at 9 kb and 13.5 kb on the olivine tholeiite composition where there has been access of small amounts of water in 3-hour runs, the orthopyroxene grows as very distinctive, large porphyroblasts which contain uncommon small inclusions of spinel, clinopyroxene and plagioclase. In some near-liquidus runs in which clinopyroxene is a co-existing primary phase as uncommon large crystals, the orthopyroxene shows included and marginal parallel intergrowth of clinopyroxene. In runs in which orthopyroxene and liquid are the only phases, the orthopyroxene does not show this intergrowth with clinopyroxene but may show some nucleation of quench clinopyroxene as feathery outgrowths, particularly at crystal corners (Fig. 2C).

Primary clinopyroxene occurs typically as small (2—10 microns) equant or short, prismatic crystals

Table 6. *Details of partial melting experiments on alkali olivine basalt composition*

Run No.	Pressure (kb)	Temp. (°C)	Time (mins)	Phases present	Comments
A3	1 atmos.	1,220	30	Ol + Cpx + Pl + ? Mt + Glass	About 50% melting. Common olivine, plagioclase and clinopyroxene, minor deep brown spinel_{ss}.
A5	1 atmos.	1,240	30	Ol + Pl + Glass	About 30% crystals, olivine and plagioclase in similar abundance. No semi-opaque spinel_{ss}.
A10	1 atmos.	1,250	30	Ol + Glass	Uncommon olivine, very near liquidus.
A9	1 atmos.	1,260	30	? Mt + Glass	Minor brown spinel (magnetite_{ss}) possibly due to some oxidation.
788	9.0	1,180	60	Ol + Cpx + Pl + Spinel + Glass	Near solidus run. Olivine (minor) with abundant fine equant clinopyroxene, uncommon spinel and common low R.I. plagioclase + some glass (probably) intergranular and amoeboid patches.
401	9.0	1,220	60	Ol + Cpx + Glass	Olivine and clinopyroxene as small euhedral or subhedral crystals. Clinopyroxene identified by microprobe. Glass > Crystals.
786	9.0	1,240	60	Ol + Glass	Olivine crystals in glass. Clinopyroxene not identifiable optically nor by X-ray means.
399	9.0	1,260	60	Ol + Glass	Small euhedral olivine in glass. Rare quench clinopyroxene.
785	11.3	1,270	60	? Ol + Cpx + Opx + Glass	Moderately common small euhedral to subhedral crystals of clinopyroxene, probable minor olivine and rare orthopyroxene laths. Glass $\gg$ crystals.
523	13.5	1,250	60	Cpx + Glass	Abundant small anhedral clinopyroxenes in intergranular glass. No definite spinel. Crystals > glass.
521	13.5	1,270	45	Cpx + Glass	Medium grained anhedral or subhedral clinopyroxene in glass. No orthopyroxene. Glass > crystals.
420	13.5	1,290	60	Opx + Cpx + Glass + quench cpx	Rare large orthopyroxene and clinopyroxene primary crystals in glass and quench clinopyroxene in feathery and aggregate textures.
494	13.5	1,290	30	? Opx + Cpx + Glass + ? quench cpx	Small subhedral clinopyroxene and possibly some orthopyroxene with glass and minor ?quench clinopyroxene outgrowths and anhedral aggregates. Apparently much greater crystallization than previous run.
501	13.5	1,295	30	Glass + quench cpx	Patchy development of fine aggregates of anhedral clinopyroxene in glass, probably entirely quench but may be primary.
497	13.5	1,300	30	? Opx + ? Cpx + Glass + quench cpx	Small, rectangular crystals of probable orthopyroxene with clinopyroxene outgrowths and similar crystals of clinopyroxene with quench outgrowths. Abundant glass and anhedral quench clinopyroxene aggregates.

0	30	Glass + quench cpx	No primary crystals. Very little quench clinopy
0	60	Cpx + Ga + Glass	Mainly fine clinopyroxene with moderately co hedral crystals. Minor (~10%) intergranular gl
0	60	Cpx + Glass	Abundant granular clinopyroxene with intergra coloured glass. About 50% glass.
0	60	Cpx + Glass	Uncommon large clinopyroxene crystals in glass
5	60	Glass + quench cpx	Some patches of feathery and fine aggregate qu
0	60	Glass + quench cpx	Some patches of feathery and fine aggregate qu
0	60	Cpx + Ga + Glass	Abundant garnet and clinopyroxene with minor segregations showing brownish or mauve colour
0	60	Cpx + Ga + Glass	Euhedral garnet and sub-hedral, equant clinopy glass. Cpx > Ga.
0	30	Ga + Glass + quench cpx	Euhedral garnet, locally in clear glass, more c quench clinopyroxene.
0	30	Glass + quench cpx	No garnet. Patchy quench clinopyroxene.

and X-ray data are commonly necessary. The identification as primary or quench clinopyroxene is very difficult in some runs where the quench clinopyroxene forms fine-grained anhedral aggregates. More commonly the quench clinopyroxene is feathery, turbid, without crystal form and with wavy extinction. In some runs at 22.5 or 27 kb, larger crystals of clinopyroxene are distinguished as quench products by small inclusions of turbid, low R.I. material either in stringers or randomly distributed. Primary clinopyroxene crystals in these runs are clear and have lower refractive indices than enclosing quench clinopyroxene.

In the compositions studied in this paper, plagioclase appears only in runs at temperatures close to the solidus. Under these conditions plagioclase could not be distinguished from glass by optical means and was identified by X-ray diffraction methods.

In subsolidus or near-solidus runs, garnet forms euhedral or subhedral crystals usually from 10—20 microns grain-size and commonly containing included pyroxene or other phases. In near-liquidus runs garnet is euhedral, without inclusions and up to 50 microns grain-size (Fig. 2D).

Table 7. *Details of partial melting experiments on picrite composition*

Run No.	Pressure (kb)	Temp. (°C)	Time (mins)	Phases present	Comments
614	13.5	1,320	40	Ol + Opx + Cpx + Glass + quench cpx	Opx uncommon but as large phenocrysts.
623	13.5	1,340	40	Ol + Opx + Cpx + Glass	Glass approx. 40—50% of run. Opx as uncommon large porphyroblasts. Cpx > Ol.
572	13.5	1,360	30	Ol + Glass	Abundant euhedral olivines.
612	13.5	1,360	40	Ol + Opx + Glass	Common olivine, very rare orthopyroxene.
479	13.5	1,380	30	Ol + Glass	Euhedral olivine + glass.
618	18.0	1,350	40	Ol + Ga + Cpx + ? Glass	Near solidus run. Minor garnet. Minor olivine. Small amount of intergranular low R.I. phase may be glass.
462	18.0	1,370	30	Ol + Cpx + Glass	Crystals > glass. Small equant clinopyroxene with minor olivine.
465	18.0	1,390	30	Ol + Cpx + Glass + quench cpx	Euhedral olivine. Crystals < glass + quench cpx. No orthopyroxene.
476	18.0	1,400	20	Ol + Opx + ? Cpx + Glass + quench cpx	Common olivine. Uncommon orthopyroxene, some with parallel Cpx intergrowth. No definite separate primary Cpx.
467	18.0	1,410	30	Ol + Opx + ? Cpx + Glass + quench cpx	Common olivine. Uncommon orthopyroxene, some with parallel Cpx intergrowth. No definite separate primary Cpx.
468	18.0	1,430	30	Ol + Glass	Moderately common olivine euhedra.
620	22.5	1,430	20	Ol + Cpx + Ga + ? Spinel + Glass	Near solidus run. Mainly. Mainly Cpx + minor garnet + minor olivine + minor brownish high R.I. phase (? Spinel or Mg-rich ilmenite).
621	22.5	1,450	20	Cpx + Glass + quench cpx	Near liquidus run. Uncommon large clinopyroxene with quench outgrowths of Cpx.
615	22.5	1,460	20	Glass + quench cpx	Above liquidus.
625	27.0	1,200	120	Cpx + Ga + Ol + ? Magnetite-ulvospinel	Common garnet and clinopyroxene. Minor olivine. Minor semi-opaque (possibly magnetite-ulvospinel$_{ss}$).
619	27.0	1,450	20	Cpx + Ga + ? Spinel	Subsolidus run. Common garnet and clinopyroxene. Minor translucent, high R.I. grains may be spinel$_{ss}$.
473	27.0	1,480	20	Cpx + Ga + ? Spinel + Glass	Near solidus run. Cpx and Ga well crystallized. Possible minor olivine. Minor high R.I. phase (spinel$_{ss}$ or ilmenite$_{ss}$).
475	27.0	1,500	20	Opx + Cpx + Ga + Glass + quench cpx	Uncommon garnet euhedra, uncommon orthopyroxene laths. Common Cpx but most is probably quench clinopyroxene. Possible olivine, minor high R.I., birefringent rutile or Mg-ilmenite ?
616	27.0	1,510	20	Glass + quench cpx	Above liquidus.
474	27.0	1,520	20	Glass + quench cpx	

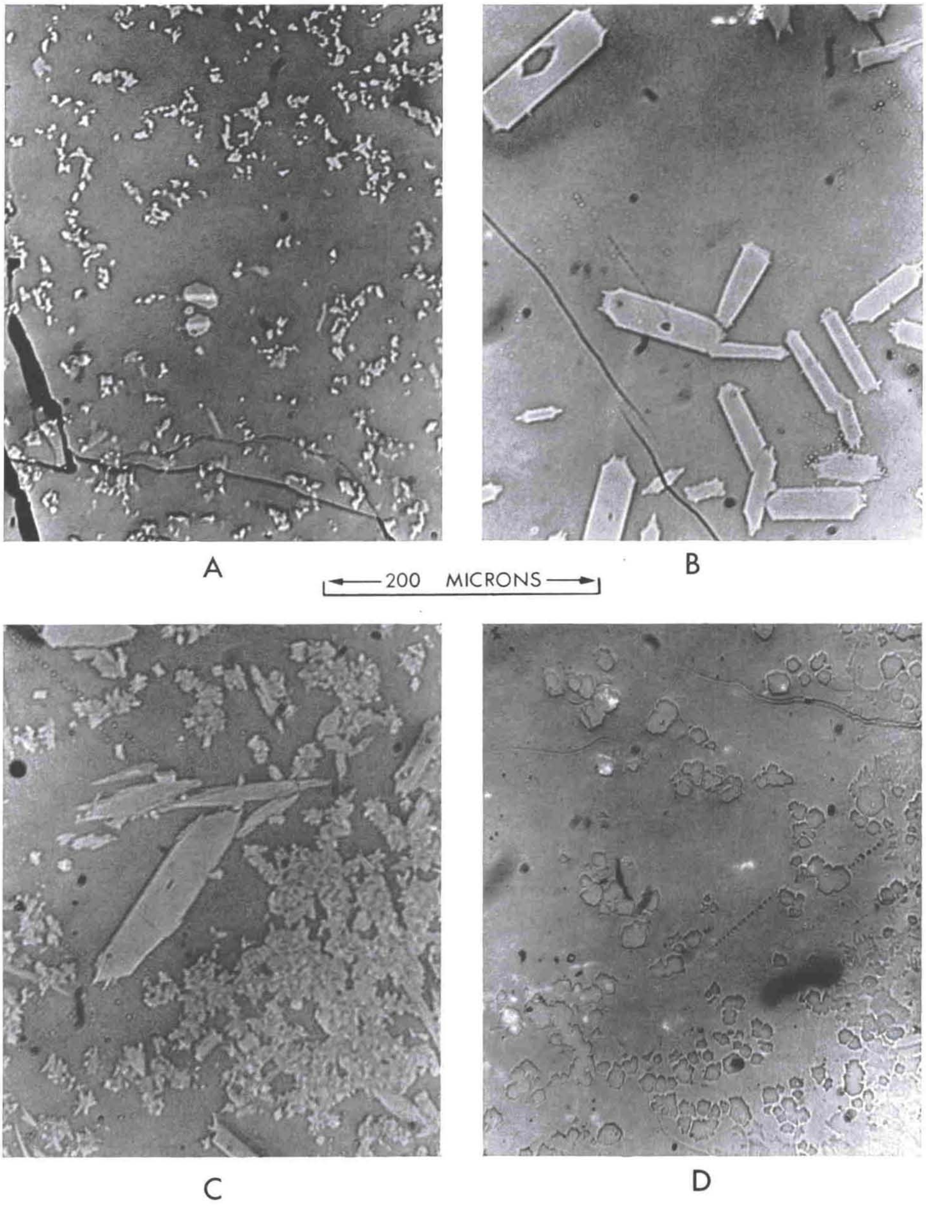

Fig. 2 A—D. Polished mounts for electron microprobe analyses showing crystalline phases in glass at various pressures. A. Olivine (small, highly reflecting grains) and orthopyroxene (larger, lath-shaped) crystals in glass. Olivine tholeiite at 9 kb, 1,290° C. White trace across larger orthopyroxenes shows path of electron beam during microprobe analyses. B. Orthopyroxene crystals in glass. Olivine tholeiite at 13.5 kb, 1380° C. Incipient nucleation of quench clinopyroxene particularly at corners of orthopyroxene. C. Large orthopyroxene and fine aggregates of quench clinopyroxene. Olivine basalt at 13.5 kb, 1,320° C. D. Garnet crystals and patchy aggregates of finegrained quench clinopyroxene (right hand side of picture) in 27 kb, 1,450° C run

either due to loss of iron to the platinum capsule or, in some cases, to splitting of the sample capsule during the run. The liquidus of the olivine tholeiite is at approximately 1,340° C and that of the alkali olivine basalt at about 1,260° C. In both compositions plagioclase is the second phase to appear (at 1,260° C in the olivine tholeiite and at 1,240° C in the alkali olivine basalt). Clinopyroxene appears at a similar temperature (1,220 ± 10° C) in both compositions and thus, in the olivine tholeiite composition there is a temperature interval of about 30° C over which precipitation of olivine and plagioclase together will determine the fractionation trend.

b) Crystallization at 9 kb

The liquidus phase in all three basalt compositions studied at 9 kb is olivine. In the olivine tholeiite composition olivine is the only phase crystallizing from the liquidus at 1,360 ± 10° C[3] down to 1,300° C. In the olivine basalt the liquidus was not accurately established but seems unlikely to be higher than 1,310° C and both orthopyroxene and clinopyroxene begin crystallizing with olivine between 1,260° C and 1,280° C. In the alkali olivine basalt composition, olivine occurs on the liquidus at 1,270 ± 10° C and co-exists with liquid alone down to 1,230 ± 10° C. At 9 kb, there is in all three compositions, a temperature interval of about 40° C or more in which separation of olivine alone will govern the early stages of fractionation of these basaltic liquids. It is of interest that the composition with the *lowest* normative olivine content (the olivine tholeiite) has the highest liquidus temperature at 9 kb and the largest temperature interval for the two phase olivine + liquid assemblage. This composition has the highest $Mg/Mg + Fe^{++}$ value and lowest $Na_2O + K_2O$ content and the liquidus temperatures show a dependence on alkali content and iron-enrichment closely analogous to that observed at atmospheric pressure by TILLEY, YODER and SCHAIRER (1965).

In the 1.290° C run on the olivine tholeiite composition, olivine is joined by orthopyroxene (Plate IA) and the assemblage ol + opx + liquid is stable down to 1,260 ± 10° C. At this temperature the composition is less than 50% crystallized. However, the appearance of clinopyroxene as a primary phase in the 1,250° C run also results in greatly increased degree of crystallization. There is an apparent decrease in the abundance of olivine in the 1,210 to 1,250° C runs in comparison with runs at both higher and lower temperatures. The increase in olivine in going from 1,210° C down to 1,100° C may be attributed to decreasing solubility of Al_2O_3 in pyroxenes with decreasing temperatures — it is accompanied by an increase in spinel and in plagioclase (c.f. GREEN and RINGWOOD, 1967) and implies reactions

$$\text{olivine} + \text{plagioclase} \rightleftharpoons \text{aluminous pyroxenes}_{SS}$$
$$\text{spinel} + \text{plagioclase} \rightleftharpoons \text{aluminous pyroxenes}_{SS}.$$

The lower abundance of olivine in the interval 1,210° C—1,250° C in comparison with runs at higher temperatures implies a reaction relationship between olivine and liquid — this will be further discussed in a later section, but it may be noted that the precipitating phase, replacing olivine, is not enstatite or hypersthene but aluminous clinopyroxene.

Plagioclase may be present as a minor phase at 1,230° C, 9 kb, and is definitely present in the near-solidus run at 1,210° C. The very late appearance and small content of plagioclase in this composition at 9 kb contrast with the major role that this phase plays in the atmospheric pressure runs.

Olivine is again the liquidus phase in the olivine basalt at 9 kb. Both orthopyroxene and clinopyroxene begin crystallizing at similar temperatures (about 1,260° C) but the orthopyroxene is a very minor phase. The degree of crystallization shows a marked increase with the appearance of clinopyroxene. At a lower temperature (1,240° C) clinopyroxene increases in abundance at the expense of both orthopyroxene (no longer observed) and olivine.

In the alkali olivine basalt composition, orthopyroxene does not appear at 9 kb and olivine is joined by clinopyroxene in the 1,220° C run. Clinopyroxene in this run forms medium sized euhedra (up to 10—15 microns), not distinguishable optically from olivine but clearly identifiable with the microprobe. In this composition the first appearance of clinopyroxene is apparently not accompanied by an abrupt increase in the degree of crystallization. At 1,180° C, a near-solidus run contains mainly clinopyroxene with a moderate amount of plagio-

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[3] The uncertainty in temperature reflects the spacing of runs at 10° C—20° C intervals and does not include the estimated precision or absolute accuracy of temperature measurement (see page 114).

clase, minor olivine and minor spinel. This assemblage differs from the subsolidus run at 1,100° C (Green and Ringwood, 1967) only in having lower plagioclase and olivine contents.

c) Crystallization at 11.3 kb

Only two runs, both aimed at establishing the phases present close to the liquidus, were carried out at 11.3 kb. In the olivine tholeiite composition, a run at 1,370° C showed large orthopyroxene euhedra and less common but well crystallized olivine euhedra. Either phase may be the liquidus phase (at $1380 \pm 10°$ C) but it is clear that both phases are important in the early stages of fractioning of the olivine tholeiite at this pressure. Since olivine is absent in the subsolidus assemblage in this composition at 1,100° C, 11.3 kb (Green and Ringwood, 1967), there is either a subsolidus reaction (e.g. spinel + orthopyroxene + clinopyroxene = olivine + anorthite) to yield olivine above 1,100° C or an incongruent melting relationship in the opx + cpx + plag + spinel assemblage.

A second run, at 11.3 kb 1,270° C, was carried out on the alkali olivine basalt composition. This run yielded glass as the major phase but with quite common clinopyroxene and rare orthopyroxene. The firm identification of olivine was not possible.

d) Crystallization at 13.5 kb

In contrast to the important role of olivine at 9 kb, this phase does not occur at 13.5 kb in any composition studied except the picrite. Orthopyroxene is the liquidus phase in both the olivine tholeiite and olivine basalt compositions and both orthopyroxene and clinopyroxene occur near or on the liquidus in the alkali olivine basalt. In the olivine tholeiite composition, there is an extensive field of crystallization of orthopyroxene alone (Fig. 2B) from a temperature of 1,400° C or slightly greater, down to about 1,360° C. Orthopyroxene is then joined by clinopyroxene and these two phases crystallize together down to 1,300° C. At 1,350° C orthopyroxene is more abundant than clinopyroxene and both occur as large, well-formed crystals. In this run, glass is more common than crystals but in the 1,310° C run crystals are more abundant than glass and the small, equant or short, prismatic crystals of clinopyroxene are much more abundant than the distinctive large orthopyroxene laths. Spinel first appears in the 1,290° C run and remains a minor phase down to a temperature between 1,100° C and 1,180° C where garnet first appears.

Plagioclase does not appear in the olivine tholeiite until practically at the solidus at 1,250° C approximately (for the dry composition). In 1 hr and 3 hr runs at 1,220° C, the presence of minor amounts of glass and uncertainty in plagioclase identification are attributed to access of small amounts of water to the charge, resulting in depression of the solidus. The 3 hr run at 1,180° C showed appreciable plagioclase, but less than observed at 1,100° C—this could result from the presence of minor glass in the run or from increasing solubility of Al_2O_3 in the pyroxenes at higher temperatures. The replacement of the px + plag + spinel assemblage by px + plag + ga with decreasing temperature at 13.5 kb is matched by the same reactions with increasing pressure at 1,100° C (Green and Ringwood, 1967).

The liquidus phase in the olivine basalt at 1,330° C is also orthopyroxene but in this composition there is a much smaller temperature interval before appearance of the second phase, clinopyroxene. Definite primary clinopyroxene first appears at 1,310° C but in this run is less abundant than orthopyroxene. There is a marked increase in the degree of crystallization between 1,310° C and 1,290° C and in the latter run clinopyroxene is much more abundant than the orthopyroxene.

In the alkali olivine basalt composition, a run for 60 min at 1,290° C, 13.5 kb is very near the liquidus with a very few large euhedral crystals of both orthopyroxene and clinopyroxene. The composition of the glass used is given in Table 1. In a later series of runs, a new batch of glass of slightly different composition (normative olivine 28%; normative nepheline 3.1%; normative diopside 16.0%) was used for 30 min runs at 1,290° C to 1,310° C. In these runs it is difficult to distinguish fine-grained anhedral quench clinopyroxene from aggregates of fine-grained primary clinopyroxene. None of these runs contain large crystals of either orthopyroxene or clino- pyroxene but the 1,290° C and 1,300° C runs both contain uncommon medium-sized rectangular laths with straight extinction and very low birefringence which are probably orthopyroxene. These crystals have rims and outgrowths of quench clinopyroxene.

At 1,270° C and 1,250° C the alkali olivine basalt crystallizes to clinopyroxene and glass and orthopyroxene is not identifiable.

The picrite differs from the three compositions described above in that olivine and not pyroxene is the liquidus phase at 13.5 kb. The liquidus temperature was not established but is greater than 1,380° C. Orthopyroxene is the second phase to appear — it is present in a 40 minute run at 1,360° C but only olivine was observed in a 30 minute run at the same temperature. There is a rather abrupt increase in degree of crystallization between 1,360° C and 1,340° C and clinopyroxene is the major phase at 1,340° C. Olivine continues to remain stable with both the orthopyroxene and clinopyroxene down to at least 1,320° C. The roles of orthopyroxene and clinopyroxene in the picrite composition are similar to those in the olivine basalt composition in that orthopyroxene appears at the higher temperature but there is only a small temperature drop before the crystallization of abundant clinopyroxene.

e) Crystallization at 18 kb

In all compositions, except the alkali olivine basalt, the liquidus phase at 18 kb is the same as that observed at 13.5 kb. Orthopyroxene is the liquidus phase in the olivine tholeiite and olivine basalt composition but in the alkali olivine basalt composition, orthopyroxene does not appear at all at 18 kb and clinopyroxene is the liquidus phase. The temperature interval over which orthopyroxene alone co-exists with the liquid is much smaller at 18 kb than 13.5 kb in the olivine tholeiite and olivine basalt compositions.

