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THE ORIGIN OF BASALTIC AND
NEPHELINITIC MAGMAS

by

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* *The Bennett Lectures*

THE ORIGIN OF BASALTIC AND
NEPHELINITIC MAGMAS

Preface by M. J. LeBas

In the past it has been the task of the 'hard-rock' geologist to observe and describe rocks as they occurred. In recent decades, the art of description (petrography) has been supplemented by the science of experimentation. The experiments have been directed at trying to imitate, on a laboratory scale, the processes which are believed to have occurred in nature.

The particular field of petrogenesis studied by Dr. Green is that of the basaltic magmas. The work has been done in the Department of Geophysics and Geochemistry within the Research School of Physical Sciences of the Australian National University at Canberra, Australia, where he is a Senior Fellow.

The origin of basaltic magmas which are so abundantly extruded from volcanoes over the whole world, undoubtedly gives us the key to the origin of all igneous rocks. It also tells us a great deal about the chemical and physical processes that have taken place over geological time in the Earth's crust and mantle. In the near future, we should be in a position to theorize on how the material beneath the crust of the earth, say that below Europe, has changed with geological time; perhaps quite a different state of affairs exists beneath the Atlantic.

In the 1960s the experimental investigations on basaltic rocks have become fairly sophisticated and so rather restricted to a specialist band of geologists, physicists and chemists. This being the case, it is necessary every now and again to explain to one's fellow geologists what is happening. Dr. Green's *Bennett Lecture* does just this.

Together with Professor Ringwood at Canberra, Dr. Green subjected rocks to high temperatures (1,000 - 1,500° C) and pressures (approx. 10,000 - 30,000 atmospheres, or 10 - 30 kilobars), and then 'watched' to see how the mineral components reacted during progressive crystallization under various physical conditions. From this study has emerged a hypothesis which is able to account for the known variations in time of the basaltic lavas of the world. This hypothesis, though it may be the leading one at present, is not without its rivals, and an interesting and instructive controversy is arising.

The lecture demonstrates that it is permissible to re-enact geological history in a crucible. Firstly, it reviews mineralogical evidence for the nature of the upper mantle which is considered to be the source of basaltic magmas, and defines the various magma types. Secondly, it describes what happens when model ultrabasic rock begins to melt, and traces various alternative consequences. Thirdly, it reconciles these observations on synthetic rocks with observations made on actual rocks such as the extrusive and intrusive basaltic magmas of Hawaii, Antarctica, and Iceland.

Dr. Green concludes by showing that, according to their experiments, it is possible for basaltic magmas to develop by partial (1% or more)

* Delivered in the Department of Geology in the University of Leicester on 10th November 1969.

melting of the ultrabasic upper mantle of the earth. He postulates on this evidence that the upper mantle has a particular composition, and "pyrolite" is the name he gives to the hypothetical rock. He further shows that primary melts derived from pyrolite can have different chemical compositions depending on the depth at which they were produced and the water content of the parent pyrolite — also, that each of these melts can change its composition through fractional crystallization. This will depend on the length of time that the melts remain at the various depths. Volcanic eruptions occur whenever some geological incident brings such melts to the surface.

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1. INTRODUCTION

Basaltic magmas are derived from the earth's mantle. Investigation of the origins of diverse magmas and deduction of the nature of the earth's upper mantle composition and mineralogy are two complementary problems. In this lecture, I will briefly review ideas on the nature of the Earth's upper mantle, as deduced from geological processes which sample the upper mantle and from constraints imposed on their source region by basalt magmas themselves. This will lead to derivation of a specific model composition for the upper mantle. The mineralogical variation with depth for this composition has been studied experimentally. The influence of a small (0.1 - 0.2%) quantity of water in the upper mantle source region is treated in some detail showing that this will lead to the formation of a low velocity zone within the upper mantle, characterized by very small degrees of partial melting.

The variations among natural basalts are briefly reviewed to select those magmas of direct mantle derivation. The experimental crystallization of such basaltic magmas under pressures and temperatures equivalent to upper mantle conditions and with various water contents, allows the recognition of conditions at which these magmas may be in equilibrium with olivine, enstatite, and possibly clinopyroxene, and spinel or garnet i.e. the residual phases of a source peridotite composition. In this way, an attempt is made to outline an internally consistent model of source composition, derivative liquids, peridotitic residues, and magmatic accumulates from the primary liquids under high and low pressure.