The liquidus of the olivine tholeiite at 18 kb is about 1,430° C and orthopyroxene co-exists with liquid down to 1400° C or 1,410° C. At 1,400° C, 1,380° C and 1,375° C both pyroxenes form large crystals, in some cases showing parallel intergrowth of the two pyroxene types. In the lower temperature runs clinopyroxene is more abundant than the orthopyroxene and at 1,370° C orthopyroxene is not identifiable, the run consisting entirely of clinopyroxene + liquid. The disappearance of orthopyroxene at lower temperatures in the partial melting interval is analogous to the behaviour of the alkali olivine basalt at 13.5 kb. The runs at 18 kb differ from those at lower pressures in showing the appearance of garnet at temperatures above the solidus. Garnet first appears at 1,350° C and increases in abundance in lower temperature runs. The solidus at 18 kb is close to 1,300° C giving a partial melting or crystallization interval of about 130° C.

The behaviour of the olivine basalt at 18 kb is notable for the very small temperature interval over which crystallization and rapid changes in phase assemblages occur. The liquidus is at 1,350 ± 10° C and duplicate runs at 1,340° C both yielded orthopyroxene + liquid. At 1,335° C, the mineral assemblage consists of large primary orthopyroxene and primary clinopyroxene with liquid but at 1,330° C orthopyroxene is not present, the degree of crystallization is much greater and garnet appears as a minor phase with the abundant clinopyroxene. The solidus for this composition is near 1,300° C. The sequence of appearance of phases in the olivine basalt matches that observed in the olivine tholeiite but the different fields for the various phase assemblages are compressed into a melting interval of about 50° C.

The liquidus of the alkali olivine basalt at 18 kb is at 1,325° C approximately. The rare clinopyroxenes present in the 1,320° C run are large and well crystallized but there is an abrupt increase in the degree of crystallization between 1,320° C and 1,300° C. Clinopyroxene is the only primary crystalline phase present in the 1300° C run but garnet is moderately common at 1,250° C.

In the picrite composition, olivine is still the liquidus phase at 18 kb and is joined by orthopyroxene in the 1,410° C and 1,400° C runs. In these runs there is no definite identification of primary clinopyroxene but this becomes a major phase and orthopyroxene disappears in the 1,390° C and 1,370° C runs. Minor garnet first appears in a near-solidus run at 1,350° C Unlike orthopyroxene, olivine persists as a stable phase right through to the subsolidus assemblage.

f) Crystallization at 22.5 kb

At 22.5 kb, clinopyroxene is the liquidus phase in both the olivine tholeiite and the picrite compositions. No runs have been carried out at this pressure on either the alkali olivine basalt or olivine basalt compositions. In the olivine tholeiite composition, the amount of clinopyroxene crystallizing increases rapidly over a small temperature interval. A run at 1,410° C, only 30—40° C below the liquidus, contains common garnet and is very near the solidus. The crystallization interval in the picrite at 22.5 kb is also small. Clinopyroxene appears to be the liquidus phase at 1,450° C and there is only minor glass present at 1,430° C. In the picrite composition, garnet and olivine appear at similar temperatures and there is also a small amount of a brownish high R.I. phase. The determination of the solidus is particularly difficult in this composition due to outgrowths of quench clinopyroxene on the abundant primary clinopyroxene crystals.

g) Crystallization at 27 kb

In all three basaltic compositions, garnet is the liquidus phase at 27 kb and is joined by clinopyroxene at about 10° C below the liquidus. In the picrite composition garnet, clinopyroxene and orthopyroxene all appear together in the near-liquidus run at 1,500° C. Orthopyroxene disappears at lower temperature (between 1,480° C and 1,500° C) and minor amounts of either spinel$_{ss}$ or ilmenite$_{ss}$ occur. The run at 1,480° C is very near the solidus and a run at 1,430° C is below the solidus. The latter run does not contain detectable olivine but olivine is a minor but definite phase at 27 kb, 1,200° C. The melting interval of the picrite at 30 kb is an extremely small one of about 40° C — the melting interval of similar compositions at atmospheric pressure is about 375° C (1,425° C to 1,050° C, cf. TILLEY, YODER and SCHAIRER, 1965).

In the olivine tholeiite composition the liquidus is at about 1,500° C and the solidus at 1,430° C. Garnet is probably the liquidus phase although the presence of quench clinopyroxene in the 1.490° C run makes the absence of primary clinopyroxene difficult to establish. The presence of euhedral garnet in clear glass in parts of the olivine basalt runs (at 1,450° C and 1,460° C) and in the alkali olivine basalt run at 1,430° C demonstrates that garnet is the liquidus phase in these compositions (Fig. 2D). The quench clinopyroxene forms feathery and turbid aggregates and, in some cases, quite large anhedral crystals which are distinguishable as quench rather than primary crystals only by their patchy occurrence, the presence of characteristic lines of small inclusions and the absence of a change in refractive index at edges against feathery, definite quench material. Garnet and clinopyroxene are the only phases observed in the crystallization sequence. The solidus of the alkali olivine basalt at 30 kb is about 1,350° C giving a melting interval for this composition of about 90° C. The solidus of the olivine basalt was not determined.

h) Stability of Orthopyroxene as a Liquidus Phase in the 9—18 kb Pressure Interval

YODER (1964) found enstatite as a liquidus phase in the forsterite-albite system at 9 kb but interpreted the enstatite as metastable, proxying for olivine. Extending this interpretation, TILLEY and YODER (1964) suggested, as a possible interpretation of GREEN and RINGWOOD's (1964) results, that orthopyroxene appearing as a liquidus phase from basaltic liquids at moderate pressure might also be metastable, crystallizing in place of clinopyroxene. Arguments against this interpretation may be developed from the observed readiness of clinopyroxene to nucleate, even during quenching; from the experimental runs in which both orthopyroxene and clinopyroxene occur as co-existing phases and from the restriction of orthopyroxene as a liquidus phase to a definite pressure interval. To demonstrate the stability of orthopyroxene, rather than either garnet or clinopyroxene, as the liquidus phase at 13.5 kb in the olivine tholeiite composition, the glass was run for 4 hrs at 27 kb, 1,100° C. This yielded an assemblage of fine-grained clinopyroxene and common, larger subhedra or euhedra of garnet. The garnet + clinopyroxene assemblage was then re-run for 1 hr at 1,380° C, 13.5 kb yielding an assemblage of euhedral orthopyroxene, uncommon relict garnet and glass + minor quench clinopyroxene. The relict garnet shows clear evidence of partial solution and is not a stable phase under these conditions. The opx + liquid assemblage obtained from an eclogite (ga + cpx) starting material is the same as that obtained under the same run conditions using glass starting material. The runs carried out in the olivine tholeiite for only 5 mins. at 13.5 kb demonstrate that clinopyroxene, rather than orthopyroxene, may crystallize metastably in very short runs but these runs also show that at temperatures of 1,400° C, run lengths of about 5 minutes may be sufficient to allow nucleation and moderate growth of stable liquidus phases. These observations effectively refute TILLEY and YODER's (1964) suggestion that the orthopyroxene reported in our earlier runs was a product of metastable crystallization.

i) Microprobe Analyses of Crystals in Partial Melting Runs

Microprobe analyses of olivine, pyroxenes and garnets are listed in Tables 8—12. The analysis of clinopyroxene was only practical in runs in which this phase grew as large well-formed crystals, i.e. near-liquidus runs. The determination by direct analysis of only three elements (Fe, Ca and Al) means that the contents of Ti and Na are unknown — these constituents are unlikely to be of significance in either olivine or orthopyroxene but the clinopyroxene may have appreciable Ti (0.5—2% TiO_2) and low Na (0.5% Na_2O) contents if comparison is made with early-formed clinopyroxenes of natural basalts. This is further substantiated by later microprobe analyses in this laboratory of clinopyroxenes in other basaltic compositions by T. H. GREEN and R. F. BULTITUDE (unpublished data). These analyses include direct determination of Na, Ti contents of the clinopyroxenes.

j) Olivine Analyses (Table 8)

The olivines in all compositions show an extremely low Al_2O_3 content and a low but measurable CaO content of 0.3%. This compares closely with the 0.4% CaO in olivine from Kilauean tholeiites reported in a preliminary study by MURATA, BASTRON and BRANNOCK (1965). Liquidus olivines have consistently higher Mg/Mg + Fe molecular ratios than olivines formed below the liquidus and the correlation of nominal temperature with Mg/Mg + Fe ratios for a given composition gives an independent check on the reliability of relative temperature intervals between runs. On this basis the olivine (100 Mg/Mg + Fe = 92.2) at 4.5 kb, 1,350° C in the olivine tholeiite is closer to the liquidus than the olivine (100 Mg/Mg + Fe = 90.8) at 10 kb, 1,350° C. The partition of magnesium and iron between liquidus olivine and liquid composition has previously been discussed and an empirical mean partition coefficient of 1.33 has been derived for the relationship between basaltic liquid and liquidus olivine or orthopyroxene of composition range 92 > 100 Mg/Mg + Fe > 82. This empirical partition coefficient is defined by the ratio

$$\left(\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}}\right)_{\mathrm{Crystal}} \Big/ \left(\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}^{++}}\right)_{\mathrm{Liquid}} = \mathrm{K}_{\mathrm{O,\,L}}.$$

In each run, ten separate olivine crystals were analyzed and the analysis of each crystal is the mean of 3 to 10 separate spot analyses 2 to 3 microns apart. The analysis given in Table 8 is the mean of the ten separate crystal analyses. It was observed that some olivine crystals showed slight zoning with more iron rich cores than rims. Also it was observed that crystals located close to the platinum

Table 8. *Compositions of olivines analysed by electron microprobe. Fe, Ca, Al determined by direct analysis; other components calculated*

	Olivine tholeiite								Olivine basalt	Alkali olivine basalt	
Pressure (kb) Temp. (°C)	4.5 1,350	9.0 1,350	9.0 1,330	9.0 1,310	9.0 1,290	9.0 1,270	9.0 1,250	9.0 1,230	9.0 1,280	9.0 1,260	9.0 1,220
SiO_2	41.1	40.9	40.8	40.7	40.3	40.2	39.9	39.8	40.7	39.6	39.2
Al_2O_3	n.d.**	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
FeO	7.7	8.9	9.9	10.3	12.3	12.9	14.0	14.7	10.1	16.1	18.0
MgO	50.9	49.9	49.0	48.7	47.1	46.6	45.8	45.2	48.9	44.0	42.5
CaO	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3	0.3
$\frac{100\ \text{Mg}}{\text{Mg}+\text{Fe}}$ (mol)	92.2	90.9	89.8	89.4	87.2	86.4	85.3	84.7	89.6	83.0	80.8
Mol. Proportions											
Fo	91.8	90.6	89.5	89.0	86.8	86.1	84.9	84.3	89.3	82.7	80.5
Fa	7.8	9.0	10.1	10.6	12.8	13.5	14.7	15.3	10.3	16.9	19.1
La	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.4	0.4
Coexisting Crystalline phases	Nil	Nil	Nil	Nil	Opx*	Opx*	Opx* + Cpx	Opx + Cpx + ? Plag	Nil	Nil	Cpx*

* Denotes crystalline phase analyzed by electron microprobe.

** "not detected", i.e. $Al_2O_3 < 0.2\%$.

wall of the capsule were slightly poorer in iron than those near the centre of the capsule. As an example the FeO content of a large olivine grain (25 μ traverse) in the 10 kb, 1,350° C, olivine tholeiite run ranged from 9.3% FeO in the centre to 9.0% FeO at the edge and averaged 9.2% FeO. In the mount as a whole the mean FeO content of the crystals ranged from 9.2% FeO to 8.7% FeO in two crystals located near the capsule wall. The mean FeO content of all the crystals was 8.9%. The variation from 8.7% FeO to 9.3% FeO represents a change in the 100 Mg/Mg + Fe ratio from 91.1 to 90.5. Similarly the maximum variation in crystal composition in the olivine of the 10 kb, 1,260° C run on the alkali olivine basalt composition is from 15.5% FeO to 16.6% FeO with an average of 16.1% FeO. This variation represents a change in 100 Mg/Mg + Fe ratio from 83.6 to 82.4 with a mean of 83.0. Thus we consider that there is an uncertainty in composition of about ± 0.3% forsterite in olivines around Fo_{90} and about ± 0.6% forsterite in olivines around Fo_{80}. This uncertainty derives from real variations in 100 Mg/Mg + Fe ratio of crystals with position in the mount and may be a consequence of Fe loss to the Pt crucible at the margins of the sample or of small temperature gradients within the sample capsule.

k) Orthopyroxene Analyses

The analyses of orthopyroxenes listed in Table 9 also demonstrate an increase in FeO content and decrease in the 100 Mg/Mg + Fe ratio with decrease in temperature at a given pressure. The partition of Fe and Mg between liquid and liquidus crystals where orthopyroxene is the liquidus phase is similar to runs in which olivine is the liquidus phase (cf. Table 10). The orthopyroxenes are distinctive in all cases in having a moderate or high Al_2O_3 content, similar to or higher than that observed in enstatites of the Ol + Opx + Cpx + Sp assemblage of peridotite nodules, and higher than the Al_2O_3 contents of orthopyroxenes crystallizing at low pressure from basaltic or andesitic magmas. At any given pressure, the Al_2O_3 content of orthopyroxene increases with decrease in temperature, as long as no other co-existing Al_2O_3-rich phase appears. At constant temperature, the Al_2O_3 content of orthopyroxene increases with increase in pressure (cf. runs on olivine tholeiite at 1,400° C, 13.5 kb and 18 kb), provided closely similar phase assemblages are compared. Alternatively, comparing orthopyroxenes of similar 100 Mg/Mg + Fe ratios, those crystallized at higher pressure, have higher Al_2O_3 contents.

The series of analyses of orthopyroxenes from the olivine tholeiite composition at 13.5 kb best illustrate the change in orthopyroxene composition with increasing degree of crystallization. The liquidus orthopyroxene at 1,400° C contains only 3.8% Al_2O_3 but this increases to 7.8% Al_2O_3 in the 1,310° C run. The run at 1,290° C reverses this trend with a drop in Al_2O_3 to 6.4%. We attribute the lower Al_2O_3 content of the 1,290° C orthopyroxene to the first appearance of spinel as a co-existing phase — this will alter the partition relationship of Al_2O_3 between liquid and crystallizing orthopyroxene. The Al_2O_3 content of the liquidus orthopyroxene at 18 kb is greater than that of the liquidus orthopyroxene at 13.5 kb in both the olivine basalt and olivine tholeiite compositions. The contrast between the Al_2O_3 content of the orthopyroxene in the olivine basalt, 18 kb 1,340° C

Table 9. *Compositions of orthopyroxenes analyzed by electron microprobe. Fe, Ca, Al determined by direct analysis, other components calculated assuming ideal orthopyroxene composition*

	Olivine tholeiite										Olivine basalt			Alkali olivine basalt
Pressure (kb)	9.0	9.0	9.0	13.5	13.5	13.5	13.5	13.5	18.0	18.0	13.5	18.0	18.0	13.5
Temp. (° C)	1,290	1,270	1,250	1,400	1,380	1,350	1,310	1,290	1,420	1,400	1,320	1,340	1,335	1,290
SiO_2	54.5	53.9	52.9	55.9	54.4	53.8	52.5	53.1	54.9	53.4	53.8	51.8	54.3	50.6
Al_2O_3	4.9	5.4	6.9	3.8	5.7	6.0	7.8	6.4	5.1	7.3	6.4	10.0	5.9	10.6
FeO	7.8	8.3	9.1	5.5	6.5	7.2	8.7	9.7	5.8	6.4	6.6	6.5	6.4	9.7
MgO	30.0	29.6	28.5	32.8	31.2	30.2	28.3	28.3	32.0	30.4	30.3	29.5	30.8	26.6
CaO	2.9	2.8	2.6	2.0	2.2	2.8	2.7	2.5	2.2	2.5	2.9	2.2	2.6	2.5
$\frac{100\ Mg}{Mg+Fe}$ (mol)	87.3	86.4	84.9	91.4	89.6	88.2	85.3	83.9	90.8	89.5	89.1	89.0	89.6	83.0
Mol. Proportions														
Ens	78.1	77.0	74.4	84.5	80.6	78.1	74.1	74.2	82.5	78.6	78.4	76.2	79.8	69.9
Fs	11.3	12.1	13.3	8.0	9.4	10.5	12.7	14.3	8.3	9.2	9.6	9.4	9.3	14.3
Wo	5.5	5.3	5.0	3.7	4.2	5.2	5.1	4.8	4.0	4.7	5.4	4.2	4.8	4.8
Al_2O_3	5.1	5.6	7.3	3.8	5.8	6.2	8.1	6.7	5.2	7.5	6.6	10.2	6.1	11.0
Coexisting Crystalline phases	Ol*	Ol*	Ol* + Cpx	Nil	Nil	Cpx	Cpx*	Cpx + Sp**	Nil	Cpx*	Nil	Nil	Cpx*	Cpx*

* Denotes crystalline phase analyzed by electron microprobe.

** Spinel has approximately 14% FeO, $70 \pm 5\%$ $Al_2O_3 < 1.3\%$ CaO.
(Calculated composition assuming $FeAl_2O_4$, $MgAl_2O_4$ solid solution 14% FeO, 67.3% Al_2O_3, 18.7% MgO.)

Table 10. *Comparison of 100 Mg/Mg + Fe⁺⁺ values of liquidus olivine and liquidus orthopyroxene crystals*

	Liquidus olivine			Liquidus orthopyroxene		
	Pressure (kb)	Temp. (°C)	$\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}^{++}}$	Pressure (kb)	Temp. (°C)	$\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}^{++}}$
Olivine Tholeiite	4.5	1350	92.2	13.5	1400	91.4
,, ,,	9	1350	90.9	18	1420	90.8
Olivine Basalt	9	1280	89.6	13.5	1320	89.1
,, ,,				18	1335	89.6
,, ,,				18	1340	89.0
Alkali Olivine Basalt	9	1260	83.0	13.5	1290	83.0

(10% Al_2O_3) run and the 1,335° C (5.9% Al_2O_3) run is not readily understood but may be an effect of the presence of aluminous clinopyroxene (9.9% Al_2O_3) in the 1,335° C run and its absence in the 1,340° C run.

The CaO content of the orthopyroxene is lower (2.0—2.2% CaO) in runs in which there is no co-existing clinopyroxene and increases to 2.5 or 2.8% CaO in these runs in which stable clinopyroxene occurs. This is shown by the olivine tholeiite runs at 13.5 kb and 18 kb and the olivine basalt runs at 18 kb but the olivine basalt 13.5 kb 1,320° C run and the 9 kb olivine tholeiite runs are exceptions to this generalization.

Table 11. *Compositions of clinopyroxenes analyzed by electron microprobe. Fe, Ca, Al determined by direct analysis, other components calculated assuming ideal clinopyroxene composition and little or no Na or Ti substitution*

	Olivine tholeiite		Olivine basalt	Alkali olivine basalt		
Pressure (kb)	13.5	18.0	18.0	9.0	13.5	18.0
Temp. (°C)	1,310	1,400	1,335	1,220	1,290	1,320
Analysis No.	1	2	3	4	5	6
SiO_2	49.3	52.3	50.7	49.4	49.2	49.1
Al_2O_3	11.3**	7.3	9.9	9.0	11.2	11.0
FeO	8.3	6.0	5.5	9.5	9.0	7.6
MgO	21.2	25.1	23.1	16.9	20.9	19.5
CaO	9.9	9.3	10.8	15.2	9.7	12.8
$\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}}$	82.1	88.2	88.2	76.1	80.6	82.1
Mol. Proportions						
Ens	56.5	66.0	61.0	46.1	56.0	52.5
Fs	12.4	8.8	8.1	14.5	13.5	11.3
Wo	19.1	17.6	20.6	29.5	18.6	24.6
Al_2O_3	12.0	7.6	10.3	9.9	11.9	11.6
Coexisting						
Crystalline phases	Opx*	Opx*	Opx*	Ol*	Opx*	Nil

* Denotes crystalline phase analyzed by electron microprobe.

** Doubtful value, possibly too high due to interference from glass. Crystals are very small.

l) Clinopyroxene Analyses

Four of the analyzed clinopyroxenes of Table 11 co-exist with orthopyroxene + liquid, one co-exists with olivine + liquid and one co-exists only with liquid. The four clinopyroxenes which co-exist with orthopyroxene are lower in CaO content and may be classified as sub-calcic augites. Within this group there is an increase in CaO content with decrease in temperature (analyses 2 and 3) and the more magnesian clinopyroxenes have higher CaO contents at similar temperatures than more iron-rich clinopyroxenes (analyses 1 and 3, Fig. 3). These relationships are a consequence of the widening of the Opx-Cpx immiscibility gap (Boyd and Schairer, 1964; Davis and Boyd, 1966) with decreasing temperature and the narrowing of this gap with increasing Fe-content or decreasing Mg/Mg + Fe content of the pyroxenes.

Table 12. *Garnet compositions determined by electron microprobe. Fe, Ca, Al determined by direct analysis, other components calculated*

	Olivine basalt		
Pressure (kb)	18	18	27
Temp. (°C)	1,330	1,320	1,460
SiO	41.6	41.4	42.1
Al_2O_3	23.3	22.6	23.9
FeO	11.5	11.8	8.7
MgO	16.9	16.6	19.5
CaO	6.4	6.6	5.7
Al_2O_3 *	23.6	23.6	24.0
$\frac{100\ Mg}{Mg + Fe}$ (mol.)	72.3	71.5	80.0
Mol. Proportions			
Almandine	23.0	23.6	17.2
Pyrope	60.5	59.4	68.3
Grossular	16.5	17.0	14.5
Coexisting			
Crystalline Phases	Cpx	Cpx	Nil

* Al_2O_3 value calculated from FeO, CaO values assuming garnets are ideal almandine-pyrope-grossular solid solutions.

In contrast to the four clinopyroxenes described above, the pyroxene compositions in equilibrium with the alkali olivine basalt liquid at 1,320° C, 18 kb and in equilibrium with olivine + liquid at 1,220° C, 9 kb (analysis 4), lie outside the pyroxene immiscibility gap for their compositions and P-T conditions. These pyroxenes are richer in CaO than those co-existing with orthopyroxene.

The clinopyroxenes generally contain more Al_2O_3 than the co-existing orthopyroxene although more data is needed to substantiate this. Thus clinopyroxene analyses 2 and 5 have similar or slightly greater Al_2O_3 contents than their co-existing orthopyroxenes but clinopyroxenes 1 and 3 have much greater Al_2O_3 contents than co-existing orthopyroxene. The 100 Mg/Mg + Fe values for the clinopyroxenes are consistently lower than co-existing olivine or orthopyroxene.

The analyzed clinopyroxenes and co-existing orthopyroxenes have been projected on to the experimentally determined phase diagram for the $CaMgSi_2O_6$-$MgSiO_3$ system at atmospheric pressure (Boyd and Schairer, 1964). Davis (1963) and Davis and Boyd (1966) have demonstrated that the position of the pyroxene solvus at a given temperature is virtually unaffected by pressure. The data from the microprobe analyses demonstrate that the width of the pyroxene solvus is greatly decreased by increasing Fe content. Surprisingly, the solubility of diopside-hedenbergite in enstatite-hypersthene is less than that determined for the simple magnesian system, but the solubility of orthopyroxene in calcic clinopyroxene is much greater. The data of Fig. 3 demonstrate preliminary results in a method of obtaining a pyroxene-diagram for natural basaltic magmas at high pressures.

m) Garnet Analyses

In the olivine basalt composition, two garnets co-existing with clinopyroxene at 18 kb and the liquidus garnet at 27 kb have been analyzed. As the oxidation state of iron cannot be determined with the microprobe the garnets are assumed to contain no andradite molecule but to consist entirely of grossular + almandine + pyrope solid solutions. With this assumption it is possible to calculate the whole garnet composition from the Fe and Ca determinations. The Al_2O_3 contents calculated in this way compare well with the directly determined Al_2O_3.

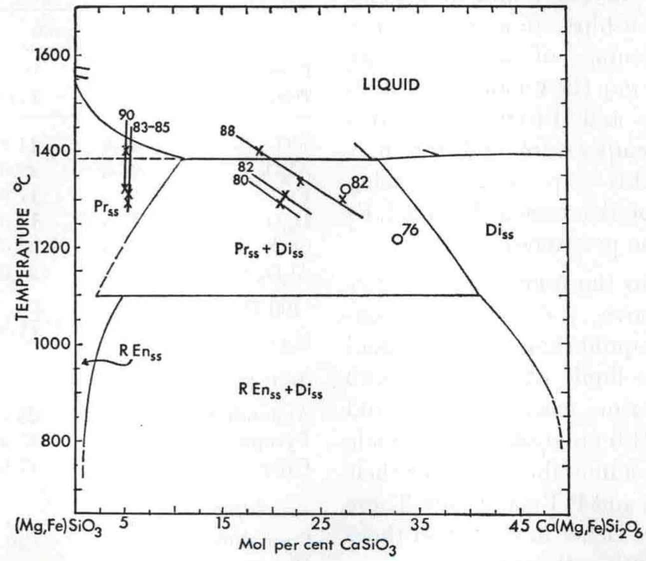

Fig. 3. Diagram illustrating the degree of solid solubility in co-existing orthopyroxene and clinopyroxene. The pyroxene solvus for pure $MgSiO_3$—$CaMgSi_2O_6$ is taken from Boyd and Schairer (1964) and the compositions of the pyroxenes observed in the partial melting runs on basalt are projected on to this solvus. × = two pyroxene + liquid assemblages; ○ = clinopyroxene + liquid only; 90, 88 etc. values of 100 Mg/Mg + Fe for the pyroxene — note that the clinopyroxene always has a lower 100 Mg/Mg + Fe ratio than the co-existing orthopyroxene

No phases co-existing with garnet have been analyzed but the liquidus garnet at 27 kb with 100 Mg/Mg + Fe = 80 is much more iron-rich than the liquidus olivine at 9 kb (100 Mg/Mg + Fe = 89.6) or the liquidus orthopyroxene at 13.5 kb (100 Mg/Mg + Fe = 89.1) or 18 kb (100 Mg/Mg + Fe = 89.0).

Fractional Crystallization of Basaltic Magmas

The nature of the fractionation trend and the derivative liquid compositions from any primitive basaltic magma may be determined if we know the composition of the crystalline phase or phases and the proportions of liquid and crystals.

Our data show that the nature of the liquidus phase, the sequence of appearance and nature of other phases, and the temperature interval for crystallization are strongly dependent on load pressure. The data on which the following discussion is based are shown in Figs. 4, 5, 6 and 7. Principal fractionation trends at several different pressures are compared in Figs. 8 and 9.

a) Fractionation at Upper Crustal Levels (<15 km)

The behaviour of the three basalts during fractionation in the upper continental crust or within the oceanic crust or uppermost mantle (i.e. at pressures <5 kb) can be discussed on the basis of the experimental results at atmospheric pressure on the olivine tholeiite and alkali olivine basalt compositions. A further, most important, source of information on the fractionation of basaltic magmas at low pressures is the detailed and comprehensive study of basalt melting relations at 1 Atmosphere carried out by Tilley, Yoder and Schairer (1963, 1964, 1965). The observed sequence of appearance of phases at atmospheric pressure (Table 4) and in particular the appearance of plagioclase before clinopyroxene, could also have been deduced by comparison of the olivine tholeiite and alkali olivine basalt with the chemical compositions and crystallization sequences of the basalts studied by Yoder and Tilley (1962); Tilley, Yoder and Schairer (1963, 1964, 1965) and Yagi (1964). Although the early crystallization of olivine in the olivine tholeiite composition results in a fractionation trend of SiO_2, Al_2O_3, and CaO

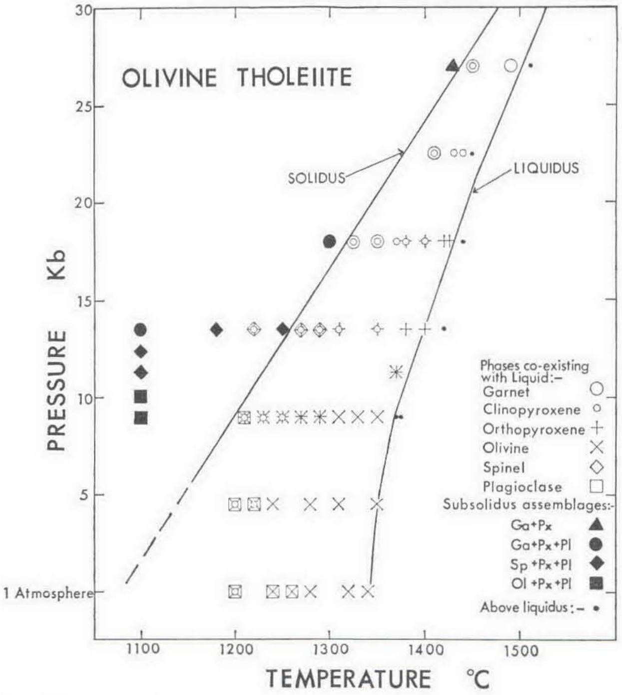

Fig. 4. Detail of melting experiments carried out on the olivine tholeiite composition at high pressures. Run descriptions are given in Table 4

enrichment at low pressure, the appearance of very calcic plagioclase as the second phase curtails the Al_2O_3 and CaO enrichment trend (Table 13), instead maintaining the well documented, tholeiitic fractionation to iron-rich, Na + K enriched quartz tholeiites with 50% or more SiO_2 and 12—14% Al_2O_3.

As the olivine tholeiite composition was initially derived from an estimate by Macdonald and Katsura (1961) for the average magma composition of the 1959—1960 Kilauea Iki lava lake, it is of interest to compare the low pressure fractionation trend of the olivine tholeiite with the thoroughly documented fractionation trends of the 1959—1960 Kilauea Iki eruption. Tilley, Yoder and Schairer (1963, 1964, 1965) have shown experimentally that rocks and glasses derived from the 1959—1960 Kilauea Iki and other Kilauean eruptions contain olivine as the liquidus phase and that olivine alone remains the liquidus phase even into the field of quartz-normative liquids. The same conclusion also follows from the quartz-normative composition for the low magnesia terminus of the olivine control lines in the 1950—1960 Kilauea Iki differentiation series of Murata and Richter (1966a, b). Tilley et al. (op. cit.) have shown that clinopyroxene is usually the second phase to crystallize from Kilauean lavas but may appear simultaneously with plagioclase or be very closely followed by plagioclase. The

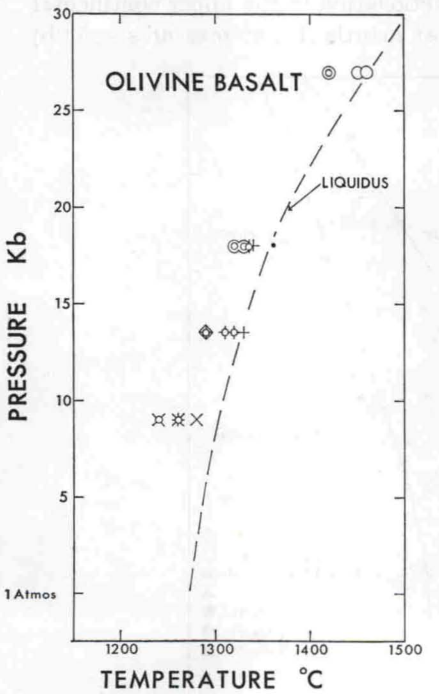

Fig. 5

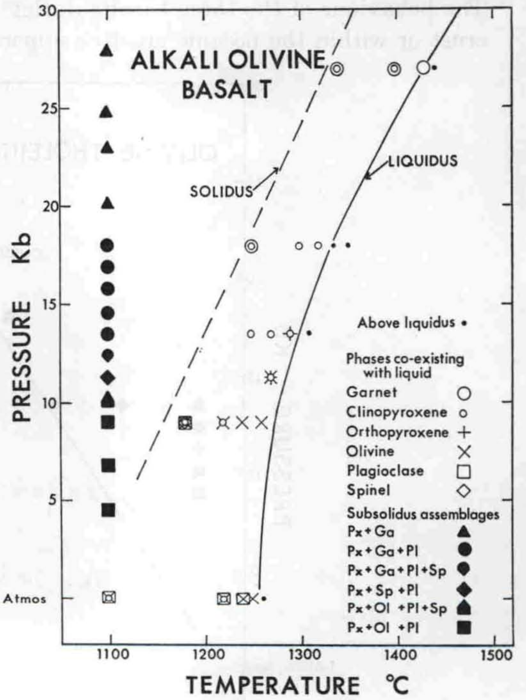

Fig. 6

Fig. 5. Detail of melting experiments on the olivine basalt composition at high pressures. Symbols are the same as those in Figs. 4 and 6. Run descriptions are given in Table 5

Fig. 6. Detail of melting and subsolidus experiments on the alkali olivine basalt composition. Run descriptions are given in Table 6

Table 13. *Hypothetical fraction of the olivine tholeiite at Atmospheric pressure*

Nature and estimated percentage of minerals		15% Olivine (Fo_{85})	18% Olivine (Fo_{85}) 12% Plagioclase (An_{80})
Composition of crystal extract			
SiO_2		40.0	43.1
Al_2O_3		—	13.4
FeO		14.3	8.7
MgO		45.7	27.4
CaO		—	6.4
Na_2O		—	1.0
Composition of residual liquid			
	(Initial liquid)		
SiO_2	46.95	48.1	48.6
TiO_2	2.02	2.4	2.9
Al_2O_3	13.10	15.5	13.0
Fe_2O_3	1.02	1.2	1.4
FeO	10.07	9.3	10.7
MnO	0.15	0.2	0.2
MgO	14.55	9.0	9.1
CaO	10.16	12.0	11.8
Na_2O	1.73	2.0	2.0
K_2O	0.08	0.1	0.1
P_2O_5	0.21	0.2	0.3
CIPW Norm			
Or	0.67	0.7	0.8
Ab	14.7 } 42.9	16.8 } 50.5	16.7 } 43.9
An	27.6	33.0	26.4
Di	17.0	20.2	24.6
Hy	12.3	15.3	20.2
Ol	21.9	7.1	3.2
Ilm	3.8	4.6	5.5
Mt	1.4	1.8	2.0
Ap	0.5	0.5	0.6
$\frac{100\,Mg}{Mg+Fe^{++}}$	72.1	63.3	60.3

appearance of plagioclase before clinopyroxene in our olivine-tholeiite composition and the relatively early appearance of plagioclase in relation to the Kilauea Iki fractionation (MURATA and RICHTER, 1965) are attributed to compositional differences. Thus the analyses that MACDONALD and KATSURA (1961, Table 1, analyses 2 and 3) used to estimate their parent magma do not fall on MURATA and RICHTER's (1966a, b) differentiation trend but are distinctly higher in Al_2O_3 (cf. page 110 and footnote). The experimental and natural studies of low pressure fractionation are sufficient to form the conclusion that separation of olivine, plagioclase and clinopyroxene, in that order, from the chosen olivine tholeiite composition will result in derivative liquid compositions equivalent to Fe-enriched quartz tholeiites, with normal Al_2O_3 contents. At these low pressure conditions, derivative liquids from the olivine tholeiite composition will move out of the undersaturated olivine tholeiite field into the oversaturated quartz-tholeiite field (Fig. 1).

The results from the alkali olivine basalt composition are consistent with results from similar rocks studied experimentally by TILLEY, YODER and SCHAIRER (1963, 1964, 1965) and YODER and TILLEY (1962). The effect of olivine and calcic plagioclase separation would be to increase the normative nepheline content and also the SiO_2 content. The relatively early appearance of plagioclase prevents any large increase in Al_2O_3 content. The appearance of clinopyroxene amongst the phases crystallizing would result in either an increase or possibly slight decrease of normative nepheline content depending upon the composition of the particular pyroxene. In general terms, crystal fractionation at low pressures of compositions close to our alkali olivine basalt would yield liquids retaining normative nepheline but becoming richer in SiO_2, alkalis, Al_2O_3 (in early fractionation stages) and $Fe/Fe + Mg$ ratio. Some hawaiites and mugearites are probably natural examples of this low pressure fractionation trend (MUIR and TILLEY, 1961) although TILLEY, YODER and SCHAIRER (1965) have recently suggested that crystal flotation involving plagioclase enrichment with extraction of olivine and clinopyroxene may be involved in the genesis of hawaiite and mugearite.

Although the olivine basalt composition contains 1% of normative hypersthene, fractionation at low pressure may produce residual liquids closer to those derived from the alkali olivine basalt than from the olivine tholeiite. As in the other compositions, the crystallization sequence at atmospheric pressure is probably olivine-plagioclase-clinopyroxene. Separation of olivine and calcic plagioclase will result in a small increase in normative hypersthene content but the appearance of clinopyroxene may reverse this trend and even produce nepheline-normative liquids. This may occur if the clinopyroxene has sufficiently high $(Mg, Fe) SiO_3$ solid solution and low $Al_2O_3 + Na (Al, Fe^{+++}) Si_2O_6$ solid solution to yield a hypersthene normative composition. The olivine basalt and compositions derived from it in early stages of low pressure fractionation belong to the transitional group of rocks discussed by POLDERVAART (1964), which are hypersthene normative but show crystallization and fractionation sequences more typical of the alkali basalt suite. Although a nepheline normative liquid *may* be produced by low pressure fractionation from the olivine basalt composition it must be emphasized that this can only result after crystallization of large amounts of olivine, plagioclase, clinopyroxene and titanomagnetite — the residual liquid would in fact be trachytic and not basaltic.

b) Fractionation at about 15—35 km Depth

The experimental partial melting runs carried out at 9 kb provide the data on which the following discussion is based. The liquidus phase in all compositions studied is olivine but the second phase to crystallize is orthopyroxene in the olivine tholeiite, orthopyroxene + sub-calcic clinopyroxene in the olivine basalt and clinopyroxene in the alkali olivine basalt. This contrasts with the atmospheric pressure sequence of crystallization in which plagioclase is the second phase to crystallize. In addition to the late appearance of plagioclase at 9 kb, the experimental results suggest that the early crystallizing olivine may bear a reaction relationship towards clinopyroxene. In discussion of the experimental results it was noted that there was a decrease in intensity of olivine as clinopyroxene became more abundant in the assemblages. However, olivine does not entirely disappear in any of the compositions, and microprobe analyses show that, in the olivine tholeiite at least, the olivine continues to become more Fe-rich even after crystallization of clinopyroxene commences. Our data therefore suggest that olivine partially reacts with liquid to precipitate aluminous, calcic clinopyroxene. This implies that it is possible to derive residual liquid compositions from these olivine-rich compositions which do not have olivine on the liquidus at 9 kb, but instead have clinopyroxene or clinopyroxene + orthopyroxene as liquidus phases. Furthermore, these derivative liquids, if crystallized at atmospheric pressure would have either olivine or plagioclase, or both, as liquidus phases. These aspects of basalt fractionation are being intensively studied by T. H. Green (Department of Geophysics and Geochemistry, Australian National University) and his results provide a much more complete documentation of the relationships than our present data.

The compositions of residual liquids can be calculated using the analyzed minerals of Tables 8, 9 provided that an estimate can be made of the proportions of crystals present. This can be done approximately by visual methods and the use of polished surfaces offers the possibility of modal analyses either on the specimen itself or on photographic reproductions of the surface. In the present paper we have used the known compositions of olivines and orthopyroxenes crystallized at various temperatures to estimate the proportions of crystals present. In a previous section we have discussed the empirical relationship between the molecular ratio 100 $\mathrm{Mg/Mg+Fe^{++}}$ of a basaltic liquid and the same ratio of its liquidus olivine.

An empirical partition coefficient $\mathrm{K_{O,L}}$ has been defined as

$$\left(\frac{100\,\mathrm{Mg}}{\mathrm{Mg+Fe}}\right)_{\mathrm{Olivine}}\Bigg/\left(\frac{100\,\mathrm{Mg}}{\mathrm{Mg+Fe^{++}}}\right)_{\mathrm{Liquid}}=\mathrm{K_{O,L}}$$

and for $(100\,\mathrm{Mg/Mg+Fe})_{\mathrm{Olivine}}$ between 83 and 92, $K_{\mathrm{O,L}}$ has a mean value of 1.33 at 9 kb. $\mathrm{K_{O,L}}$ increases slightly with decreasing values of $(100\,\mathrm{Mg/Mg+Fe})_{\mathrm{Olivine}}$ but as a first approximation the mean value of 1.33 is used in the following calculations (Table 3).

At 9 kb, the liquidus olivine of the olivine tholeiite at a temperature of about 1,360° C has $100\,\mathrm{Mg/Mg+Fe} = 92$. At a lower temperature, e.g. 1,310° C, the olivine is more abundant and has $(100\,\mathrm{Mg/Mg+Fe})_{\mathrm{Olivine}} = 89.4$. Olivine of this composition is the liquidus phase of the glass present in the run, and assuming the empirical relationship that $\mathrm{K_{O,L}} = 1.33$ the liquid at this temperature has $(100\,\mathrm{Mg/Mg+Fe^{++}})_{\mathrm{Liquid}} = 89.4/1.33 = 67.2$. The first appearance of orthopyroxene in the olivine tholeiite is at 1,290° C, 9 kb and in a similar way the residual liquid at this temperature may be calculated to have the value $(100\mathrm{Mg/+Fe^{++}}) = 65.6$. The ratio [MgO]/[FeO] where [MgO] and [FeO] are expressed in weight percent can then be calculated for the residual liquid

$$\frac{[\mathrm{MgO}]}{[\mathrm{FeO}]}_{\mathrm{Liquid}} = 1.07.$$

Since orthopyroxene is extremely rare in the 1,290° C run the residual liquid must be derived from the original composition by extraction of olivine alone. The olivine at 1,290° C 9 kb contains 12.3% FeO and 47.1% MgO and if the percentage of olivine crystals present is m% then

$$\frac{[\mathrm{MgO}]}{[\mathrm{FeO}]}_{\mathrm{Liquid}} = \frac{14.55 - \frac{m}{100} \times 47.1}{10.07 - \frac{m}{100} \times 12.3} = 1.07.$$

$$\text{i.e. } m = 11.1\%.$$

The composition of the liquid phase, assuming 11% crystallization of olivine as analyzed at 1,290° C, 9 kb, is given in Table 14.

Between 1,290° C and about 1,260° C the olivine tholeiite composition consists of olivine, orthopyroxene and liquid. The first appearance of clinopyroxene was at 1,250° C. Both olivine and orthopyroxene have been analyzed from this run and have $(100\ \mathrm{Mg/Mg+Fe})_{\mathrm{Opx}} = 84.9$ and $(100\ \mathrm{Mg/Mg+Fe})_{\mathrm{Ol}} = 85.3$ respectively. The residual liquid has a calculated $100\ \mathrm{Mg/Mg+Fe^{++}})_{\mathrm{Liquid}} = 64$ but to calculate the degree of crystallization, the relative proportions of olivine and orthopyroxene crystallizing between 1,290° C and 1,250° C must be estimated. If this crystal extract consists of 25% olivine and 75% orthopyroxene then its composition is

SiO_2	49.66%
Al_2O_3	5.23%
FeO	10.38%
MgO	32.83%
CaO	2.10%

and, using the liquid composition determined at 1,290° C, the percentage of olivine + orthopyroxene crystals precipitating between 1,290° C and 1,250° C may be calculated as 3.4%, i.e.

0.9% Olivine
2.5% Orthopyroxene.

This method of estimation of degree of crystallization is potentially very useful, particularly as more data on liquid and liquidus phase partition coefficients become available. Using the estimates of degrees of crystallization obtained in this way as a guide, Table 14 contains the calculated compositions of the liquid phase in the olivine tholeiite at about 1,290° C assuming 11% crystallization of olivine, and about 1,250° C assuming 12% crystallization of olivine and 3% crystallization of orthopyroxene.

The appearance of clinopyroxene at 1,250° C appears to coincide with a relatively rapid increase in the proportion of crystals present. Although no analysis of the clinopyroxene was possible, its composition may well be similar to the clinopyroxenes which co-exist with orthopyroxene at 1,290° C, 13.5 kb and 1,310° C, 13.5 kb in the alkali basalt and olivine tholeiite composition. The clinopyroxene at 9 kb, 1,230° C would probably contain a lower Al_2O_3 content but similar CaO content and similar 100 Mg/Mg + Fe ratio to that from the alkali olivine basalt at 1,290° C, 13.5 kb. In Table 14 an approximate residual liquid composition assuming 30% crystallization (12% olivine, 4% orthopyroxene, 14% clinopyroxene) has been calculated for the olivine tholeiite at 9 kb and about 1,230° C.

Table 14. *Fractionation of the olivine tholeiite at 9 kb*

P, T conditions (kb) (°C)		9 1,290	9 1,250	9 1,230
Nature and estimated percentage of crystals		11% Olivine	12% Olivine 3% Orthopyroxene	12% Olivine 4% Orthopyroxene 14%. Clinopyroxene
Composition of Crystal Extract				
SiO_2		40.3	42.5	45.9
Al_2O_3		—	1.4	6.2
FeO		12.3	13.0	11.3
MgO		47.1	42.3	31.6
CaO		0.3	0.8	5.0
Composition of Liquid Phase				
	(Initial liquid)			
SiO_2	46.95	47.6	47.6	47.3
TiO_2	2.02	2.3	2.4	2.9
Al_2O_3	13.10	14.8	15.2	16.1
Fe_2O_3	1.02	1.1	1.2	1.5
FeO	10.07	9.8	9.6	9.5
MnO	0.15	0.2	0.2	0.2
MgO	14.55	10.5	9.5	7.2
CaO	10.16	11.4	11.8	12.4
Na_2O	1.73	2.0	2.0	2.5
K_2O	0.08	0.1	0.1	0.1
P_2O_5	0.21	0.2	0.2	0.3
CIPW norm of Liquid Phase				
Or	0.67	0.7	0.7	0.8
Ab	14.7 } 42.9	16.5 } 48.3	16.8 } 49.6	21.0 } 54.2
An	27.6	31.1	32.1	32.4
Di	17.0	19.2	20.2	22.5
Hy	12.3	13.8	12.9	5.2
Ol	21.9	12.2	10.5	10.1
Ilm	3.8	4.3	4.6	5.5
Mt	1.4	1.6	1.8	2.1
Ap	0.5	0.5	0.5	0.6
$\frac{100\ Mg}{Mg+Fe^{++}}$	72.1	65.5	64.3	57.2

If the Al_2O_3 content of the clinopyroxene is lower than the value used (cf. the clinopyroxene from the alkali basalt at 9 kb, 1,220° C) the resulting liquid will have higher Al_2O_3 and higher normative anorthite content.