2. NATURAL PROCESSES SAMPLING THE UPPER MANTLE

(a) *Kimberlite Pipes* Kimberlite pipes transport xenoliths and xenocrysts of high-pressure type from depths of 80 kilometres or more. The dominance of peridotite xenoliths over rare eclogite (quartz-free), kyanite eclogite and grosspyroxite xenoliths has previously been noted (Ringwood 1966a). The peridotites themselves commonly consist of olivine + enstatite + rare chromite, olivine + enstatite + garnet, olivine + enstatite + clinopyroxene and olivine, enstatite, clinopyroxene and garnet. Magnesium-rich ilmenite, or phlogopite may occur as minor phases in some samples. The mineral assemblages of the peridotites indicate temperatures well below their solidus — this is apparent in the low Al_2O_3 content (1 - 2% Al_2O_3) of orthopyroxene coexisting with garnet and in the generally low degree of solid solution between orthopyroxene — clinopyroxene pairs (Boyd 1967, Green and Ringwood 1969). This is particularly important in evaluating attempts to use the olivine + orthopyroxene + garnet assemblage of some kimberlite inclusions to argue that residues from basalt magma extraction from the mantle source rock should be olivine + orthopyroxene + garnet \pm clinopyroxene (O'Hara 1968, Carswell and Dawson 1970)

rather than olivine + aluminous orthopyroxene + aluminous clinopyroxene (Green and Ringwood 1967a, b). However the garnet: pyroxene ratio of many kimberlite xenoliths is such that at temperatures near their solidi (200-600°C above their probable equilibration temperature, depending on the water content of the source rock at the time of melting), the garnet would be taken up in the aluminous pyroxene solid solutions giving olivine + orthopyroxene ± clinopyroxene mineralogy. This would be valid for depths of 80-150 kilometres in the upper mantle. There is no unique composition for peridotite inclusions in kimberlite, they exhibit a small range in $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}}$ ratio from ~89 to ~93, and relatively larger variations in CaO, Al₂O₃, Na₂O and TiO₂ contents. They are probably residue from basalt magma extraction but represent re-equilibration of the residual phases at temperatures well below the peridotite solidus. Their contents of minor elements such as Ba, Rb, K, etc. are likely to be grossly contaminated by their host 'magma' (kimberlite); this process can be seen in the growth of secondary carbonates, serpentine or phlogopite in some inclusions.

The inclusions of kimberlite thus suggest an upper mantle which is inhomogeneous and at least partially residual in nature and in which magnesian olivine and orthopyroxene are the most abundant phases. Their investigation appears unlikely to lead to the selection of a specific individual xenolith composition as representative of any part of the upper mantle.

(b) *Xenoliths in Alkali Olivine Basalts, Basanites and Nephelinites*

Olivine-rich and nepheline normative magmas quite commonly contain xenoliths of coarse-grained lherzolite with the typical mineral assemblage of olivine > enstatite > clinopyroxene > spinel. Hornblende, phlogopite, and apatite occur as accessory minerals in some examples. There is variation from sample to sample in the relative proportions of the major and minor minerals and in the presence or absence of particular accessory minerals. The pyroxenes are characteristically aluminous (2-5% Al₂O₃ in orthopyroxene) and the degree of solid solution between the two pyroxenes indicates temperatures below the solidus temperatures for the lherzolites. The spinel is aluminous spinel with variable chromite solid solution and the stability of the olivine + aluminous pyroxenes + spinel assemblage relative to lower or higher pressure alternatives indicates pressures at equilibration of > 8 kb and < 30 kb, probably < 23 kb. The lherzolite inclusions have a limited range of $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}}$ values from > 88 to < 92. Detailed geochemical and isotopic studies (see Cooper and Green 1969, for references) demonstrate that the lherzolites (at least the intensively studied examples) are accidental xenoliths with no cogenetic or cognate relationship to their host magmas. It has been inferred that the lherzolite inclusions include residual peridotite after magma extraction, but range from such refractory residue to materials which have not suffered partial melting and complete magma extraction but may have lost a small low temperature melt or fluid phase. Again it is possible to ascertain which lherzolites are not refractory and residual in character, only after detailed geochemical study, and it is not yet possible to select a specific lherzolite composition as appropriate for a mean upper mantle composition. Rather it appears that the mantle

source region for the lherzolite inclusions is appreciably inhomogeneous on the hand-specimen scale.