The calculated compositions of Table 14 illustrate the fractionation trend of the olivine tholeiite assuming quite low degrees of crystallization at pressures around 9 kb, i.e. depths of 30—35 kms. The derivative liquids are very distinctive in showing very little change in SiO_2 content and in normative olivine content once fractionation has proceeded to the stage of separation of both pyroxene and olivine (Fig. 8). In contrast, Al_2O_3, CaO and Na_2O are steadily enriched in the residual liquid as fractionation proceeds (Figs. 8, 9) and the resulting compositions have high normative plagioclase contents. The derivative liquids are classifiable as high-alumina olivine tholeiites and closely resemble analyses of high alumina olivine tholeiites, "oceanic tholeiites" or "high-alumina basalts" given by YODER and TILLEY (1962), ENGEL, ENGEL and HAVENS (1965) and KUNO (1960). It must be emphasized that the enrichment in CaO, Al_2O_3 and Na_2O is caused by the absence of plagioclase as an early-crystallizing phase and by the co-precipitation of olivine and orthopyroxene, maintaining an olivine normative character and essentially constant SiO_2 content. The appearance of clinopyroxene does not greatly modify this trend *provided* that the clinopyroxene is sub-calcic (i.e. co-existing with orthopyroxene) and of low or moderate Al_2O_3 content. The increasing proportion of low-calcium pyroxenes extracted as fractionation proceeds results in decreasing normative hypersthene — in terms of the basalt tetrahedron (YODER and TILLEY, 1962), residual liquids from the olivine tholeiite fractionate rapidly towards the plagioclase apex of the olivine tholeiite volume and also fractionate slightly towards the Cpx-Ol-Plag plane of critical undersaturation (Fig. 1).

In the alkali olivine basalt composition, the analyses of both olivine and clinopyroxene at 9 kb 1,220° C can be used to calculate a possible residual liquid composition (Table 15) assuming 15% crystallization (5% olivine, 10% clinopyroxene). As with the olivine tholeiite, fractionation produces little change in SiO_2 content but an increase in Al_2O_3, Na_2O and normative plagioclase contents, and a decrease in $100\ Mg/Mg + Fe^{++}$ ratio (Figs. 8, 9). The absence of orthopyroxene and more calcic nature of the clinopyroxene prevent marked enrichment in CaO content though it must be pointed out that a higher proportion of olivine to clinopyroxene than that assumed in Table 15 would result in more marked increase in Al_2O_3 content and increasing SiO_2 and CaO contents. In terms of normative minerals, the separation of olivine and clinopyroxene results in increased nepheline and derivative liquids remain critically undersaturated with low or moderate degrees of fractionation.

The analytical data on phases crystallized from the olivine basalt composition at 9 kb are insufficient for a quantitative discussion of fractionation. However, the appearance of both orthopyroxene and clinopyroxene with olivine in the 9 kb, 1,260° C run will cause derivative liquids to move towards high Al_2O_3, CaO, and Na_2O contents while maintaining approximately constant SiO_2. Derivative liquids would almost certainly become nepheline-normative with sufficient fractionation. The absence of orthopyroxene in the 1,240° C, 9 kb run indicates a more calcic clinopyroxene and fractionation beyond this temperature would probably show decreasing CaO content.

In general, analyses of natural high-Al_2O_3 basalts in the literature have higher SiO_2 contents than the calculated compositions. Closer similarity in this component would be produced if a slightly greater proportion of olivine crystallized before the appearance of orthopyroxene and we anticipate that this would happen in the olivine tholeiite at slightly lower pressures than 9 kb. We anticipate a transition with increasing pressure between the low pressure fractionation trend to quartz tholeiite without Al_2O_3 enrichment and the higher pressure trend to high-alumina olivine tholeiites or high-alumina alkali olivine basalts. Increasing pressure will enlarge the temperature interval between the appearance of olivine and plagioclase and will also gradually extend the field of orthopyroxene crystallization into olivine-normative (particularly with $Ol > 15\%$) compositions. Thus

Table 15. *Fractionation of alkali olivine basalt at 9 kb*

P, T conditions (kb) (°C)	9 1,270		9 1,220	
Nature and estimated percentage of crystals	Nil		5% Olivine 10% Clinopyroxene	
Composition of crystal extract				
SiO_2			46.0	
Al_2O_3			6.0	
FeO			12.3	
MgO			25.5	
CaO			10.2	
Composition of liquid phase				
	(Initial liquid)			
SiO_2	45.39		45.3	
TiO_2	2.52		3.0	
Al_2O_3	14.69		16.2	
Fe_2O_3	1.87		2.2	
FeO	12.42		12.4	
MnO	0.18		0.2	
MgO	10.37		7.7	
CaO	9.14		9.0	
Na_2O	2.62		3.1	
K_2O	0.78		0.9	
P_2O_5	0.02		—	
CIPW norm				
Or	4.5	48.7	5.5	52.0
Ab	18.0		19.0	
An	26.2		27.5	
Ne	2.2		3.9	
Di	15.7		14.2	
Ol	25.8		21.0	
Ilm	4.8		5.7	
Mt	2.9		3.2	
$\frac{100\text{ Mg}}{\text{Mg}+\text{Fe}^{++}}$	59.8		52.5	

the same parent magma may fractionate to yield high-alumina basalts of slightly different SiO_2 contents and normative olivine and hypersthene contents depending on the particular pressure in the 5—10 kb range at which fractionation occurs.

Kuno (1960, p. 125) considers that high-alumina basalts are chemically transitional between tholeiites and alkali basalts, except in Al_2O_3 content. Yoder and Tilley (1962, p. 416—417) also note that high-alumina basalts are not restricted to either the tholeiites or alkali basalt and include analyses of hawaiites within the group. Our data suggests that at 8—10 kb pressure there is no well defined thermal divide close to the Ol-Cpx-Plag plane of critical undersaturation but rather this fractionation trend runs obliquely to this "plane" and directly towards plagioclase enrichment. Thus, while a "parental" olivine tholeiite with low normative hypersthene content may fractionate with moderate degree of crystallization to yield nepheline-normative, high-Al_2O_3 residual liquids, parental olivine tholeiites with higher normative hypersthene are unlikely to yield nepheline-normative residual liquids but would produce high-Al_2O_3, olivine tholeiites (5—15% normative olivine).

c) Fractionation at 35—70 km Depth

The experimental partial melting runs carried out at 13.5 and 18 kb provide the data on which the following discussion is based. The liquidus phases in all compositions studied, except the picrite[4], are pyroxenes, and orthopyroxene plays a dominant role in the early stages of fractionation within this depth interval. Clinopyroxene appears early in the fractionation sequence and, if accompanied by orthopyroxene, is a very sub-calcic variety. Both clinopyroxene and orthopyroxene are aluminous. Garnet does not appear above the solidus at 13.5 kb but appears late in the crystallization sequence at 18 kb. Spinel and plagioclase appear near the solidus at 13.5 kb but are absent at 18 kb.

The compositions of derivative liquids from the basaltic compositions can be calculated using the analyzed minerals of Tables 9, 11 provided that an estimate can be made of the proportions of crystals present. The mineral analysis data show that co-existing olivine and orthopyroxene have almost identical ratios of 100 Mg/Mg + Fe, at least for values of this ratio between 85 and 90 (Table 10). Also, where the compositions have orthopyroxene as the liquidus phase, this has a similar 100 Mg/Mg + Fe ratio to the liquidus olivine observed at lower pressure. For these reasons it is possible to use the empirical partition coefficient

$$\left(\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}}\right)_{\mathrm{Opx}} \Big/ \left(\frac{100\ \mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}^{++}}\right)_{\mathrm{Liquid}} = 1.33$$

to obtain an estimate of the degree of crystallization in the olivine tholeiite (cf. pages 143—144). At 13.5 kb, orthopyroxene is the only phase present on the liquidus and very uncommon clinopyroxene first appears in the 1,350° C run. The calculated proportion of orthopyroxene crystallized between the liquidus and the first appearance of clinopyroxene is 14.5%, using the composition of the orthopyroxene in the 1,350° C, 13.5 kb run in the calculations. This is in good agreement with the amount of orthopyroxene observed optically. At 18 kb, clinopyroxene is absent in the 1,425° C and 1,420° C runs but moderately common in the 1,400° C run. Assuming separation of orthopyroxene (as analyzed at 1,400° C, 18 kb) alone between the liquidus and 1,400° C the percentage of orthopyroxene crystallized may be calculated as 12.5%. This is likely to be an over-estimate as clinopyroxene probably appears at about 1,410° C.

In Table 16 we have calculated derivative liquid compositions assuming 15% crystallization of orthopyroxene alone at 13.5 kb and crystallization of 10% orthopyroxene + 5% clinopyroxene at 18 kb. The derivative liquids are very similar in chemistry and in normative mineralogy although the extraction of

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[4] **Orthopyroxene occurred as the liquidus phase in the picrite at 1,325° C, 18 kb in the preliminary runs reported by GREEN and RINGWOOD, 1964. The absence of olivine presumably resulted from the access of water, possibly with some resulting oxidation, and suppression of the liquidus temperature by about 100° C.**

sub-calcic clinopyroxene at 18 kb results in a residual liquid with slightly higher hypersthene and lower anorthite content. Both compositions lie very close to the "critical plane of undersaturation" in the basalt tetrahedron (Fig. 1) and are olivine-rich basalts of the type transitional between olivine tholeiite and alkali olivine basalt. In chemistry and normative mineralogy the calculated compositions are very close to the "olivine basalt" composition chosen for detailed study and further fractionation beyond 15% crystallization can be best interpreted from the experimental results on this and the alkali olivine basalt. However it is worth noting that crystallization below 1,350° C at 13.5 kb in the olivine tholeiite involves extraction of both orthopyroxene and co-existing sub-calcic clinopyroxene. Thus the nature of the crystal extract remains rich in SiO_2 and in hypersthene molecule, and derivative liquids will move into the nepheline normative field. A similar relationship holds for the 1,380° C and 1,375° C runs at 18 kb. The disappearance of orthopyroxene in the 1,370° C 18 kb run would result in extraction of much higher CaO content in the crystalline phase and the appearance of garnet in the 1,350° C run would further divert the fractionation trend away from the direct path from hypersthene-normative to nepheline-normative olivine basalts.

In the olivine basalt composition at 13.5 kb there is a temperature interval of 10—20°C over which orthopyroxene alone separates from the liquid. Our data are insufficient to use the pyroxene compositions to estimate the degree of crystallization at 13.5 kb and instead we have calculated the derivative liquid composition assuming 10% crystallization of orthopyroxene as indicated by optical examination of polished surfaces. At 18 kb the orthopyroxene co-existing with liquid at 1,340°C has a lower $100\,Mg/Mg+Fe$ ratio (89.0) than the orthopyroxene co-existing with clinopyroxene + liquid in the lower temperature and more extensively crystallized run at 1,335°C. This is apparently in conflict with the empirical partition relationship between liquid and orthopyroxene discussed previously. However the co-existing clinopyroxene at 1,335°C has $100\,Mg/Mg+Fe = 88.2$ so that the mean value of this ratio for the crystalline phase is <89.0. It is clear however that very little orthopyroxene may be extracted at 18 kb from the olivine basalt before the appearance of subcalcic clinopyroxene. In Table 17 we have calculated a residual liquid at 18 kb assuming crystallization of 5% orthopyroxene + 10% clinopyroxene as analyzed at 1,335°C, 18 kb. The residual liquids calculated for 10% crystallization at 13.5 kb and 15% crystallization at 18 kb are very closely similar. Both show decreased SiO_2 content from the original liquid and both are nepheline normative with over 2% nepheline and 24% normative olivine. In both chemistry and normative mineralogy these compositions are very close to the alkali olivine basalt composition and the experimental results on the latter composition may be used to further investigate the fractionation trend beyond 10 and 15% crystallization. The appearance of minor orthopyroxene at the liquidus of the alkali olivine basalt composition at 13.5 kb, supports the estimate of 10% orthopyroxene crystallized from the olivine basalt at 13.5 kb. These two compositions effectively demonstrate that orthopyroxene may crystallize from nepheline-normative magmas at moderate pressure and that in the chosen compositions it is the dominant role of orthopyroxene which determines the liquid fractionation trend from olivine-rich tholeiite to alkali olivine basalt.

Table 16. *Fractionation of olivine tholeiite at 13.5 kb and 18 kb*

P, T conditions (kb)		13.5	18
(°C)		1,350	1,400
Nature and estimated percentage of crystals		15% Orthopyroxene	10% Orthopyroxene + 5% Clinopyroxene
Composition of crystal extract			
SiO_2		53.8	53.0
Al_2O_3		6.0	7.3
FeO		7.2	6.3
MgO		30.2	28.6
CaO		2.8	4.8
Composition of liquid Phase			
	(Initial liquid)		
SiO_2	46.95	45.7	45.9
TiO_2	2.02	2.4	2.4
Al_2O_3	13.10	14.3	14.1
Fe_2O_3	1.02	1.2	1.2
FeO	10.07	10.6	10.7
MnO	0.15	0.2	0.2
MgO	14.55	11.9	12.1
CaO	10.16	11.3	11.1
Na_2O	1.73	2.0	2.0
K_2O	0.08	0.1	0.1
P_2O_5	0.21	0.2	0.2
CIPW norm of liquid Phase			
Or	0.6	0.6	0.6
Ab	14.7 } 42.9	17.3 } 47.6	17.3 } 46.7
An	27.6	29.7	28.8
Di	17.0	20.5	20.3
Hy	12.3	0.2	2.0
Ol	21.9	24.9	24.2
Ilm	3.8	4.5	4.6
Mt	1.4	1.8	1.8
Ap	0.5	0.5	0.5
$\frac{100\,Mg}{Mg+Fe^{++}}$	72.1	66.7	66.9

The residual liquids from the alkali olivine basalt, assuming 10% crystallization of orthopyroxene + clinopyroxene at 13.5 kb and 20% crystallization of clinopyroxene at 18 kb (Table 18) suggest that derivatives with higher normative nepheline contents, approaching olivine-rich basanites, may be produced. If both orthopyroxene and clinopyroxene crystallize together, the CaO content of derivative liquids remain roughly constant and the derivative liquids show increasing nepheline and diopside contents and decreasing albite and anorthite contents. Continuation of this trend appears appropriate for derivation of basanites. However, if orthopyroxene is absent, extraction of calcic clinopyroxene results in decreasing CaO content and atypical basaltic chemistry. The change in normative mineralogy is towards lower diopside content but the ratio of nepheline to albite

Table 17. *Fractionation of olivine basalt at 13.5 kb and 18 kb*

P, T conditions (kb)		13.5	18
(°C)		1,310	1,335
Nature and estimated percentage of minerals		10% orthopyroxene (as at 1,320° C, 13.5 kb)	5% orthopyroxene 10% clinopyroxene
Composition of crystal extract			
SiO_2		53.8	51.9
Al_2O_3		6.4	8.6
FeO		6.6	5.8
MgO		30.3	25.6
CaO		2.9	8.1
Composition of residual liquid			
	(Initial liquid)		
SiO_2	47.10	46.4	46.3
TiO_2	2.31	2.6	2.7
Al_2O_3	14.15	15.0	15.1
Fe_2O_3	0.42	0.5	0.5
FeO	10.64	11.1	11.5
MnO	0.16	0.2	0.2
MgO	12.71	10.8	10.4
CaO	9.86	10.6	10.2
Na_2O	2.21	2.5	2.6
K_2O	0.44	0.5	0.5
CIPW norm of residual liquid			
Or	2.7	3.0	3.3
Ab	18.9	16.8	17.8
Ne	—	2.4	2.3
An	27.3	28.1	27.8
Di	17.6	19.9	18.6
Hy	1.3	—	—
Ol	27.2	24.2	24.3
Ilm	4.4	5.0	5.1
Mt	0.6	0.7	0.8
$\frac{100\,\mathrm{Mg}}{\mathrm{Mg}+\mathrm{Fe}^{++}}$	68.1	63.4	61.8

with 20% clinopyroxene extracted is almost the same as for 10% (orthopyroxene + clinopyroxene) extraction. Extraction of clinopyroxene alone (if this is similar in composition to that analyzed at 1,335° C, 18 kb) appears likely to yield compositions unlike natural basanites or olivine nephelinites. The possible derivation by fractionation of these more extremely undersaturated liquids is currently being studied and factors such as the appearance of spinel or garnet and the Na_2O content of the liquidus clinopyroxenes are potentially important in this problem.

Without microprobe analyses of the phases crystallizing from the picrite, it is not possible to quantitatively discuss the fractionation of this composition. However the normative composition is such that the picrite is essentially the same as the olivine basalt composition with the addition of about 9% more olivine. The

Table 18. *Fractionation of alkali olivine basalt at 13.5 kb and 18 kb*

P, T conditions (kb)		13.5	18
(°C)		1,280	1,310
Nature and estimated percentage of minerals		2.5% orthopyroxene + 7.5% clinopyroxene (as at 13.5 kb, 1,290° C)	20% clinopyroxene (as at 18 kb, 1,320° C)
Composition of crystal extract			
SiO_2		49.6	49.1
Al_2O_3		11.0	11.0
FeO		9.2	7.6
MgO		22.3	19.5
CaO		7.9	12.8
Composition of residual liquid			
	(Initial liquid)		
SiO_2	45.39	44.9	44.5
TiO_2	2.52	2.8	3.2
Al_2O_3	14.69	15.1	15.6
Fe_2O_3	1.87	2.1	2.3
FeO	12.42	12.8	13.6
MnO	0.18	0.2	0.2
MgO	10.37	9.1	8.1
CaO	9.14	9.3	8.2
Na_2O	2.62	2.9	3.3
K_2O	0.78	0.9	1.0
P_2O_5	0.02	—	—
CIPW norms of residual liquids			
Or	4.5	5.5	6.1
Ab	18.0	15.7	17.4
Ne	2.2	4.9	5.6
An	26.2	25.3	24.8
Di	15.7	17.1	13.2
Hy	—	—	—
Ol	25.8	23.2	23.6
Ilm	4.8	5.3	6.1
Mt	2.9	3.0	3.3
$\frac{100\,Mg}{Mg+Fe^{++}}$	59.8	56.1	51.5

melting runs at 13.5 kb and 18 kb are in excellent agreement with this conclusion as olivine is the liquidus phase at both pressures and following precipitation of a moderate amount of olivine, orthopyroxene followed closely by clinopyroxene separates from the residual liquid. We conclude that initially the picrite would fractionate by separation of olivine to yield a composition with $25 \pm 3\%$ normative olivine, 4% normative hypersthene. Further fractionation by separation of orthopyroxene or of orthopyroxene + clinopyroxene accompanying the olivine would yield nepheline normative liquids closely analogous to those derived from the olivine basalt composition.

The appearance of orthopyroxene rather than olivine as the liquidus phase at 18 kb, 1,325°C under "wet" melting conditions (GREEN and RINGWOOD, 1964) suggests the possibility that low partial pressures of water may suppress the temperature of appearance of olivine to a greater degree than orthopyroxene. This aspect of high pressure fractionation needs to be further explored experimentally.

d) Fractionation at about 70—100 km Depth

The experimental results at 22.5 kb and 27 kb are relevant to discussion of fractionation below 70 km depth. At 22.5 kb, clinopyroxene is the liquidus phase in all three basaltic compositions and also in the picrite. By analogy with the clinopyroxene in the alkali basalt at 20 kb, such clinopyroxenes are likely to contain CaO contents greater than the liquid, high Al_2O_3 content, and may also contain moderate Na_2O as jadeite solid solution. Fractionation by separation of clinopyroxene at this pressure would produce lower SiO_2 contents, lower CaO contents and lower 100 Mg/Mg + Fe ratios. Depending on the Na_2O content of the clinopyroxene, hypersthene normative liquids could possibly yield derivative nepheline-normative liquids but high degrees of crystallization would be required and resultant liquids would be atypical of basalt in having low CaO contents.

At 27 kb, garnet is the liquidus phase in the olivine basalt and alkali olivine basalt compositions and possibly garnet alone occurs on the liquidus of the olivine tholeiite at 27 kb. In all three cases, there is a relatively small drop in temperature before the appearance of clinopyroxene. The melting interval in all compositions and particularly in the picrite is quite small at 27 kb so that quite large amounts of a phase may precipitate over a small temperature interval.

We have calculated a residual liquid composition (Table 19) assuming 10% crystallization of garnet from the olivine basalt and using the composition of the garnet analyzed from the 27 kb, 1,460°C run. The effect of garnet extraction from this particular composition is to produce a nepheline-normative (0.6% Ne) liquid. This result supports the conclusions of YODER and TILLEY (1962) that the plane of critical undersaturation is not a thermal divide at 27—30 kb and, in particular, that extraction of garnet may produce liquids of alkali basalt type. However, separation of 10% garnet without the appearance of clinopyroxene is rather doubtful in the compositions studied, particularly in the olivine tholeiite where clinopyroxene possibly occurs with garnet at the liquidus. Extraction of garnet results in increased SiO_2 content and particularly in decreased Al_2O_3 content and is not as "efficient" as aluminous orthopyroxene in developing nepheline-normative residual liquids, e.g. while extraction of 10% garnet from the olivine basalt at 27 kb produces a residual liquid with 0.6% nepheline, extraction of 10% aluminous orthopyroxene at 13.5 kb produces a liquid with 2.4% normative nepheline. Extraction of garnet with its high Al_2O_3/SiO_2 ratio causes a decrease in normative anorthite and this imposes a rather severe restriction on the amount of garnet which can be extracted while retaining a basalt-like chemistry and normative mineralogy.

Calculations of the effects of extracting 10% garnet from the alkali-olivine basalt demonstrate an increase in normative nepheline content (from 2.2% to 4.0%) and decrease in normative anorthite content (from 26.2% to 21.5%). Similar

Table 19. *Fractionation of olivine basalt at 27 kb*

P, T conditions (kb)		27
(°C)		1,460
	Composition of initial liquid	Residual liquid after extraction of 10% garnet (1,460°C, 27 kb)
SiO_2	47.10	47.6
TiO_2	2.31	2.5
Al_2O_3	14.15	13.0
Fe_2O_3	0.42	0.5
FeO	10.64	10.9
MnO	0.16	0.2
MgO	12.71	12.0
CaO	9.86	10.3
Na_2O	2.21	2.5
K_2O	0.44	0.5
Norms		
Or	2.7	3.1
Ab	18.9	20.0
Ne		0.6
An	27.3	22.7
Di	17.6	23.0
Hy	1.3	—
Ol	27.2	25.2
Ilm	4.4	4.7
Mt	0.6	0.7
$\frac{100\,Mg}{Mg + Fe^{++}}$	68.1	66.3

extraction of 10% garnet from the olivine tholeiite decreases the normative hypersthene from 12.3% to 10.1% and normative anorthite from 27.6% to 23.3%. To obtain a nepheline normative liquid from the olivine tholeiite would require separation of more than 30% garnet without the appearance of clinopyroxene. This requirement is in direct conflict with our experimental data. Furthermore, it would produce non-basaltic residual liquids.

In the picrite composition at 27 kb, garnet, orthopyroxene and clinopyroxene all occur very close to the liquidus and the direction of fractionation cannot be estimated. Olivine is not a liquidus phase in the picrite at 27 kb suggesting that this composition is not appropriate as a "minimum melting liquid" derived by partial melting of mantle peridotite. We would anticipate that such a liquid would have olivine, with orthopyroxene or possibly clinopyroxene, as liquidus phases — these being the major phases of the residual peridotite with which the liquid would be in equilibrium.

We conclude that separation of garnet or a garnet-rich mixture of garnet + clinopyroxene may produce nepheline normative liquids from compositions rich in olivine but poor in hypersthene. Such fractionation, if significant, must be restricted in basaltic rocks as extraction of garnet with its high-Al_2O_3 content leads to residual liquids inconsistent with basaltic chemistry. If this process is significant in producing liquids (at 27 kb or similar pressures) which are parental to the low pressure divergent alkali olivine basalt and olivine tholeiite sequences, then in general alkali olivine basalts should have lower Al_2O_3/SiO_2 ratios than olivine tholeiites of similar Al_2O_3 or SiO_2 content. This is the reverse of the observed situation cf. MACDONALD and KATSURA (1964, Table 9, p. 124) and the association of higher Al_2O_3 and lower SiO_2 contents with alkali olivine basalts relative to olivine tholeiites of similar $100\ Mg/Mg + Fe^{++}$ ratios, is consistent with fractionation control by aluminous orthopyroxene rather than garnet.

Discussion of the role of garnet in the basaltic compositions at 27—30 kb may be irrelevant if these particular compositions are appreciably divergent from the composition of liquids which are partial melting derivatives of a mantle peridotite composition and thus in equilibrium with residual crystals of that peridotite. The presence of clinopyroxene and garnet on or very near the liquidus in these compositions suggests that at 27 kb these liquids could be in equilibrium with residual eclogite — rather than residual peridotite or dunite. Experimental work in progress in this laboratory has shown that garnet is unstable at 27 kb, 1,500°C in several peridotite compositions which are appropriate chemically for derivation of basaltic magmas leaving residual dunite or enstatite peridotite. Such compositions contain olivine + orthopyroxene + clinopyroxene ± spinel at the solidus at 27 kb and liquids derived from them by partial melting at this pressure may be expected to have one or more of these phases on the liquidus (cf. pages 160, 166, 167). It appears that the picrite composition may be closer to the composition of liquids

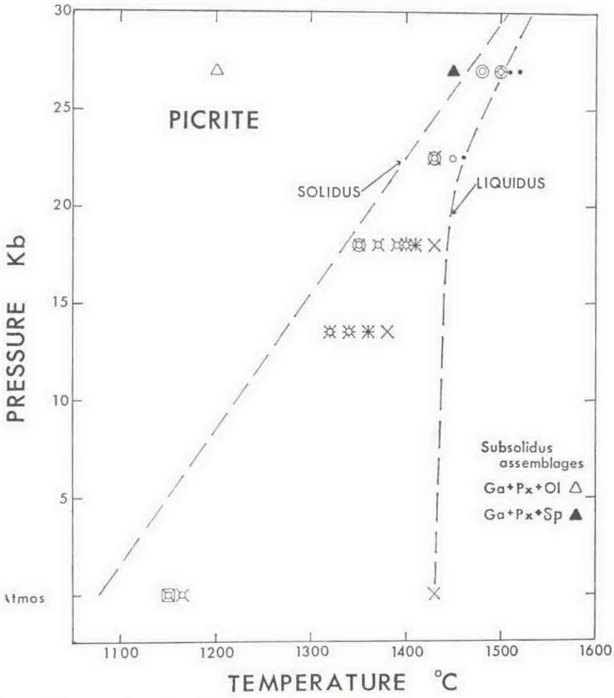

Fig. 7. Detail of melting and subsolidus experiments carried out on the picrite composition at high pressure. Solidus and liquidus positions are approximate only. Data at 1 Atmosphere is entirely interpreted from the literature (TILLEY, YODER and SCHAIRER, 1963, 1964, 1965). Symbols are the same as those of Figs. 4 and 6

which may be derived from parental peridotite at 27 kb, particularly if in fact orthopyroxene is the liquidus phase at 27 kb, 1,500°C $< T <$ 1,510°C.

e) Summary of Fractionation Trends

The data and discussions of previous sections demonstrate that an olivine-rich tholeiite magma may fractionate to yield distinctive basaltic magma types at different pressures. These fractionation trends are depicted in Figs. 8—10.

(i) *Low Pressure or Shallow Crustal Fractionation* ($<$15 km depth). Fractionation under these conditions is dominated by olivine as the liquidus phase with clinopyroxene and/or plagioclase appearing at lower temperatures. The field of olivine + liquid in compositions with 20—25% normative olivine may cover a wide temperature interval and, because of the Bowen reaction relationship, olivine remains the liquidus phase, in many compositions, into the quartz-normative tholeiite field. Thus, in the olivine tholeiites, olivine may show a reaction relationship with liquid, and fractionation of olivine-rich tholeiitic liquids (such as the olivine tholeiite) may yield quartz-normative residual liquids.

(ii) Fractionation at about 15—35 km Depth. Early stages of fractionation are dominated by olivine as at low pressure but the olivine is joined by orthopyroxene as the second phase in the olivine tholeiite composition and possibly the olivine basalt composition and by clinopyroxene in the alkali olivine basalt composition. Plagioclase only appears at temperatures very near the solidus. The crystallization of relatively large proportions of olivine and pyroxene(s) *before* the appearance of plagioclase contrasts with the crystallization sequence at low pressure and results in a marked increase in Al_2O_3 while retaining a relatively constant SiO_2 content at about 47—49% SiO_2. In this load pressure or depth interval an olivine tholeiite will not fractionate to yield an oversaturated quartz-tholeiite but instead will fractionate to yield high-Al_2O_3 olivine tholeiites (5—10% normative olivine).

(iii) Fractionation at about 35—70 km Depth. In this depth interval, olivine does not appear as the liquidus phase except in the picrite composition. Fractionation is dominated by separation of aluminous orthopyroxene or orthopyroxene + subcalcic augite. In either case magnesium-iron metasilicate is the major component removed with lesser but important amounts of alumina and calcium metasilicate. Extraction of these pyroxenes results in decrease in SiO_2 content, increase in $Na_2O + K_2O$ content and smaller percentage increases in Al_2O_3 and CaO contents. The fractionation trend yields olivine-rich, alkali basaltic magmas (i.e. nepheline normative) from parental olivine-rich tholeiitic magmas, and is in direct contrast to the shallow crustal fractionation trend of olivine tholeiites towards olivine-poor or quartz tholeiitic magmas.

(iv) Fractionation at about 70—100 km Depth. The appearance of garnet as the liquidus phase, without clinopyroxene, means that separation of garnet alone may occur over a very limited temperature interval and result in a fractionation trend producing higher SiO_2 + alkalis, lower Al_2O_3, MgO + FeO and approximately constant CaO in derivative liquids. The early appearance of clinopyroxene accompanying the garnet and the separation of garnet + clinopyroxene would change the fractionation trend, probably retaining an overall basaltic chemistry for the residual liquids but leading to increased FeO/MgO, increased K_2O, TiO_2 and possibly Na_2O, depending on the Na_2O content of the clinopyroxene.

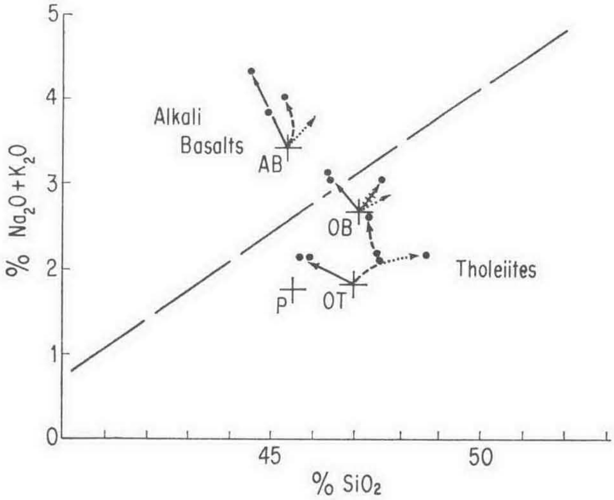

Fig. 8. Total alkalies vs silica plot for the compositions used in the experimental melting study. The line dividing the alkali basalt from the tholeiite fields is from MACDONALD and KATSURA (1964). *AB* alkali olivine basalt composition (Table 1); *OB* olivine basalt composition (Table 1); *OT* olivine tholeiite composition (Table 1); *P* Picrite (Table 1). ● Calculated composition (Tables 12—19). ·······> Fractionation trend at atmospheric pressure; ----> fractionation trend at about 9 kb (30 km); ——> fractionation trend at about 13—18 kb (40—60 km); -|-|-|-> fractionation trend at about 27 kb (100 km)

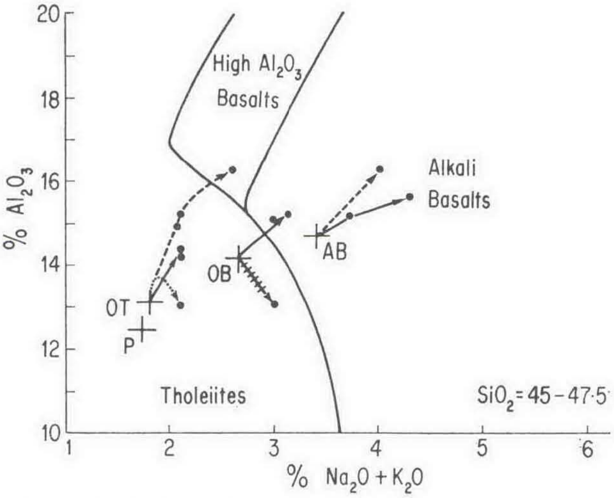

Fig. 9. Diagram for total alkalies v alumina for the compositions used in the experimental study and for the calculated fractionation trends of Tables 12—19. Basalt field boundaries from KUNO (1960). Symbols as in Fig. 8

If basaltic magmas in general are derived by partial melting of a peridotitic mantle at a depth of 100 to 150 km and segregate from residual unmelted peridotite at a depth of 100 km or less then the above diverse fractionation would argue for the recognition of one "primary" basaltic magma (an olivine-rich and hypersthene-normative composition) and three derivative magma types of similar status:

(a) the alkali olivine basalt magma,

(b) the high alumina olivine tholeiite magma,

(c) the olivine tholeiite and quartz tholeiite magmas of low alumina content.

It must be emphasized that no sharp division between these magma types would be anticipated but rather a continuity in compositions since all may have the same parental liquid. The experimental data provide support for the concept that high-alumina basalts, like alkali olivine basalts, require recognition as a specific magma type indicative of origin or fractionation with a defined depth interval of the mantle cf. Tilley (1950), Kuno (1960), Green, Green and Ringwood (1967). Rocks derived from these two magma types, and also those derived from a "parental" olivine tholeiite magma, most commonly reflect the effects of low pressure fractionation super-imposed on the chemistry of the liquid actually derived from the mantle. Low pressure fractionation of alkali olivine basalts will only increase the divergence of these liquids from liquids of the olivine or quartz tholeiite type (cf. pages 107—111), thus facilitating recognition of the type of magma derived from the mantle. In contrast, high alumina olivine tholeiites, if allowed to fractionate at shallow depth, will do so by precipitation of plagioclase, olivine and clinopyroxene. Derivate liquids, if separated, will move towards tholeiites with lower Al_2O_3 and finally into the quartz tholeiite field. The effect of low pressure fractionation on high-alumina olivine tholeiites and on "parental" olivine-rich tholeiite is to produce a convergence of derivative liquids, making identification of the nature of the mantle-derived magma more difficult.

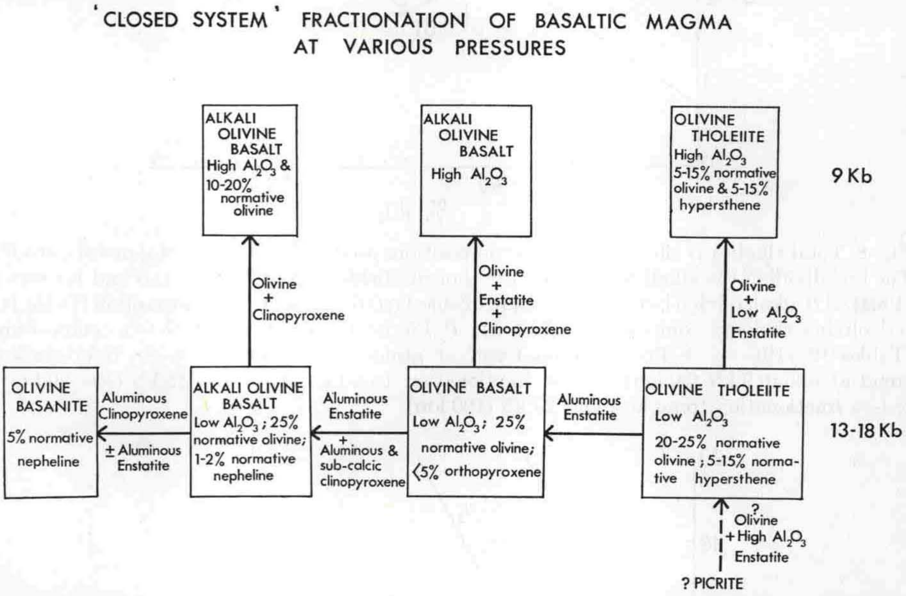

Fig. 10. Diagrammatic summary of the effects and directions of fractionation of basaltic liquids at moderate to high pressures. Refer also to Figs. 8 and 9 for plots of fractionation trends on the conventional "alkalies vs silica" and "alkalies vs alumina" diagrams

Consideration of low pressure fractionation of basaltic magmas leads to a concept of two-fold division of magmas into nepheline-normative and hypersthene-normative types separated by a thermal divide (Yoder and Tilley, 1962). The study of fractionation at high pressure accentuates the transitional and continuous character of chemical variation in magma types but delimits three distinctive trends of fractionation characteristic of certain pressure intervals. These three trends require consideration in any petrogenetic classification of basaltic rocks.

The Generation of Basaltic Magmas

In the previous sections we have discussed the fractionation of basaltic magmas at various depths in the mantle and demonstrated mechanisms by which a "primitive" olivine-rich tholeiite magma may produce derivative liquids of alkali olivine basalt, high-alumina basalt or quartz tholeiite type. An important alternative hypothesis (e.g. Kuno, 1960; Kushiro and Kuno, 1963) maintains that the compositions of the principal basaltic magmas are determined by the depth in the mantle at which partial melting occurs rather than by subsequent fractionation processes. The data obtained from our experiments are used in the following sections to evaluate this hypothesis in detail.

An essential prerequisite to a discussion of the generation of basaltic magmas in the mantle is a consideration of the mineralogical and chemical constitution of the upper mantle. We will take up this subject in the next section, proceed then to the physical conditions of magma formation and finally investigate the chemical and mineralogical equilibria involved and their effect upon the composition of the resultant magma.

a) Chemistry and Mineralogy of Parental Mantle

It can be argued plausibly on general petrological and geochemical grounds that the chemical composition of the primary undifferentiated upper mantle should be somewhere between those of typical basalt and typical alpine peridotite. A chemical and petrological model for the upper mantle based upon this postulate has been developed by Ringwood (1962a, b; 1966a, b), Green and Ringwood (1963) and Green (1966a). In this model, the primary undifferentiated composition of the upper mantle is assumed equal to approximately 1 part of basalt to 3 parts of peridotite. This primary composition is called pyrolite (pyroxene-olivine rock). It is emphasized that the 3:1 proportion is not regarded as critical or unique, and substantial variations in this ratio are possible. Nevertheless an approximate 3:1 ratio is suggested by certain geochemical considerations and is convenient for the formulation of a specific model. The composition of pyrolite as derived by Ringwood (1966a) is given in Table 20.

An important property of compositions close to pyrolite is the ability to crystallize in four distinct mineralogical assemblages over the range of P, T conditions existing in the upper mantle. These are

1. Olivine + amphibole ± enstatite ± spinel (Ampholite);
2. Olivine + pyroxenes + plagioclase + chromite (Plagioclase pyrolite);
3. Olivine + aluminous pyroxenes ± spinel (Pyroxene pyrolite);
4. Olivine + pyroxenes + garnet (Garnet pyrolite).

The stability fields of these mineral assemblages for the model pyrolite composition given in Table 20 are currently under experimental investigation. A preliminary outline of the pyrolite stability fields was given by Ringwood, MacGregor and Boyd (1964), and Ringwood (1966b). Our latest experimental results require some modifications of the earlier boundaries of pyrolite stability fields but these are not of a fundamental nature. Much of the revision is caused by the necessity to introduce a pressure correction to earlier results because of non-uniform distribution of pressure in the furnace assemblies (Green et al., 1966). A provisional outline of stability fields for the pyrolite composition (Table 20) according to our latest experimental results is given in Fig. 11.

Table 20
Model composition of pyrolite (Ringwood, 1966a)

SiO_2	45.16
TiO_2	0.71
Al_2O_3	3.54
Fe_2O_3	0.46
FeO	8.04
MnO	0.14
MgO	37.47
CaO	3.08
Na_2O	0.57
K_2O	0.13
Cr_2O_3	0.43
NiO	0.20
P_2O_5	0.06

For the present purposes, it is important to observe that the boundary between the stability fields of pyroxene pyrolite and garnet pyrolite intersect the pyrolite solidus at a depth of 100 km. *Thus, the formation of magmas by fractional melting of pyrolite at depths smaller than 100 km would occur in the stability field of olivine + aluminous pyroxenes. Garnet would not play a significant role in the genesis of magmas by fractional melting at depths smaller than 100 km.* These considerations constitute a serious objection to the views of Yoder and Tilley (1962), O'Hara (1965) and others, that the principal basalt magma types are formed by direct partial melting of "garnet peridotite" in the upper mantle. In order to stabilise garnet at shallower depths in the mantle, a pyrolite composition possessing a much higher ratio of R_2O_3 ($Al_2O_3 + Cr_2O_3 + Fe_2O_3$) to total pyroxene than appears reasonable would need to be assumed. Furthermore, a model invoking such a composition would encounter further difficulties because of the tendency for garnet to melt incongruently to aluminous enstatite at an early stage of fractional melting in ultramafic compositions, so that magmas produced at relatively low pressures (15—30 kb) are in equilibrium with residual olivine and aluminous pyroxene for a very wide range of possible mantle compositions.

b) Physical Processes of Magma Generation

The formation of a magma in the mantle requires the supply of a large amount of thermal energy, in excess of 100 cals per gram of magma, to a localised region. Physical processes, e.g. thermal conduction, radioactive heat generation, mass transfer, which might be responsible for the supply of this energy operate on a comparatively long time scale. In contrast, the time scale required for separation of crystals from liquid within the mantle directly by gravity or indirectly by deformational processes ultimately of gravitational origin, is probably smaller by orders of magnitude. Because of these conditions, the formation of magmas in the mantle will almost always be the result of *partial* melting rather than of complete melting. Where a substantial degree of partial melting occurred throughout a large volume, the magma will tend to segregate from residual crystals into a self-contained magma body which thereafter evolves independently of the refractory residuum with which it was formerly associated. The degree of fractional melting which is required before the magma separates from residual crystals doubtless varies according to physical conditions, but perhaps ranges mostly between 20 and 40 percent (by volume).

Many processes of magma generation have been advocated in the past, e.g. melting by relief of pressure, localised melting caused by liberation of energy during earthquakes, melting caused by accumulation of heat in regions characterised by a high concentration of radioactivity, and melting connected with rising "convection" cells or "advective movement" in the mantle. After an examination of possibilities, the authors are of the opinion that the only generally satisfactory mechanism for producing basalt magmas and their observed distribution over the face of the earth and through geologic time, is some process connected with "convection" or "advection" in the mantle. Such processes have been advocated by many authors (e.g. Holmes, 1926, 1927; Verhoogen, 1954). Accordingly we will base our discussion of magma generation on this type of process. Nevertheless it should be pointed out that much of the following discussion is of a general nature and could be applied to some other models of magma generation.

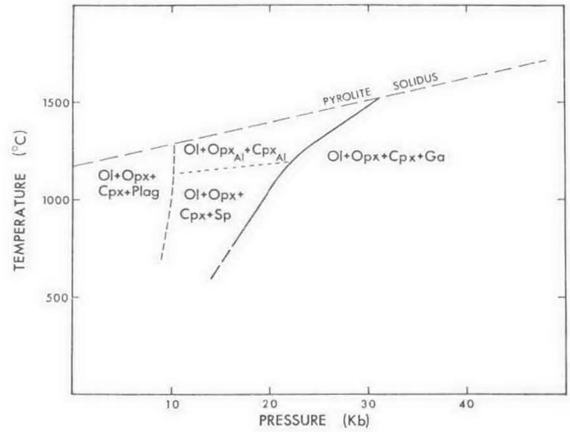

Fig. 11. Preliminary experimental determination of the stability fields of different mineral assemblages in the pyrolite composition of Table 20. The data are for an anhydrous composition and do not include the "ampholite" (olivine + amphibole ± enstatite) assemblage

The condition for gravitational instability in the mantle is that the actual temperature gradient should exceed the adiabatic gradient. This condition is therefore most favourable in the upper few hundred kilometers of the mantle, where the geothermal gradient is greatly in excess of the adiabatic gradient. Indeed, as shown by Clark and Ringwood (1964) it is probable that the density of the mantle actually decreases with depth down to 100 km or so in many regions. Although the outermost mantle many be potentially gravitationally unstable, the triggering-off of an actual instability leading to some form of mass-transfer requires other favourable conditions controlled particularly by rheological properties and by the presence of horizontal inhomogeneities in density, caused either by temperature or chemical differences. Because of the close approach of the actual temperature gradient to the melting point gradient in the uppermost mantle, this region will constitute a zone of low strength and high mobility as previously pointed out by numerous authors. Also, horizontal inhomogeneities are most pronounced in the upper mantle. All these factors contribute towards the occurrence of mass transfer processes in the upper mantle, and their essential restriction to this region.

Processes of mass transfer in the upper mantle are commonly referred to as "convection" a term which characteristically applies to quasiregular, thermally generated motions in a viscous fluid. The analogy has frequently been applied to the earth and it has been argued e.g. Vening Meinesz (1952, 1962), Runcorn (1962) that the mantle is characterised by regular arrays of convection cells, extending as deep as the core. Such schemes appear unrealistic and implausible for many reasons. Elsasser (1963) has provided a stimulating discussion of the subject, and argued convincingly for restriction of mass transport processes to the upper mantle. Furthermore he emphasizes the probable extreme irregularity both in time and configuration, of the processes to be expected in the upper mantle. An additional complication in the models which we shall discuss is that mass motions are accompanied by partial melting and chemical differentiation, and are hence irreversible. Clearly, "convection" in the conventional sense is not an ideal term[5] to apply to such complex processes.