Other coarse-grained inclusions occur with the dominant lherzolite and include olivine orthopyroxenites, websterites, spinel websterites, and garnet clinopyroxenite. These compositions have lower $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}}$ ratios, decreasing approximately in the order listed above, and show clear evidence of exsolution and partial reaction and recrystallization of the assemblages in a high pressure cooling history following initial crystallization. These inclusions are interpreted as random xenoliths of small bodies of *high pressure* ($> 10 \text{ kb}$) accumulates and accumulates + entrapped liquid. Another group of xenoliths, which may occur with or without the lherzolite xenoliths described above, includes dunites ($\text{Fo}_{87.5} - \text{Fo}_{84}$) with minor chromian spinel, clinopyroxene and minor plagioclase, and possibly related wehrlites and olivine clinopyroxenites (Jackson and Wright 1970). This group contains more iron-rich olivine than the lherzolite inclusions and the mineral assemblages lack the clear evidence of a high pressure cooling history which is evident in the garnet pyroxenites. They are interpreted as accumulates or recrystallized accumulates from basaltic magmas at shallower depths (< 20 kilometres approximately). Inclusions of olivine gabbro, gabbro, and anorthosite (approximately An_{80}) are interpreted as accumulates from shallow magma chambers (< 10 kilometres). Ultimately the detailed study of xenolithic inclusions and their host magmas, together with knowledge of the crystallization sequence in various basalt magma types, should provide important evidence on the natural processes of crystallization and fractionation of basaltic magmas at various depths.

(c) *High-temperature peridotite intrusions* There are some peridotite intrusions, usually of dome or plug-like form and emplaced in metamorphic terrains, which have high temperature, dynamothermal, metamorphic aureoles. The peridotites themselves provide evidence of initial crystallization at high pressure and high temperature (to olivine + aluminous pyroxene + spinel) and subsequent movement as high temperature crystalline diapirs to emplacement in the crust at lower pressures. The mineralogy of these peridotites is olivine $>$ enstatite $>$ clinopyroxene $>$ spinel (high pressure assemblage). The major element composition and mineralogy is such that these rocks could potentially yield some proportion of basaltic magma by partial melting but detailed geochemical studies such as that of Frey (1969) show evidence for more complex history and selective depletion in some elements by an earlier melting or volatile-extraction process. Inhomogeneity within the peridotites due to lenses or bands of dunite or pyroxenite and modal variation within the peridotite again pose difficulty in selecting a valid mean composition for these mantle-derived peridotites.

To summarize, at least three geological processes sample materials which, because of their high pressure mineralogy, demonstrably crystallized under upper mantle conditions. All three processes indicate the importance of peridotitic composition with major olivine, enstatite as the second most abundant phase, essential clinopyroxene and a minor but essential alumina-rich phase which may be plagioclase, spinel or garnet depending on the final P, T conditions of equilibration. In all three processes, there is a

problem of representative sampling and inhomogeneity. The partially residual or depleted character of many samples, relative to a potential basalt source composition, has already been demonstrated. These factors illustrate the difficulty of natural mantle-sampling processes in determining the mean upper mantle composition.

3. THE PYROLITE MODEL FOR THE UPPER MANTLE

If basaltic magmas are liquids derived from the upper mantle as products of partial melting then it would be possible to estimate the source-rock composition if the compositions of liquid and residue and their relative proportions were known. This is the rationale of the 'pyrolite' model for the mean upper mantle composition and it is emphasized that this approach is a general one (Ringwood 1962). Green and Ringwood (1963) calculated a specific pyrolite composition by combining an average basalt composition (Nockolds 1954) and very refractory dunite in the proportions 1:3. The proportion 1:3 was chosen to yield Al_2O_3 and CaO contents similar to those of the natural mantle-derived peridotites of the previous section and also similar to an estimated upper mantle composition based on the chondritic earth model. Later experimental work on basaltic compositions permitted more restrictive limitations on the nature of liquids and residues in basalt genesis. Ringwood (1966a) calculated a second pyrolite composition using Hawaiian olivine tholeiite as the liquid and harzburgite (olivine 80%, enstatite 19%, chromite 1%) as the residue. These were combined in the 1:3 ratio to give the composition, shown in column 9 of table 1, which has orthopyroxene : olivine and pyroxene : Al_2O_3 ratios much closer to those of the mantle-derived peridotites. With the models of liquid compositions, nature of residues, and proportions of liquid : residue (which are developed later in this lecture) one could calculate many such 'pyrolite' compositions, and if the concept of a mean upper mantle composition (acting as the source for a variety of magma types) has any validity then these calculated compositions should be very similar. Also as more extensive studies of mantle-derived peridotites become available, these may provide closer constraints by indicating trends of increasingly residual character from 'pyrolite'. At the present time the pyrolite composition of Ringwood (1966a) appears adequate in deriving internally-consistent models of mantle composition and basalt genesis, at least to the level of precision currently possible.