The model for magma generation which we have in mind is given in Fig. 12. It is characterised by a highly specific relationship between the actual temperature distribution and the pyrolite solidus, as shown in the diagram. Gravitational instability in the upper mantle combined with a suitable combination of horizontal inhomogeneity and rheological properties causes a source-mass (S) of solid pyrolite to rise diapirically (in the manner of a salt dome) from the low-velocity zone. It is possible that the initial triggering-off was connected with stresses associated with seismic activity, i.e. the diapir may be derived from an earthquake source-region; however this is not essential. The rising diapir is sufficiently large and hence possesses sufficient thermal inertia in relation to its velocity, so that it cools adiabatically and does not interact by thermal conduction with the surrounding mantle. The adiabatic gradient, of the order of 0.3° C/km (Birch, 1952) is much smaller than the gradient of the pyrolite solidus. Accordingly partial melting in the rising diapir will occur as the temperature of the diapir, following the adiabat from S, intersects the solidus at F (Fig. 12). This causes an increase of the density contrast between rising diapir and surrounding mantle and accordingly, an increase in its rate of upward movement. We assume that the partially melted diapir remains adiabatic. As it rises, and the pressure decreases further, the degree of partial melting increases. The absorption of latent heat accordingly steepens the effective adiabatic gradient, and the temperature of

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[5] In a previous paper (Ringwood and Green, 1966) we adopted the term "advection" for the process, following Elsasser (1963). Although this term adequately specifies the complexity of the mass transport envisaged, it also carries the implication that the horizontal dimensions of the motions greatly exceed the vertical dimensions. This may not necessarily apply in the upper mantle.

the rising diapir (now a crystal-liquid mush) behaves as shown. It is probable that at this stage, the velocity of upward movement is sufficiently slow to permit the liquid component of the mush to remain in chemical equilibrium with the residual unmelted crystals. Eventually, the degree of partial melting becomes sufficiently extensive (20—40%) so that the liquid segregates from residual refractory crystals, and forms an independent, homogeneous magma body. This may be termed the stage of *magma segregation* (M, Fig. 12). From this stage onwards, the magma is no longer in equilibrium with the residual crystals with which it was originally associated. Instead, it may fractionate independently by cooling and crystal settling as it rises towards the surface.

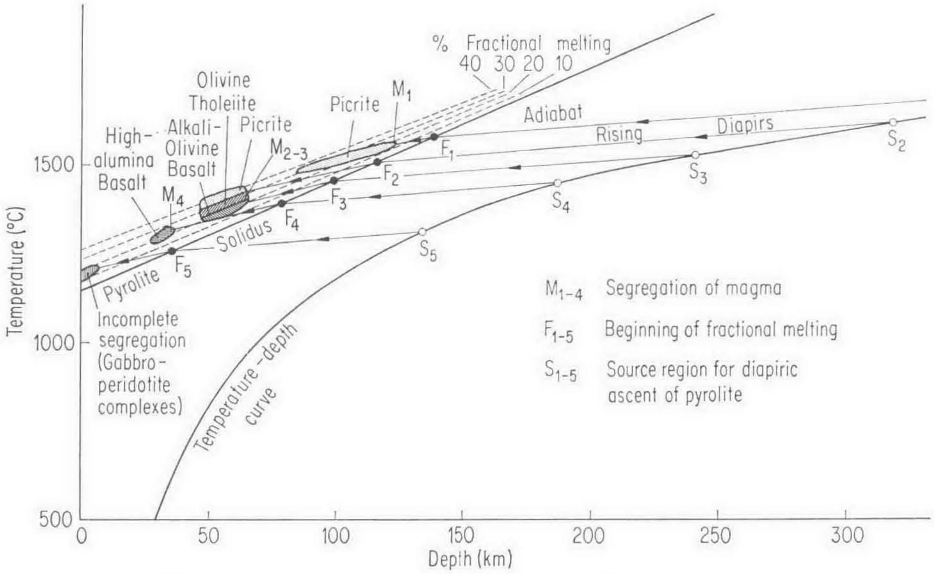

Fig. 12. A model for magma formation by fractional melting of mantle pyrolite. S_1—S_5 represent arbitrary source regions from which there is diapiric ascent of bodies of subsolidus pyrolite to intersect the pyrolite solidus at points F_1—F_5. Partial melting begins at these points and the rising crystal-liquid mushes follow the courses indicated until segregation of magma from residual crystals at M_1—M_4. The nature of the magma is determined by equilibria occurring at M_1—M_4 and not at F_1—F_4

The above outline suggests that the processes of magma generation in the mantle are more complex than sometimes assumed. Thus the initial source region (S) of the rising solid diapir, which may be associated with seismic activity, is probably much deeper than the point (F) at which the earliest liquid forms by fractional melting. As we shall see, the chemical composition of the liquid first formed at (F) is sensitively dependent upon depth. However, the magma which finally segregates at the much shallower level (M) may have an entirely different composition. As long as crystals and liquid in the rising crystal mush remain in chemical equilibrium, as appears likely, the nature of the liquid will change continuously with pressure, and will retain no "memory" of its earlier, deeper origin. The overall chemistry of the magma is in fact, not determined until it segregates from residual refractory crystals at M. Thus the depth of *magma segregation* M is decisive in determining the nature of the magma. This depth may not be related in any simple manner to the initial depth of partial melting (F). Accordingly, in subsequent sections, we will discuss the partial melting of pyrolite according to the depths at which magma may segregate from residual crystals.

c) Partial Melting of Pyrolite — General Discussion

The rationale of the pyrolite model was to select a composition which on fractional melting would yield a generally "basaltic" magma, leaving an unmelted residuum similar in composition to peridotite or dunite. In one sense, fractional melting may be regarded as the reverse of fractional crystallization, providing that the nature of the crystalline phases are similar in both cases. This relationship is independent of the actual proportions of phases which may be present. We have seen that at depths between about 30 kms and 100 kms, the mineral assemblage of pyrolite at the solidus consists essentially of olivine, aluminous enstatite and aluminous subcalcic clinopyroxene. The experiments upon the crystallization of basalt magmas at different pressures showed that the principal magma types could be derived by the separation of olivine and aluminous pyroxenes in various proportions according to P, T conditions and magma compositions. Moreover the Fe/Mg ratios of olivines and pyroxenes occurring near the liquidi of the experimental basaltic compositions are similar to the Fe/Mg ratios of olivines and pyroxenes of mantle-derived ultramafic material such as kimberlite inclusions, peridotite nodules and alpine peridotites. Because of these relationships, we are able to apply our results on fractional crystallization of basalt magmas to fractional melting of pyrolite within the above depth range. The inferred relationships are summarized in Fig. 13 and elaborated in the following sections.

d) Depth of Magma Segregation 0—15 km

It has long been known that at atmospheric pressure, the field of crystallization of olivine extends into quartz normative basaltic compositions. Accordingly, it is possible to derive quartz tholeiite from an olivine tholeiite magma by early crystallization of olivine, followed by segregation of crystals from magma, so that Bowen's reaction relationship is prevented. Conversely, segregation of small amounts of magma from fractional melting of pyrolite under restricted conditions may yield an oversaturated tholeiite magma together with residual dunite (Reay and Harris, 1964).

Boyd et al. (1964) showed that a pressure of a few kilobars was sufficient to prevent the incongruent melting of enstatite. T. H. Green is currently studying this relationship in natural basaltic magmas (personal communication). He finds that saturated tholeiites may form by fractional melting of pyrolite in the depth interval 0—15 km. Below this depth, the olivine reaction relationship does not occur, and the liquid fractions are always olivine-normative.

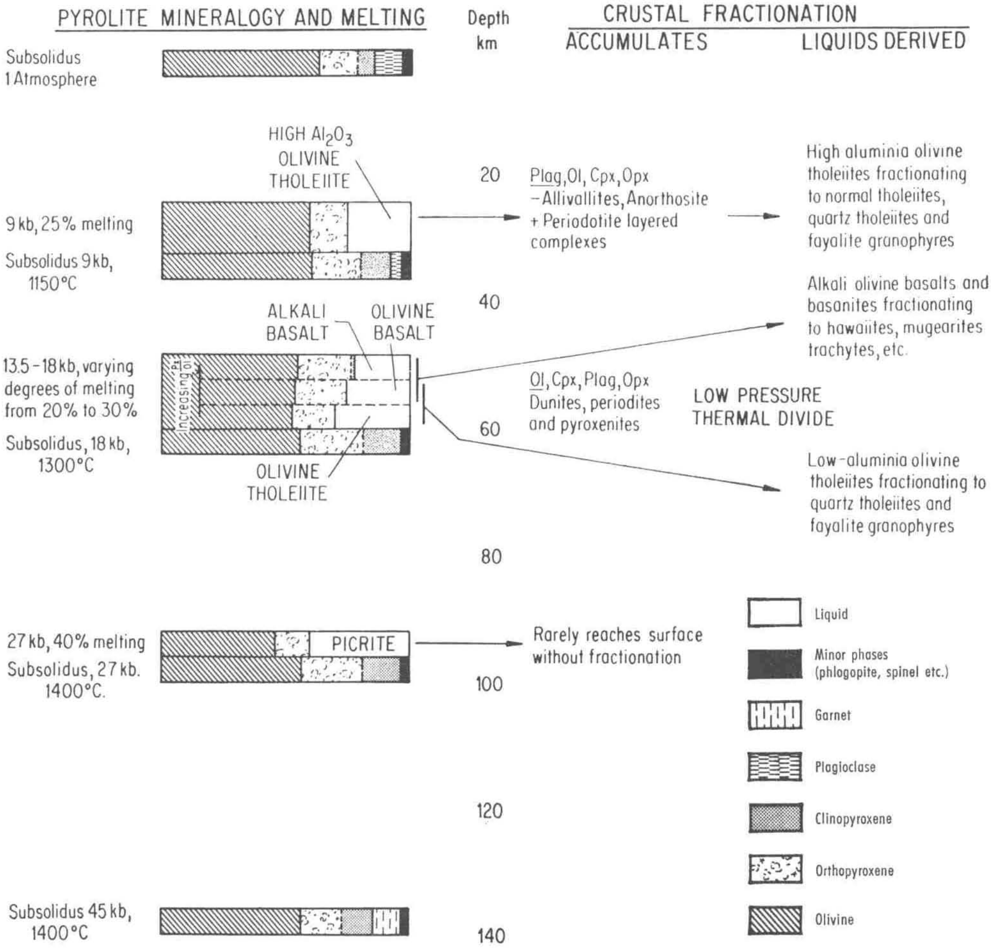

Fig. 13. Diagram illustrating the variation in the near-solidus mineralogy of pyrolite, the degree of partial melting and nature of both the liquid and the refractory residuum at various depths within the mantle

e) Depth of Magma Segregation 15—35 km

Results of experimental crystallization of the olivine tholeiite are given in Table 14. At 1,250° C, the separation of about 12 percent of olivine (Fo_{91-86}) and 3 percent of orthopyroxene (En_{87}, $Al_2O_3 = 5\%$) produced a residual liquid similar to high-alumina olivine tholeiite. Since the liquid is saturated with these phases under the stated conditions, it would be possible to increase these proportions to any desired amount without affecting liquid composition, e.g. 60 percent of olivine and 15 percent of enstatite (5% Al_2O_3) could be added, giving a bulk composition essentially identical to pyrolite. It follows that fractional melting of 20—25% of pyrolite followed by magma segregation under these P, T conditions would yield a high alumina olivine tholeiitic liquid.

This liquid might thus be erupted directly to the surface after segregation or it might undergo crystal fractionation at a slightly higher level under closed system conditions leading to even greater enrichment in Al_2O_3 due to separation of olivine and pyroxenes as previously discussed.

f) Depth of Magma Segregation 35—70 km

The compositions of liquidus orthopyroxenes from the olivine tholeiite and olivine basalt are close to the compositions of natural orthopyroxenes of the olivine + aluminous orthopyroxenes + spinel mineral assemblage in peridotites. Previously Green and Ringwood (1964) have noted that near-liquidus olivine and orthopyroxene from the picrite composition at 13.5 kb closely match the compositions of olivine and orthopyroxene from the olivine + aluminous pyroxenes + spinel assemblage of lherzolite nodules. Olivine was not observed during the crystallization of olivine tholeiite or of the olivine basalt at 12—20 kb. However it was present on the liquidus of the picrite composition. It follows from a consideration of these three compositions that the olivine tholeiite and olivine basalt are almost saturated with olivine between 12—20 kb. We have already noted that they are saturated with highly aluminous orthopyroxene. Accordingly, a fairly large degree of fractional melting of pyrolite under these conditions would yield an olivine tholeiite magma closely resembling our chosen composition, but containing slightly more normative olivine. Such an olivine tholeiite would be in equilibrium with a refractory assemblage of olivine + aluminous orthopyroxene possessing Fe/Mg ratios identical to natural peridotite nodules.

The crystallization experiments upon picrite, olivine tholeiite and olivine basalt at 12—20 kb showed that these magmas would fractionate by means of the crystallization of substantial amounts of aluminous orthopyroxene ± aluminous clinopyroxene to the composition of alkali-olivine basalts. Thus we see that an alkali olivine basalt (with a slightly higher content of normative olivine than the one which we investigated), can be in equilibrium with olivine + aluminous enstatite ± aluminous clinopyroxene. This assemblage would be identical on cooling at pressure to the assemblage found in many lherzolite nodules which are commonly found in alkali olivine basalts and basanites (pages 181—185). These relationships show that olivine-rich alkali basalt may be formed by direct fractional melting of pyrolite.

Whether an olivine tholeiite or an alkali olivine basalt is formed by melting and magma segregation under these conditions depends simply upon the extent of partial melting. With a relatively small degree ($\sim$20%) of partial melting of pyrolite, the residual crystals consist of olivine, abundant aluminous enstatite ± aluminous clinopyroxene, and the liquid has the composition of an olivine-rich alkali basalt. However, as the temperature is increased, the extent of partial melting increases, the clinopyroxene and then a large amount of aluminous orthopyroxene enter the liquid, changing its composition to an olivine tholeiite at about 30 percent partial melting. The situation is exactly the reverse of fractional crystallization of the olivine tholeiite.

The series of liquids developed by different degrees of partial melting at 35 to 70 km would be lower in SiO_2 content and in Al_2O_3 content than the series of liquids developed at 9 kb (30 km approx.) with similar degrees of melting of pyrolite. This is because the ratio of pyroxene/olivine, and particularly of orthopyroxene/olivine in the residual crystals is always higher at 13.5—18 kb than at 9 kb for a given degree of partial melting. In addition the residual pyroxene at 13.5—18 kb contains more Al_2O_3 in solid solution, resulting in lower Al_2O_3 content of the liquid fractions.

g) Depth of Magma Segregation Around 90 km

We have argued that the mineralogy of pyrolite at depths of around 90 km (P ~27 kb) and temperatures near the basalt solidus or liquidus (1,420—1,520° C) will consist essentially of olivine + aluminous enstatite + aluminous clinopyroxene. Garnet is absent from this assemblage but we have shown that garnet is a major phase in the subsolidus assemblages of the basaltic compositions at 27 kb and is the liquidus phase of at least two of the three basaltic compositions at 27 kb. These relationships demonstrate that the three basaltic compositions chosen are not compositions which may be derived by direct partial melting and magma segregation from pyrolite at 27 kb. Compositions which could be derived by partial melting of pyrolite at 27 kb should have olivine + orthopyroxene or olivine + clinopyroxene as liquidus phases or near-liquidus phases. The appearance of orthopyroxene near the liquidus of the picrite composition at 27 kb suggests that this composition is closer to being an appropriate partial melt at this pressure and re-inforces the previous arguments that liquids derived by partial melting and magma segregation at pressures near 30 kb (i.e. 100 km) would have the chemistry of picrites with over 30% normative olivine.

The deduced and observed nature and behaviour of the partial melt at 27 kb illustrates the important role which buffering by residual crystals may play in liquid fractionation. If a picritic liquid is segregated free from residual olivine + pyroxenes and fractionates essentially as a closed system then garnet and clinopyroxene will be the major phases controlling fractionation. If, however, a picritic liquid does not segregate from residual crystals but begins to crystallize again then garnet and olivine are incompatible and will react principally to form aluminous enstatite + olivine. Garnet as a phase will have no role in either the fractionation sequence or in the solidus assemblage at 27 kb unless some segregation of partial melt and residual crystals occurs.

Application of the Hypotheses of Partial Melting and Fractionation to Natural Basaltic Provinces

The conclusions on partial melting in the upper mantle and basalt fractionation at high pressure which have been detailed in previous sections should be relevant to discussions of any basaltic magma province. However, we will briefly apply our conclusions to an interpretation of the causes of the main features of Hawaiian volcanism, and the rather distinctive basaltic magmas of the mid-oceanic ridges.

a) Hawaiian Volcanism

The extensive chemical data on Hawaiian basalts published by MACDONALD and KATSURA (1964) and their review of the main features of the individual volcanoes, together with the review by YODER and TILLEY (1962), serve as the basis for the following discussion. The major eruptive phase of each of the Hawaiian volcanoes comprises voluminous, rapid and rather homogeneous eruptions of olivine tholeiite to quartz tholeiite liquids. The average lava compositions for each of seven major volcanoes is quartz normative but all workers recognize the important role that fractionation at very shallow depths (2—5 km) exercises in producing the variations in lava composition and particularly the trend to quartz-normative tholeiites. The liquids derived from the mantle and "parental" for the low pressure sequence of fractionation, are probably of olivine tholeiite composition, but neighbouring volcanoes may show lava compositions of a consistently distinctive character, e.g. the Pololu Series of Kohala is distinctive in low K_2O, low TiO_2 and high-Al_2O_3 contents and in high Al_2O_3/SiO_2 ratio. Individual lavas are closely similar if not identical to the high-Al_2O_3 "oceanic" tholeiites of ENGEL, ENGEL and HAVENS (1965). The early, voluminous tholeiites of the Hawaiian Islands pass *transitionally* to an upper zone of interlayered tholeiite and alkali olivine basalt types — the transitional nature of this change of magma type is most significant, as is the restriction of the alkali olivine basalts to the late, waning stage of activity of any particular volcano. In some volcanoes [of Kohala type (MACDONALD and KATSURA, 1964)] the upper zone containing alkali olivine basalts is disconformably overlain by a thin veneer of mugearite and related rocks, i.e. lavas derived by shallow level fractionation of the alkali olivine basalts. A similar relationship appears to hold for Piton des Neiges volcano, Reunion Id. (UPTON and WADSWORTH, 1965).

Geophysical studies on the currently active Kilauea volcano indicate that seismic activity correlatable with the ascent of magma for the 1959-60 eruption of Kilauea Iki was first observed at depths of 50—60 km. A reasonable interpretation of this is that 50—60 km marks the depth at which magma segregated and accumulated as a discrete body. We have shown that magma segregation following a high degree of partial melting at 35—70 km will produce olivine tholeiite magmas. If these magmas move rapidly from this depth to shallow levels and then fractionate to a variable extent at 2—5 km, the observed sequence of early tholeiitic series will result. Rapid and voluminous tholeiite eruptions will continue as long as the melting process is active and the zone of pyrolite is not exhausted of its basaltic component. Variations towards high-alumina tholeiites, e.g. Pololu Series, will result if fractionation occurs at intermediate depths (20—30 km) or alternatively if magma segregation for other volcanoes occurs at shallower depths ($\sim$30 km) than deduced for Kilauea. The latter explanation is preferred.

In the waning phase of activity of each volcano the ascent of magma batches becomes less frequent and cooling at the depth of segregation may occur. Cooling of an olivine tholeiite liquid at 35—70 km depth will cause crystallization of aluminous enstatite and the magma composition will change towards alkali olivine basalt. The Hawaiian sequence of rock types, the transitional nature of the change from tholeiitic series to alkali olivine basalt and the relative volumes of tholeiite and alkali types are all consistently explained by the hypothesis of high pressure fractionation of olivine tholeiite during waning phases of activity. Like the tholeiitic series, the alkali olivine basalt magmas may accumulate and differentiate in shallow level magma chambers. It is significant that the most undersaturated rocks of the late alkali olivine basalt series appear to be basanites, i.e. compositions near the suggested limit of fractionation towards nepheline-enrichment at 35—70 km. Some Hawaiian volcanoes have a much later period of eruption of olivine nephelinites and nephelinites but these seem unlikely to be derived by extreme fractionation at 35—70 km of the alkali olivine basalt series.

The Hawaiian volcanic province is one in which the observed proportions and sequence of magma types support an hypothesis of origin of alkali olivine basalt magma by crystal fractionation of olivine tholeiite at 35—70 km depth. Complete transition between the magma types occurs at the depth of fractionation but later, shallow level fractionation of most liquids has imposed the effects of the low pressure thermal divide between the nepheline-normative and hypersthene-normative rock series leading to a greater chemical separation of the magma series. Perhaps the most notable effect of low pressure fractionation is that the average composition of the tholeiitic series for the Hawaiian volcanoes is almost exactly saturated in silica with either minor olivine or quartz in the norm (MACDONALD and KATSURA, 1964). These liquids represent the effects of precipitation of some 20% (if olivine only precipitated) to 30% (olivine ± clinopyroxene ± plagioclase) of crystals at very shallow depths from the olivine-rich tholeiitic liqudis present at 35—70 km depth. Geophysical data showing the existence of high density and seismically fast material at shallow depths beneath the central regions of Hawaiian volcanoes (STRANGE, WOOLLARD and ROSE, 1965; FURUMOTO, THOMPSON and WOOLLARD, 1965) may be consistent with such olivine-rich, low pressure accumulative material.

b) Oceanic Ridge Volcanism

In recent years, papers by ENGEL and ENGEL and their co-workers have focussed attention on the distinctive chemical characteristics of basalts dredged from the ocean floor at depths around 3,000 metres and located along the mid-oceanic ridges (Mid-Atlantic Ridge, Indian Ocean ridges, and the East Pacific Rise). Earlier analytical work by WISEMAN (1940) and independent work by NICHOLLS, NAWALK and HAYS (1964) support the conclusions of ENGEL and ENGEL (1964a, b), ENGEL, FISHER and ENGEL (1965) and ENGEL, ENGEL and HAVENS (1965) that a characteristic magma is of olivine tholeiitic character with $SiO_2 = 49—50\%$, $Al_2O_3 = 15—17\%$, $Na_2O = 2.5—3.0\%$, $K_2O < 0.3\%$ and $100\ Mg/Mg + Fe^{++} \sim 65$. This composition is closely similar to the high-alumina, olivine tholeiite composition derived by fractionation of low-Al_2O_3, olivine-rich tholeiite at 9 kb (pages 142—148) and also interpreted as a direct product of magma segregation from mantle pyrolite at depths around 30 kms.

ENGEL and ENGEL (1964b) and ENGEL, ENGEL and HAVENS (1965) have argued that these high-alumina olivine tholeiites are the only "primary" basaltic magma generated by partial melting within the mantle and that other tholeiitic and alkali magmas are derived by processes of crystal differentiation and volatile transfer, mainly in volcanic conduits at depths less than 3 km. The possible directions of fractionation of high-alumina olivine tholeiites at various pressures are at present under study by Mr. T. H. GREEN and his data is unfavourable to any hypothesis of crystal fractionation at shallow depths leading to alkali olivine basalt compositions (T. H. GREEN, personal communication).

There is an alternative process consistent both with our results on high pressure fractionation and partial melting and with the observed features of chemistry and distribution of volcanics along the mid-oceanic ridges. The high-Al_2O_3 olivine tholeiites have chemical compositions appropriate for liquids segregated from a pyrolite composition at the comparatively shallow depth of about 30 kms. With about 25% melting and segregation of the magma from pyrolite at about 30 kms depth, high-Al_2O_3 olivine tholeiitic liquids could be produced leaving residual olivine + low-Al_2O_3 enstatite. Apart from the probable dropping of minor (0—10% approx.) olivine during ascent, the high-Al_2O_3 olivine tholeiites show little chemical change consequent on low pressure fractionation. These magmas thus approach the simplest model of partial melting with extraction of a rather constant proportion of magma from a given volume of pyrolite and transport of the magma through the oceanic crust with a minimum of chemical change by either fractionation or processes of wall-rock reaction. In comparison with other tholeiitic provinces it may be relevant that the environment of intrusion or extrusion of the oceanic tholeiites requires 3—5 km less vertical travel than the exposed Hawaiian tholeiites and there is no thin zone of near-surface material of low density to be penetrated.