The stability relations of various mineral assemblages for the pyrolite composition have been experimentally determined (fig. 1) (Green and Ringwood 1967a, 1969; Green and Hibberson 1970). The boundary for the disappearance of spinel at high temperature to yield the olivine + aluminous pyroxenes assemblage and the boundary for appearance of garnet from the latter assemblage are sensitive to bulk composition, particularly the pyroxene : R_2O_3 proportions. In the pyrolite composition of Green and Ringwood (1963), spinel persists to the solidus and there is no distinct inflexion in the boundary for the appearance of garnet from the olivine + aluminous pyroxenes + spinel assemblage. In considering partial melting of the pyrolite composition (Ringwood 1966a) under dry conditions, the first liquids will be in equilibrium with olivine + pyroxenes + plagioclase at low pressures, olivine + aluminous pyroxenes at intermediate pressures and olivine + aluminous pyroxenes + garnet at high pressures. With increasing degree of melting, the minerals in equilibrium with the

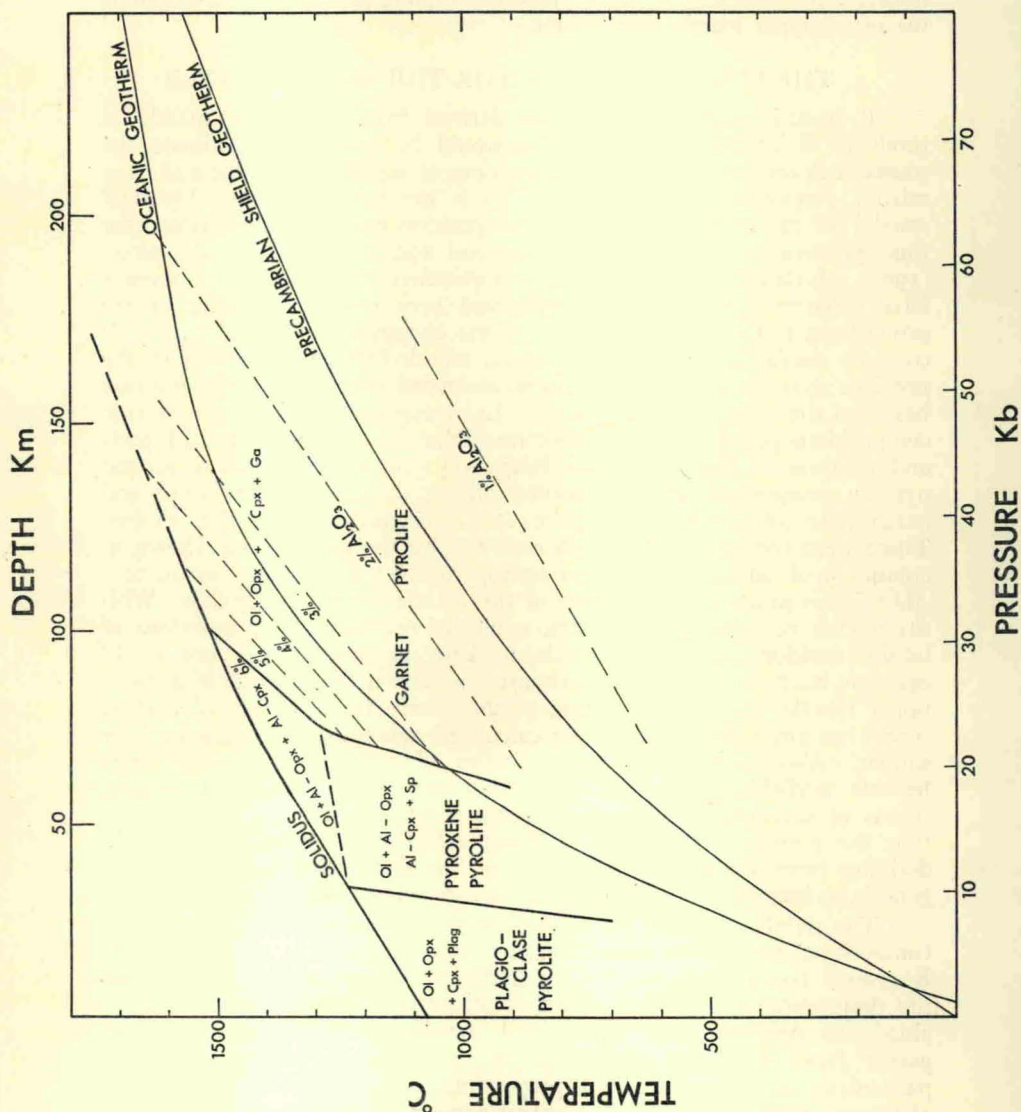