The distinctive chemical compositions of the oceanic tholeiites seem to require particularly intense or active upward movement of mantle such that separation of magma and crystals does not occur until depths of about 30 km. If the proposed rapid mantle upwelling process in any part of the mid-oceanic ridge begins to wane in intensity it is probable that the zone of magma segregation will move to greater depths and also the frequent, fissure type of eruption will change towards a localized and central type of eruption. If this process occurs then there will be a gradual transition from high-alumina olivine tholeiite to low-alumina olivine tholeiite or to alkali olivine basalt.

c) Other Provinces

The Hawaiian alkali olivine basalts are interpreted as the product of crystal fractionation of olivine tholeiite magma. Other basaltic provinces of regional extent are characterized by alkali-type magmas ranging from alkali olivine basalt to nephelinites and excluding the tholeiitic suite entirely. KUNO (1960) has drawn attention to this feature in Japan and correlated the distribution of the alkali series with a region characterized by particularly deep melting of the mantle. The Tertiary volcanic provinces of Eastern Australia are almost entirely of the alkali olivine basalt + basanite + nephelinite type. In contrast Tertiary basalts of Western Australia and the Jurassic dolerites of Tasmania are tholeiitic. The Tasmanian tholeiitic province, as well as the much more widespread Jurassic Ferrar dolerites of Antarctica and Karroo dolerites of South Africa form distinctive magma provinces. Other examples could be cited supporting the observation that basaltic provinces of regional extent may be typified by alkali olivine basalt magmas alone or by tholeiitic magmas alone. Examples such as these are interpreted as direct partial melting products from the mantle, with relatively constant conditions of melting and magma segregation over the region. Extensive fractionation at low pressure within the crust appears to be a normal characteristic of the continental tholeiitic series.

Trace Element Contents and Isotopic Ratios of Basaltic Rocks in Relation to Partial Melting and Fractionation Processes

The relationships between basalt magma types which have been deduced in previous sections, are based on a consideration of the relative concentrations of the major elements (Si, Al, Fe, Mg, Ca, Na and Ti). These elements largely determine the normative and modal mineralogy and thus the classification of basaltic magma types. However, the minor elements K, Ti, P, and trace elements U, Th, Ba, Cs, Rb, Sr, Zr, Hf and the rare earth group, are frequently much more abundant in alkali olivine basalts than predicted by simple crystal fractionation from olivine-rich tholeiite or by differences in degree of partial melting of the same parental peridotite. Isotopic ratios also show variations amongst basalts which must be considered in formulating hypotheses in their origin. In this section trace element behaviour and isotopic ratios amongst basalts are briefly reviewed and the data derived in this way are applied to formulate a more detailed discussion of the partial melting and fractionation processes within the mantle.

a) Enrichment in K_2O and Associated Elements

The high-alumina olivine tholeiites (oceanic tholeiites) have been interpreted as the result of simple melting of 20—25% pyrolite mantle and extraction of basalt with a minimum of fractionation. In this case their K_2O content of 0.1—0.2% may be used to estimate an average K_2O content of 0.03—0.05% for their source region, assuming no retention of K_2O in residual olivine or enstatite. Many alkali olivine basalts and continental tholeiites contain 0.5 to 1% K_2O and their derivation by simple partial melting from such a source would require that these magmas represent extraction of 5% or less of liquid. Alternatively some additional process of K_2O enrichment must be postulated for these magmas.

In the Hawaiian volcanic province, comparison of average Hawaiian alkali olivine basalt with average Hawaiian tholeiite (Macdonald and Katsura, 1964) shows an increase of 100% or more in K_2O content in the alkali series. This cannot be accounted for by 10—30% fractionation or the difference between 20% and 30% partial melting of the same pyrolite parental material. In the Hawaiian example there is a smaller but similar inconsistency in the P_2O_5 contents whereas TiO_2 variation between the suites is broadly consistent with the suggested major element fractionation. The abundances of U and Th in alkali olivine basalts compared to tholeiites in Hawaii are greater by factors of 3 to 5 instead of 1.5 to 2 as expected in the simple crystal fractionation model (Heier and Rogers, 1963; Heier, McDougall and Adams, 1964; Hamilton, 1965; Tatsumoto, Hedge and Engel, 1965). Nephelinites and melilite nephelinites are in even more marked contrast having over five times the U and Th contents of the tholeiites (Heier et al., 1964; Hamilton, 1965). Although data are meagre and commonly not easily correlated with total rock chemistry, there appears to be a similar contrast in abundance in tholeiites and alkali olivine basalts for Ba (Engel et al., 1965; Le Maitre, 1962), for Zr (Engel et al., 1965; Naughton and Barnes, 1965) and probably for Cs and Hf. Recent data for Rb and Sr contents of basalts are much more extensive and demonstrate a much greater enrichment of Rb in the alkali olivine basalt series (Tatsumoto et al., 1965; Hamilton, 1965; Engel et al., 1965; Lessing and Catanzaro, 1963). Strontium shows a similar pattern although enrichment is not as marked as for rubidium. The limited data on Sr contents of nephelinites and melilite nephelinites show extremely high Sr contents (HAMILTON, 1965).

As well as large enrichment factors in the alkali olivine basalts, the above-mentioned elements show relative fractionation within the group (as might be expected from the rather diverse elements included). There is an independent and rather random variation of K_2O, TiO_2 and P_2O_5 among basalts (GREEN, 1966) e.g. there exist tholeiites with $TiO_2 < 1\%$ and others with $TiO_2 > 2.5\%$; low TiO_2 may be coupled with high K_2O or low K_2O etc. HEIER and ROGERS (1963) noted a correlation of high Th/U ratios with high K_2O contents while GAST (1965) noted that high K/Rb ratios (>1000) occurred in the oceanic tholeiites with very low K_2O contents and that this ratio decreased (down to 350) with increasing K_2O contents through the tholeiitic and alkali olivine basalt series. Continental tholeiites with K_2O contents similar to the oceanic alkali olivine basalts have even lower K/Rb ratios of about 200 (HEIER et al., 1965; COMPSTON et al., 1966; GUNN, 1965).

The data summarized above suggest that the process or processes which determine the abundances of the group of elements discussed above, are sufficiently selective to produce marked fractionation within the group. This conclusion is further illustrated and supported by recent studies of rare-earth element distributions in basalts (HASKIN, FREY, SCHMITT and SMITH, 1966; SCHILLING and WINCHESTER, 1966). The total rare earth content of basalts ranges from around 100 ppm for high-alumina oceanic tholeiites and a Kilauean tholeiite, 174 ppm for a composite sample of 282 continental basalts, 280—430 ppm for oceanic island alkali basalts and 530 ppm for a continental alkali olivine basalt. As a group, the rare earth elements are enriched in terms of their total abundance in basalts in similar manner to K, Rb, Sr etc. The overall enrichment in rare-earth abundances is accompanied by marked fractionation within the rare-earth group (HASKIN et al., 1966; SCHILLING and WINCHESTER, 1966). The high-alumina oceanic tholeiites are distinctive amongst basalts so far analyzed in having low rare-earth abundances and a distribution pattern similar to the chondritic abundance pattern, although there is relative depletion (compared to chondrites) in the lightest rare-earth elements (La, Ce, Pr). Hawaiian tholeiites show low or moderate degrees of overall enrichment and maximum enrichment factors, relative to the chondrite pattern, for Pr and Nd. Alkali olivine basalts, and basanites, show greater enrichment than tholeiites throughout the rare earth series but particularly in the lightest rare earth elements (enrichment factors, relative to a low-potassium tholeiite, of 2.5 to 3.5 for La compared to factors around 1.2 for Lu; SCHILLING and WINCHESTER, 1966). Nephelinite and melilite nephelinite from Hawaii show enrichment factors relative to tholeiite which range from over 10 for La to about 1.1 for Lu (SCHILLING et al., 1966). The processes which lead to enrichment in the rare earth elements in alkali olivine basalts and basanites relative to olivine tholeiites, also lead to marked fractionation within the rare earth group. Elements which are heavier than Gd have abundances in alkali olivine basalts and basanites which appear consistent with the fractionation or partial melting relationships outlined in previous sections. Elements which are lighter than Gd (and have progressively larger ionic radii) have abundances which are inconsistent with the simple fractionation model and resemble K, Rb, Sr etc. in their abundance relationships.

b) Isotopic Abundances and Fractionation

Data obtained from basaltic rocks indicate a rather constant initial Sr^{87}/Sr^{86} ratio of 0.703—0.704 (FAURE and HURLEY, 1963; HEDGE and WALTHALL, 1963; LESSING and CATANZARO, 1964; HEIER, COMPSTON and MCDOUGALL, 1965; GAST, 1966). LESSING and CATANZARO (1964) found no differences in Sr^{87}/Sr^{86} ratio between Hawaiian basalts of tholeiitic and alkali basalt types. This is supported by data given by HAMILTON (1965), POWELL, FAURE and HURLEY (1965) and POWELL and DE LONG (1966) and is consistent with a simple crystal fractionation relationship for these two magma series. However both POWELL et al. (1965) and HAMILTON (1965) noted that the highly undersaturated, low-SiO_2 nephelinites and melilite nephelinites from Oahu (Honolulu Series) have distinctly lower Sr^{87}/Sr^{86} ratios than either the tholeiites or the alkali olivine basalts. This problem was further amplified by POWELL and DE LONG (1966) with extensive data from the Honolulu Series, Oahu (basanites to melilite nephelinites) showing that this series has consistently lower Sr^{87}/Sr^{86} ratios than the tholeiitic Koolau Series, Oahu. However, four specimens of the Waianae Series, Oahu, including tholeiite, alkali basalt, hawaiite and rhyodacite gave low Sr^{87}/Sr^{86} ratios similar to the highly undersaturated Honolulu Series. Current data from Hawaii appear to suggest that a given Sr^{87}/Sr^{86} ratio is characteristic for a particular volcanic centre or series and not for a specific magma type.

GAST, TILTON and HEDGE (1964) found consistent differences in the Sr^{87}/Sr^{86} ratio for alkali olivine basalts and their differentiates from two oceanic islands, Gough Island ($Sr^{87}/Sr^{86} = .7045$) and Ascension Island ($Sr^{87}/Sr^{86} = .7025$). They observed that these differences in Sr^{87}/Sr^{86} were accompanied by differences in Pb^{206}/Pb^{204} ratios while Pb^{208}/Pb^{204} ratios were closely similar. GAST et al. concluded that the isotopic differences between equivalent magma types on the two islands reflect differences in the source regions of the mantle, i.e. higher Rb/Sr in the mantle acting as source for the Gough Id. basalts and higher U/Pb but not Th/Pb for the Ascension Id. source. It is relevant that LE MAITRE (1962) pointed out compositional differences between the alkali olivine basalt series of the two islands, the Gough Id. rocks having higher K_2O contents and K_2O/Na_2O ratios than those from Ascension Id.

The compilation of Sr^{87}/Sr^{86} data by GAST (1966) further suggests that regional variations in Rb/Sr in mantle source regions may cause the observed variation in Sr^{87}/Sr^{86} ratios and that this is not coupled to rock type. There is some evidence that basalt series characterized by high K_2O contents such as the alkali olivine basalts of Gough Id. and Samoa or the quartz tholeiites of Tasmania and Antarctica also have the highest Sr^{87}/Sr^{86} ratios and the lowest ratios are found in series with relatively low K_2O contents or K_2O/Na_2O ratios such as the high-alumina oceanic tholeiites (TATSUMOTO, HEDGE and ENGEL, 1965) or the Ascension and St. Helena Id. alkali olivine basalts (GAST, 1966). GAST's (1966) compilation also illustrates the greater abundance of basalts with $Sr^{87}/Sr^{86} > .705$ from continental areas rather than oceanic areas, a result accentuated by the work of HEIER et al. (1965) and COMPSTON et al. (1966) on Jurassic continental tholeiites.

TATSUMOTO et al. (1965) found that the Rb/Sr ratios of high-alumina oceanic tholeiites was too low to account for the observed Sr^{87}/Sr^{86} ratios, assuming an initial Sr^{87}/Sr^{86} ratio of .6984 for the earth system. These authors concluded that the magmas either were not representative of their source in Rb/Sr ratio or alternatively that the source region had changed its Rb/Sr ratio during earlier evolution from a high value of Rb/Sr to the present low value (0.011) observed in the oceanic tholeiites.

Inconsistencies in isotopic abundances and parent element abundances are even more apparent in the U-Th-Pb system (GAST et al., 1964; TATSUMOTO, 1966a, b; GAST, 1966). These authors have postulated multistage differentiations of the source material to account for differences between the U/Pb and Th/Pb ratios of the basalts and the ratios required by the observed Pb^{208}/Pb^{204}, Pb^{207}/Pb^{204}, Pb^{206}/Pb^{204} and assumed primordial abundance ratios. In addition, heterogeneity in U-Th-Pb in the mantle source region seems to be required by the data.

c) *Inferences from Minor Element and Isotopic Studies*

The data briefly summarized in previous sections lead to the following general conclusions:

(i) Processes other than simple crystal fractionation in a "closed-system" magma are necessary to account for observed abundances of the elements (K, Ti, P, Ba, Cs, Rb, Sr, U, Th and the lighter rare earth elements) in tholeiitic and alkali olivine basalt magmatic series.

(ii) Not only are these elements fractionated as a group but there is also fractionation within the group leading to changes in abundance ratios such as K/Rb, K/Ti, K/Zr, Rb/Sr etc. and in relative abundances of elements of the rare earth group.

(iii) The simple concept of partial melting in which elements strongly partitioned into the liquid fraction preserve the abundance ratios of the parental mantle does not account for the observed relationships of the highly enriched elements nor for observed isotopic ratios. In particular the abundance ratios Rb/Sr, U/Pb, Th/Pb of a magma probably differ from the ratios of the original parental mantle and may not be simply related to the strontium or lead isotopic ratios present in the magma.

(iv) Variations in isotopic ratios amongst basalts require that the basaltic magmas interact with regions of mantle or crust which have differed for considerable lengths of time in their Rb/Sr, U/Pb and U/Th ratios. This implies inhomogeneity in distribution of these elements within at least the uppermost mantle.

(v) In highly fractionated derivatives of basaltic magmas the late differentiates may show higher Sr^{87}/Sr^{86} ratios. Examples of the process, e.g. Skaergaard granophyres (HAMILTON, 1963) and Gough and Ascension Id. trachytes (GAST et al., 1965) suggest that fractionation of basaltic magmas, even at low pressure, is not a simple, closed-system process.

The elements K, Rb, U, Zr etc. which show relatively high enrichment factors in the alkali olivine basalts have the common property that all of them are incapable, because of their ionic radii and valency states, of appreciable substitution in the major minerals of pyrolite, i.e. in olivine, orthopyroxene, chromian clinopyroxene and possibly spinel or garnet. This group of elements is the same as that considered by Harris (1957) to be concentrated by processes of "zone refining" in the mantle and referred to as "incompatible elements" by Ringwood (1966a). Both authors stressed the crystal-chemical factors controlling the behaviour of these elements in the mantle and concluded that these "incompatible elements" would be strongly partitioned into any liquid fraction present in the mantle. In the absence of a liquid phase, these elements are considered to be present in the mantle in minor phases such as phlogopite, Mg-ilmenite. apatite or other phases stable under upper mantle pressure, temperature conditions (Harris, 1957; Hess, 1964).

In contrast to the behaviour of the highly enriched elements, for which we shall adhere to the term "incompatible elements" (Ringwood, 1966a) other minor elements such as Mn, Cr, Ni or trace elements (Cu, Co, Ga, V, Y) show comparatively small variations between tholeiitic and alkali olivine basaltic series and appear consistent with the simple crystal fractionation or partial melting hypotheses. These elements all substitute comparatively readily in the major minerals (olivine + pyroxenes ± spinel ± garnet) of the mantle and in the liquidus phases of basaltic magmas. In the present context, these elements have been grouped as "compatible elements" referring to their crystal-chemical properties in relation to the major mineral phases of the mantle.

d) Wall-Rock Reaction

The hypothesis that fractional crystallization of olivine tholeiite magma at 35—70 km depth yields derivative alkali olivine basalt liquids requires cooling of the magma body against its wall-rock environment. Cooling of a magma leading to partial crystallization may occur by two main processes —

(i) heat transfer by conduction to cooler wall-rock without intermixing of material, or

(ii) by reaction with and solution of crystalline material incorporated into the magma.

The relative importance of these two processes probably differs at different depths and for different magmatic provinces. At shallow crustal depths the large temperature contrast between wall-rock and magma *usually* results in chilling of the magma in a zone near the contact so that the remaining liquid is isolated in a protective sheath of rock with similar chemistry. Under these conditions there is a minimum of chemical exchange between magma and wall-rock and the magma may behave essentially as a closed system during any fractionation process.

At greater depths in the mantle, particularly near the depth of segregation of basaltic magmas, there may be little temperature contrast between magma and mantle wall-rock. Under such conditions, cooling of a magma may proceed mainly by a process of "wall-rock reaction". We envisage the nature of this reaction as solution of low-melting components from the wall-rock and their incorporation into the magma, accompanied by precipitation of the liquidus phase of the basalt. The concept of "wall-rock reaction" is closely related to the "zone-refining" concept discussed by Harris (1957) and our arguments have been considerably influenced by this earlier paper. However "wall-rock reaction" is applied in a much broader sense and to a wider spectrum of basalts than was attempted by Harris. In addition "zone refining" is a defined physico-chemical process and, in its geological application, was coupled to models of solution of roof material, magma convection and bottom precipitation and also to specific models of mantle temperature distribution (Harris, 1957, p. 199).

In considering the mantle environment of a magma near its depth of segregation, the chemical nature of the wall-rock will, in general, be that of pyrolite or a refractory peridotite composition or some composition intermediate between these extremes. Ascent of the magma near its source within the mantle may be by processes of stoping or by displacive intrusion. In both cases the contacts between magma and wall rock will not be chilled but will be diffuse with a gradual decrease in the ratio of liquid to crystals away from the magma body. Mantle material which is processed by the magma body by stoping or by marginal reaction may lose its low melting fraction into the basaltic liquid and the nature and proportion of the liquid fraction thus extracted will depend on the temperature of the magma and the temperature reached by the wall-rock. Lower temperature basaltic magmas will clearly be able to melt and extract only a small proportion of the potential basaltic magma in any incorporated mantle material but this small fraction may be highly enriched in elements which are not readily accepted by the major mantle minerals, i.e. the incompatible elements. In contrast, higher temperature basaltic magmas may extract almost all the potential basaltic liquid from enclosed pyrolite blocks and there would be little or no selective enrichment or fractionation in the lowest melting components of pyrolite.

In the particular case of an olivine tholeiite magma segregated at 35—70 km depth in the mantle, cooling of the magma by the "wall-rock reaction" process would include solution of the lowest melting fraction from any stoped or accessible wall-rock material and the complementary process of precipitation of the liquidus phase, i.e. aluminous enstatite. As this cooling continues, the major element chemistry of the liquid would trend towards alkali olivine basalt composition with slightly decreasing Mg/Fe ratio. The magma would also, because of cooling and changing Mg/Fe ratio, be able to melt successively smaller fractions from any stoped mantle peridotite with which it is in contact. The lowest melting fraction and most easily extracted elements from pyrolite are the "incompatible" elements previously discussed. These elements could become highly enriched in the magma depending on the amount of mantle material processed through the magma and on the proportion of mantle material soluble in the magma. The process of crystal fractionation at depths near that of magma segregation in the mantle is considered to be *normally* accompanied by "wall-rock reaction" leading to enrichment in incompatible elements to a degree greater than is possible by closed system fractionation alone. It is also apparent that with extraction of extremely small amounts of liquid from the mantle wall-rock, the opportunity for fractionation between elements is greatly enhanced. Much more data is needed on partition relationships for the incompatible elements in both major and minor minerals of the pyrolite composition. For instance, it has been suggested (Green, 1967) that the inverse relationship between K/Rb ratios and K_2O contents (Gast, 1965) could be a consequence of melting and wall-rock reaction involving a major phase with low K_2O content but high K/Rb ratio (possibly pyroxenes) and a minor phase with high K_2O content and low K/Rb ratio (possibly phlogopite). It has been shown (Haskin et al., 1965) that garnet in eclogite is enriched relative to clinopyroxene in the heavier rare earths. Aluminous enstatite is chemically similar to pyrope-almandine garnet and, if similarly enriched in the heavier rare earths, in a liquid-orthopyroxene or clinopyroxene-orthopyroxene partition, could be a major factor in producing the light-element enriched rare earth pattern of alkali olivine basalts. A knowledge of orthopyroxene-liquid, clinopyroxene-liquid and phlogopite-liquid partition relationships at high pressures would probably show methods by which fractionation could yield liquid fractions with Rb/Sr, U/Th, U/Pb and Th/Pb ratios different from those in either the initial pyrolite source for the magma or the mantle material processed by the wall-rock reaction mechanism.

A corollary of the "wall-rock reaction" mechanism is that zones of pyrolite which are still potential sources for basaltic magma in major element chemistry, may acquire lower abundances and different element-ratios for the incompatible elements by being "processed" through an early magmatic cycle. The application of this to the origin of lherzolite nodules in basalts is discussed in the following section. Batches of magma or accumulates from magmas formed in early magmatic cycles may remain as segregated bodies within the mantle and develop isotopic ratios for Sr and Pb which are of local extent and very different from the ratios in subjacent pyrolite. Such bodies may be intersected by later ascending magmas, and, if their content of incompatible elements is extracted by wall-rock reaction, may modify the composition and isotopic ratios of the later magma batch.

To summarize, the fractional crystallization and slow ascent of an olivine tholeiitic magma at 35—70 km depth or the slow ascent of an alkali olivine basalt magma derived by direct partial melting at 35—70 km, would be accompanied by extensive and highly selective reaction with mantle wall-rock. This would cause overall enrichment in incompatible elements and selective fractionation amongst them. The process would also lead to heterogeneity in incompatible element distribution and ratios in the mantle, creating the necessary conditions for development of varied Pb-isotope ratios and, to a less marked degree, Sr^{87}/Sr^{86} ratios in different mantle regions.

It is important to note that enrichment in incompatible elements accompanies fractionation of an olivine tholeiite magma near its depth of segregation so that magmas enriched in incompatible elements at 35—70 km depth are of alkali olivine basalt type. Enrichment in incompatible elements while retaining tholeiitic affinities requires that an olivine tholeiite magma cools and fractionates at lower pressures i.e. along the trend towards high-Al_2O_3 tholeiites or towards quartz tholeiites. This aspect of wall-rock reaction is treated in more detail in the following section.

e) Wall-Rock Reaction at Lower Pressures

Under some conditions it is possible that an environment suitable for extensive wall-rock reaction may occur under lower pressure conditions, e.g. if magma segregation occurs at higher levels in the mantle or if continued magmatic activity causes wall-rock temperatures in the lower crust or top of the mantle to approach the magma temperature. There are important differences between the nature of the highly enriched "incompatible elements" in a process of wall-rock reaction at low pressure in comparison with wall-rock reaction at high pressure. At high pressure, plagioclase is not a stable phase in the wall-rock mineralogy, and strontium behaves as an incompatible element since it does not readily enter either clinopyroxene or orthopyroxene. At low pressure, plagioclase is a major stable phase in the wall-rock mineralogy and as strontium substitutes for calcium in plagioclase, strontium will behave as a "compatible element" in fractionation or reaction processes. In addition, at sufficiently low pressure, plagioclase may be a liquidus or near-liquidus phase and as strontium is preferentially enriched in plagioclase (WILKINSON, 1959; HEIER, 1962, p. 436), precipitation of this phase will prevent marked enrichment of strontium in a fractionating basaltic liquid at very low pressure.

Titanium may exhibit similar differences in behaviour at high and low pressure particularly if ilmenite or titanomagnetite have the character of relatively refractory phases at low pressure in intermediate or acidic compositions. Changes in redox potential between mantle and lower crust environments may also potentially alter the role of titanium.

The Jurassic tholeiites of Antarctica, South Africa and Tasmania are considered to be possible examples of magmas which have trace element contents determined in large part as a result of wall-rock reaction and fractionation processes occurring at low pressure, and in these particular continental examples, in a crustal rather than upper mantle environment. The Ferrar Dolerites (Antarctica) and Tasmanian dolerites are highly enriched in K, Rb, U, Th but are conversely low in Sr and Ti in comparison with olivine and quartz tholeiites from oceanic islands or oceanic ridges (COMPSTON et al., 1966). In the Ferrar Dolerites there is strong K_2O enrichment accompanying increasing SiO_2 oversaturation in the sequence of magmas ranging from olivine tholeiite (0.38% K_2O, 50.40% SiO_2) to pigeonite quartz tholeiites (1.27—1.76% K_2O, 55.25—55.87% SiO_2) (GUNN, 1962, 1966). The initial Sr^{87}/Sr^{86} ratios for both Tasmanian and Antarctic tholeiites are consistently above 0.710, in contrast to the much lower values (0.702—0.704) found in alkali olivine basalts or tholeiites of oceanic areas.

It is suggested that the parental olivine tholeiite magmas for these intrusions began to crystallize at crustal levels giving rise to a spectrum of magma compositions for individual intrusions ranging from oversaturated, iron-rich pigeonite tholeiite to more magnesian olivine tholeiite (GUNN, 1962, 1966). The period of slow ascent of very large volumes of basic magma through the lower crust established conditions suitable for selective extraction of a low-melting fraction from the wallrock environment. The nature of this extraction process presents some difficulty if the average lower crust is of intermediate or acid composition, unless it is a partial melting process with migration of the liquid phase into the basaltic magma. The elements enriched in the basaltic magma are considered to be those least able to substitute in the major phases of the surrounding wallrock, i.e. in pyroxenes, plagioclase, quartz and possibly hornblende and ilmenite or titanomagnetite. The contaminating material would be anticipated to have high Sr^{87}/Sr^{86} ratios and lead isotopic ratios consistent with geologically old crustal material. HEIER et al. (1965), COMPSTON et al. (1966) and GAST (1966) have discussed more fully the role of contamination processes by crustal material enriched in Sr^{87}/Sr^{86} in deriving the particular chemistry of the Tasmanian and Antarctic tholeiites. COMPSTON et al. (1966) present arguments favouring a highly selective contamination rather than wholesale assimilation by average crustal material. The process of wall-rock reaction detailed previously goes some way towards indicating a mechanism for such selective contamination but requires that the average lower crustal material will only be melted to a very limited extent by the tholeiitic magmas and that the bulk of the wall-rock material would remain as refractory residue. A lower crust of anhydrous granulitic or anorthositic composition (RINGWOOD and GREEN, 1966a, b; T. H. GREEN, 1966) would possibly be suitable wall-rock material for the process and effects outlined above. The alternative explanation for the particular chemical compositions of the Tasmanian and Antarctic quartz tholeiites requires a mantle source region of high K, U, Th etc. content and high Sr^{87}/Sr^{86} ratio (HEIER et al., 1965; COMPSTON et al., 1966).

If the conclusions on the behaviour of Sr at low and high pressure are correct then rocks enriched in incompatible elements at low pressure should be characterized by rapid increases in such ratios as K/Sr and Rb/Sr whereas basalts enriched in incompatible elements at high pressures should show relatively small changes in K/Sr and Rb/Sr through quite large variations in overall incompatible element abundances. Without making an exhaustive literature search we have assembled data in Table 21 which supports this hypothesis. In the oceanic tholeiite to basanite series with K contents increasing by a factor of 20 and Rb contents increasing by a factor of 40, the K/Sr ratio only varies between 10 and 21 and the Rb/Sr between 0.01 and 0.07. In contrast the K-enriched quartz tholeiites, with K contents only 10 times the K content of oceanic tholeiites, have K/Sr ratios increasing with increasing K content to a maximum near 90. The Rb/Sr ratio increases to 25—40 times the Rb/Sr ratio of the oceanic tholeiites. In contrast, tholeiites and quartz tholeiites of Hawaii do not show large enrichments in K_2O contents nor large changes in K/Sr or Rb/Sr ratios.

f) Summary

The slow ascent of magmas at considerable depth within the mantle, if accompanied by cooling and partial crystallization, would also cause selective enrichment in the incompatible element content (wall-rock reaction). Repetition of such processes would produce a mantle which is quite inhomogeneous in incompatible element content and, over geological time, would become inhomogeneous in Sr and Pb isotopic ratios. Less commonly, and possibly only in continental areas, slow ascent of magmas in the lower crust or uppermost mantle would be accompanied by partial crystallization and by selective enrichment in incompatible elements extracted from the wall-rock under conditions of lower pressures. The abundances of K, Rb, Ba and probably U, Th relative to major or trace compatible elements may be used to determine whether a given magma has been altered in chemistry by wall-rock reaction. The difference in partition behaviour of Sr at low and high pressures (consequent on the stability or instability of plagioclase) leads to the promising possibility of using such ratios as K/Sr or Rb/Sr to distinguish between the effects of deep-level wallrock reaction or a shallow-level process. In attempting the interpretation of isotope and trace-element abundance data there seems little relevance in the simple model of melting in which a liquid composition is determined at source (particularly in trace element and isotopic ratio contents) and thereafter remains essentially a

K/Sr and Rb/Sr for various average and individual basalts. Enrichment in incompatible elements in the f
essure, enrichment in the remainder is considered to have occurred at high pressure

Rb (ppm)	Rb/Sr	Sr (ppm)	Rock and locality	
55	0.4	137	Average	Pigeonite tholeiite, Antarctica
31	0.25	125	Average	Hypersthene tholeiite, Antarctica
33	0.25	130	Average	Hypersthene tholeiite, Tasmania
12	0.12	100		Olivine tholeiite, Antarctica
55	0.18	310		Columbia R. Plateau, Oregon
40	0.02	1,980		Basanite, Tasmania (Tertiary)
14	0.02	770		Alkali olivine basalt (Tertiary)
30	0.05	600	Average	Alkali basalt
21	0.015	1,436		Hawaiite, Hawaii
16	0.04	413		Alkali olivine basalt, Hawaii
10	0.02	515		Ankaramite, Hawaii
47	0.04	1,249		Alkali olivine basalt Hawaii
42	0.07	568		Alkali olivine basalt Samoa
32	0.04	643		Alkali olivine basalt Ascension
20	0.07	270		Tholeiitic basalt Tasmania (Tertiary)
10	0.04	250	Average	Hawaiian tholeiite
10	0.04	230	Average	non-submarine tholeiite
8	0.02	391		Quartz dolerite, Oahu
11	0.03	335		Quartz dolerite (vein), Oahu
5	0.02	297		Tholeiite, Mauna Loa, Hawaii
1.2	0.01	115	Average	oceanic tholeiite

closed system except for depletion by fractional crystallization. There are great uncertainties in use of trace element and isotopic data obtained from basaltic rocks rich in incompatible elements to deduce element abundances and geochemical pre-histories for the mantle source region and such attempts should be viewed with caution.

The Significance of Cognate Xenoliths in Basalt Petrogenesis

a) Peridotite Nodules

Coarse-grained inclusions of peridotite have a sporadic but world-wide occurrence as inclusions in alkali olivine basalts, basanites, nephelinites and melilite nephelinites. The most abundant type of nodule may be classed as lherzolite and has the mineral assemblage olivine (Fo_{85-92}), aluminous enstatite, aluminous chrome-diopside (also with appreciable Na_2O content) and minor spinel ($MgAl_2O_4$-rich). Other types, less common and of more restricted occurrence, ranging from dunite through wehrlite to clinopyroxene (FRECHEN, 1948, 1963), are probably genetically distinct (de ROEVER, 1963; WHITE, 1966).

The lherzolite inclusions commonly have the character of xenoliths of previously consolidated rocks showing internal effects of deformation, cataclasis, grain-boundary recrystallization and dimensional or crystallographic orientation of minerals (DEN TEX, 1963). There may be internal evidence of disequilibrium and reaction between phases and the minerals of the inclusion are clearly not in equilibrium with the enclosing basaltic liquid — orthopyroxene, spinel and clinopyroxene all showing clear evidence of partial reaction. The lherzolite nodules apparently do not occur in olivine tholeiitic magmas or their derivatives (WHITE, 1966). The proportions of olivine, pyroxenes and spinel are variable amongst individual nodules but the order of abundance is almost invariably olivine $\gg$ enstatite $>$ chrome-diopside $>$ spinel. Compilations of modal analyses by VILMINOT (1965) from Rocher du Lion, Velay, France and WHITE (1966) from Hawaii indicate a greater frequency of nodules with modes containing olivine from 90—65% and pyroxenes from 10—35%. The ratio of orthopyroxene to clinopyroxene is usually between 2:1 and 3:1. Both authors plot modal data obtained from the literature giving a wider scatter and including types with clinopyroxene in excess of olivine (cf. BROTHERS, 1960). Spinel is almost ubiquitous as a minor phase (1—4%) and accessory minerals noted in the literature on nodules and in a study (GREEN — in preparation) of Eastern Australian examples, include phlogopite, hornblende, sulphides, melilite, plagioclase (An_{90} and An_{65}; WHITE, 1966) and garnet.

Nodules show evidence of internal disequilibrium — thus some Australian examples show reaction between clinopyroxene and spinel at mutual grain boundaries to yield fine grained plagioclase, olivine and chromite aggregates. WHITE (1966, p. 289—293) notes that enstatites in some inclusions are of variable Al_2O_3 content and may be zoned in Al_2O_3 content. Comparison of the mineralogy and bulk chemical composition of lherzolite nodules with the chemical compositions of high temperature peridotite intrusions (GREEN, 1963, 1964) and with other peridotites (GREEN and RINGWOOD, 1963), coupled with the internal evidence of disequilibrium in the peridotites themselves, demonstrates that the mineral assemblage of the lherzolite inclusions is unstable at high temperatures at upper crustal levels. Preliminary experimental work indicates that the olivine + aluminous pyroxenes + spinel assemblage is stable at pressures > 8 kb approx. at 1100° C in compositions close to those of the lherzolite nodules. The temperature at which the mineral assemblages of individual nodules reached equilibrium is probably variable but the degree of solid solution of the two pyroxenes is very much less than the experimentally crystallized pyroxene pairs at 1300—1400° C (p. 138, Fig. 3). The CaO content of most orthopyroxenes is $<1.0\%$ CaO and commonly around 0.5% CaO (ROSS, FOSTER and MYERS, 1954; WHITE, 1966; GREEN and EASTON, unpublished data) and the orthopyroxene very commonly has exsolution lamellae of clinopyroxene indicating a history involving cooling from an earlier, higher temperature environment. The low CaO content of olivine ($<0.1\%$ CaO; WHITE, 1966) also suggests temperatures rather less than volcanic temperatures or the temperatures of the experimental high pressure runs previously reported. These features suggest that the lherzolite nodules were at temperatures of less than 1000° C before inclusion in their host basalts. The partition of Mg and Fe^{++} between co-existing pyroxenes in lherzolite nodules has been discussed by several authors (KRETZ, 1961; BARTHOLOME, 1961; DEN TEX, 1963; GREEN, 1964; WHITE, 1966) and, in some cases, used to suggest temperatures of equilibrium crystallization of the nodules similar to or greater than those of igneous, basaltic rocks. The distribution coefficients (K_D) for the analyzed orthopyroxene and clinopyroxene pairs (Tables 9, 11) obtained experimentally range from 1.14 to 1.27 and are thus distinctly higher than the values for igneous rocks (0.7 approx.) and somewhat higher than those of lherzolite nodules (around 1.0), cf. DEN TEX, 1963; WHITE, 1966. A considerable uncertainty in the use of mineralogical criteria for estimation of crystallization temperature is that different indicators may be "frozen in" at different stages of cooling.

Hypotheses of origin of the lherzolite nodules must be consistent with the above observations and may be broadly grouped into four possibilities:

1. Lherzolite nodules are accumulates of crystals precipitated from olivine-rich picritic or basaltic magmas at moderate to high pressures.
2. Lherzolite nodules are xenoliths of "parental mantle" from which a basaltic magma can still be derived under appropriate melting conditions.
3. Lherzolite nodules are xenoliths of residual, refractory mantle peridotite after basaltic magma has been removed.
4. Lherzolite nodules are xenoliths of mantle peridotite which have been selectively depleted in those "incompatible elements" which do not readily substitute in the major minerals of the peridotite.

In the simplest statement of the first hypothesis, a crystal accumulate relationship to the magmas *in which they occur* fails to account for the evidence of cataclasis, deformation and recrystallization in the lherzolite nodules. The estimated equilibration temperature of $<1000°$C and the compositions of pyroxenes and spinel are inconsistent with liquidus temperature equilibrium with a basaltic magma. It would be necessary to postulate an earlier precipitation from another magmatic event, cooling of the accumulates with variable cataclastic effects and accidental incorporation in a later magmatic event. The chemistry and mineralogy of the lherzolite inclusions would require that the initial precipitating phases were dominantly very magnesian olivine, with much smaller quantities of aluminous enstatite and/or sub-calcic, aluminous clinopyroxene. This type of mineral accumulate is possible from magmas such as the olivine basalt, olivine tholeiite or picrite compositions at pressures around 5—10 kb. We might expect such accumulative rocks to be initially inhomogeneous, possibly ranging from dunite to pyroxenite and also controlled in their bulk chemistry by the presence or absence of trapped, interprecipitate magma. Later cooling would be accompanied by exsolution of Al_2O_3 from aluminous pyroxenes to form spinel, exsolution of orthopyroxene from subcalcic clinopyroxene and could give rise to the observed mineralogy of the nodules. This hypothesis may be applicable to a limited degree but must rest on a postulated earlier history of the nodules which cannot be directly and unequivocally deduced from the present observed chemistry and mineralogy of the nodules.

The second hypothesis has been adopted by KUSHIRO and KUNO (1963) with calculations of the degree of partial melting needed to extract basalts of average Hawaiian and Japanese types from average lherzolite nodule composition. KUSHIRO and KUNO (1963) show that only 2—9% of basalt magma could be extracted from the lherzolite before one or more of the oxides TiO_2, P_2O_5 or K_2O goes to zero. Approximately 8% of an average high-Al_2O_3, oceanic tholeiite (ENGEL, ENGEL and HAVENS, 1965) could be extracted before TiO_2 and Na_2O of the average nodule composition used by KUSHIRO and KUNO go to zero. These small degrees of partial melting seem incompatible with the relatively high $Mg/Mg + Fe^{++}$ ratios of some basaltic liquids and pose mechanical difficulties in segregating such a small liquid fraction from residual crystals. The difficulties in obtaining basaltic chemistry, particularly the contents of incompatible elements such as K, U, Th etc. are even more formidable if average nodule compositions (e.g. VILMINOT, 1965) rather than selected individual nodules are used as model parental material.

The third hypothesis has been advocated by de ROEVER (1961, 1963), O'HARA and MERCY (1963) and OXBURGH (1964a, b) and is supported by the very low K_2O (MORGAN and GOODE, 1966), TiO_2, P_2O_5 and, to a lesser extent, Na_2O, contents of the lherzolite nodules. Referring to the previous discussion of partial melting in the mantle, the nature of residual peridotite will differ considerably depending on whether an olivine tholeiite or alkali olivine basalt magma is extracted. The appreciable Na_2O, CaO and Al_2O_3 contents and the opx:cpx ratios of 3:1 or 2:1 in very many lherzolites imply that, if the nodules are residua after basalt extraction, the magma extracted was of alkali olivine basalt type and not of olivine tholeiite type. The presence of glass (with crystallites of olivine, clinopyroxene and plagioclase) apparently developed at the expense of clinopyroxene + spinel in some Australian lherzolite inclusions, implies that at least some inclusions are capable of yielding a liquid fraction at magmatic temperatures and cannot be regarded as completely refractory residues. The third hypothesis, i.e. that lherzolite nodules are residual mantle from which a basaltic magma has been extracted, is probably valid for many lherzolite nodules. The previous arguments indicate that the magma extracted from pyrolite to yield compositions of the lherzolite nodules is of alkali olivine basalt rather than olivine tholeiite type. With this type of limitation the third hypothesis passes transitionally into the fourth, i.e. that lherzolite nodules are "residual" in character only in their content of incompatible elements but have not suffered a general partial melting process.

The fourth hypothesis has been suggested by HESS (1964) and further discussed by RINGWOOD (1966a) and GREEN (1967). The hypothesis postulates that lherzolite nodules are "residual", in comparison with pyrolite in their contents of elements such as K, Ba, U, Th, Rb, Sr, Zr, Ti, P and to a lesser extent, Na and the rare earths. The selective depletion in these elements, with little change in Ca, Al, Fe etc. could be a consequence of the process of wall-rock reaction previously discussed. Thus, fragments of pyrolite included in alkali olivine basalt magma would be expected to lose any very low melting fraction into the magma — it is assumed that this low melting fraction will be highly enriched in the incompatible elements K, Ba, U, Th, Rb, Sr, Zr, P and Ti. Of the major minerals of the lherzolite, clinopyroxene may partially react with loss of jadeite component to the liquid phase. By this hypothesis, the enrichment in incompatible elements in alkali olivine basalts is a complementary process to the depletion of these elements in any fragments of mantle accidentally incorporated into a body of magma. The third and fourth hypotheses, which differ in degree rather than in kind, probably together provide an adequate explanation of the mineralogy and chemical compositions of lherzolite inclusions.

b) The Restriction of Lherzolite Nodules to Critically Undersaturated Basaltic Magmas

Both of the preferred hypotheses postulate an initially accidental inclusion of peridotitic mantle material into ascending basaltic magma. The mechanical difficulty of upward transport of xenoliths of density 3.3 in magma of density approximately 2.7 may impose some limitation on the nature of host basalts. On the other hand, the present experimental work and discussion offers a direct reason for the restriction of the lherzolite nodules to undersaturated, alkali olivine basalts, basanites etc. and their absence in olivine tholeiites or quartz tholeiites. The data and discussion on basalt fractionation and on partial melting of pyrolite demonstrated that an alkali olivine basalt liquid may be in equilibrium with olivine, aluminous orthopyroxene and aluminous clinopyroxene near its depth of segregation at 35—70 kms (cf. GREEN and RINGWOOD, 1964). If lherzolite nodules have the character suggested in either the third or fourth hypotheses then they will behave essentially as refractory xenoliths in alkali olivine basalt with little internal melting and without disaggregation of the nodule.

In contrast to their behaviour in alkali olivine basalt magma, lherzolite nodules would be unstable if enclosed in an olivine tholeiite magma at 35—70 kms depth. An olivine tholeiite or tholeiitic picrite magma at these depths will be saturated in magnesian olivine and magnesian, aluminous enstatite but will completely dissolve any aluminous clinopyroxene introduced into it.

A lherzolite nodule which is included in an olivine tholeiite magma at 35—70 kms depth will partially melt with spinel, aluminous clinopyroxene and some aluminous enstatite entering the tholeiitic liquid, i.e. probably some 20—30% of the original inclusion would melt. Under these conditions a lherzolite inclusion would disintegrate although residual xenocrysts of magnesian olivine or chromite might remain, cf. MUIR and TILLEY (1964).

An additional reason for the absence of lherzolite nodules in tholeiitic magmas is that tholeiitic liquids extruded at the surface have in many cases undergone low pressure fractionation towards quartz tholeiites. The Hawaiian tholeiites illustrate this trend very well. Any magma which has fractionated by crystal settling at low pressure will no longer retain mantle xenoliths of high pressure mineralogy.

c) Pyroxene Xenocrysts and Pyroxenite Xenoliths

The experimental evidence on the important role of aluminous pyroxenes in basalt fractionation at high pressure suggests a critical examination of the xenolithic and xenocrystal material in basalts for natural examples of high pressure accumulates. The xenoliths of pyroxenite and "eclogite" occurring in olivine nephelinite tuff at Salt Lake Crater, Oahu, have assumed significance in some discussions of the hypothesis that the earth's upper mantle is of eclogitic composition. The experimental data previously presented offer a direct explanation of the occurrence, chemistry and petrography of these xenoliths as accumulates at high pressure and temperature from an undersaturated magma (KUNO, 1964; GREEN, 1966). The inclusions show features supporting an origin as sub-calcic clinopyroxene ± olivine accumulates from undersaturated, olivine-rich basaltic magma at pressures of 13—18 kb. Cooling, following initial precipitation, would cause a decrease in solubility of Al_2O_3 in the clinopyroxene and also result in exsolution of orthopyroxene from the low-calcium clinopyroxene. High temperature aluminous pyroxenes would exsolve Al_2O_3 as spinel initially but on further cooling into the stability field of pyrope-almandine garnet, spinel and pyroxene would react to yield garnet + olivine (cf. garnet coronas around spinel) or garnet may have exsolved directly from aluminous pyroxene. We infer that the Hawaiian "eclogites" were originally high pressure, high temperature accumulates from undersaturated magma and are non-equlibrium assemblages showing partial adjustment to lower temperature but still high pressure conditions. Both olivine (minor) and clinopyroxene (major) were probably included in the initial precipitate minerals.

KUNO (1964) described aluminian bronzite (4.4% Al_2O_3) and aluminian augites (4.9—10.3% Al_2O_3, 0.5—0.9% Na_2O) as phenocrysts or xenocrysts in an alkali olivine basalt or hawaiite from Japan and concluded that these phases crystallized from the magma at high pressures. The magma composition contains 50% SiO_2, 16% Al_2O_3 and is olivine + hypersthene normative. Our experimental data suggest that liquid and crystals of these compositions could be in equilibrium at about 9 kb. KUNO (1964) also refers to a quotation by HESS (1952) of an unpublished analysis of aluminous orthopyroxene (with almost 8% Al_2O_3) in alkali olivine basalt from western Texas. JAMIESON (1966) has reported olivine phenocrysts and orthopyroxene phenocrysts and aggregates in picritic basalts from the Nuanetsi province, Rhodesia. These orthopyroxenes are interpreted by JAMIESON as products of crystallization at high pressure from a picritic magma.

We have observed very rare coarse-grained pyroxenite (opx + cpx) xenoliths in lherzolite-bearing alkali olivine basalts from western Victoria. These show complex exsolution textures consistent with a high temperature origin and may be high pressure, high temperature accumulates from their host magma. Large single crystals of black, glassy-looking clinopyroxenes are not uncommon in basalts containing lherzolite nodules in eastern Australia and these may represent crystallization products at high pressure from their host magma.

These examples of "xenoliths" and "xenocrysts" are interpreted as natural examples of material crystallized from basaltic liquids at high pressure. Many other examples undoubtedly occur and the accumulation of data on such occurrences should provide an excellent linkage with experimental studies such as that presented in this paper.

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