FIG. 1 (Green and Ringwood 1967a). Diagram illustrating the P, T, fields of different mineral assemblages in pyroxene composition under anhydrous conditions. The figures 1% Al_2O_3 , 2% Al_2O_3 , 3%, 4% etc. refer to the Al_2O_3 content of orthopyroxene in equilibrium with garnet, clinopyroxene and olivine in the garnet pyroxene field. Positions of two estimated geothermal gradients are also shown.

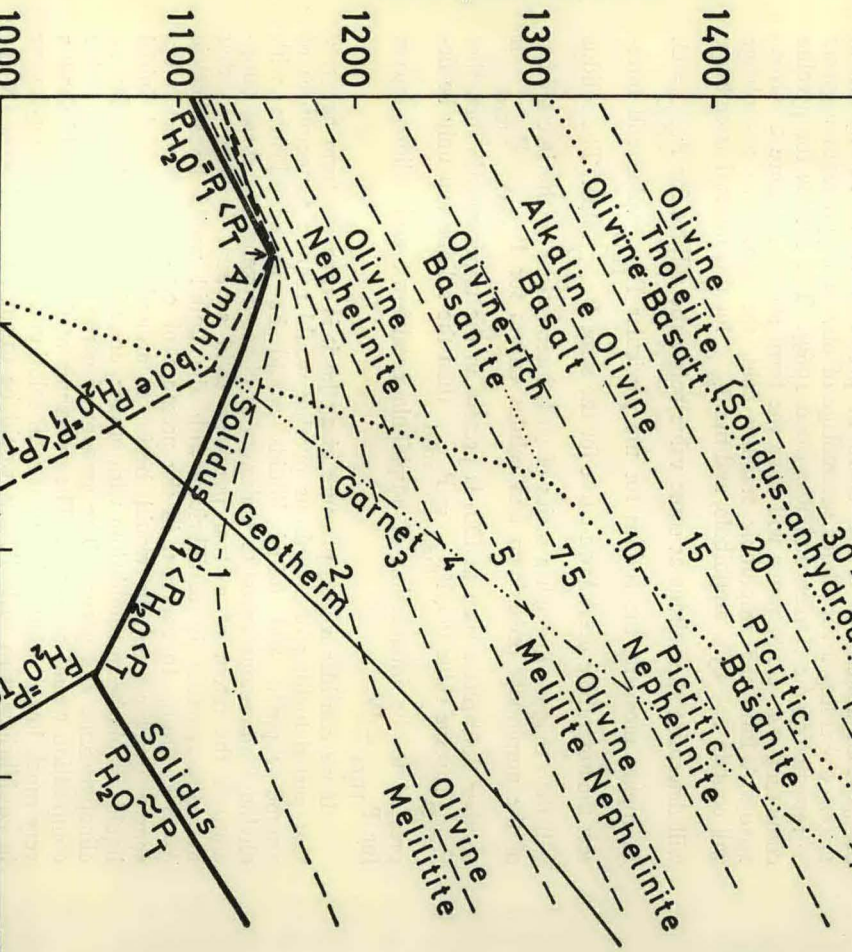
melt disappear in the order: plagioclase followed by clinopyroxene, clinopyroxene; and garnet followed by clinopyroxene (up to 36 kb) in the difference phase fields. Enstatite and major olivine remain as the residual minerals for quite high degrees of melting at pressures above 5 kb.

If the solidus of the pyrolite is lowered by the presence of water, then the characteristic mineralogy at the solidus and the nature of liquids developed in equilibrium with the near-solidus mineralogy will both differ. The presence of water causes crystallization of amphibole in the subsolidus mineralogy, but amphibole stability is limited at high pressure and this mineral breaks down to olivine + pyroxenes + garnet + water. Experimental studies indicate that for the pyrolite composition containing 0.1 - 0.2% H_2O , amphibole breaks down at pressures above 28-29 kb at $1,000^\circ C$ yielding a subsolidus assemblage of olivine + pyroxenes + garnet + rare phlogopite (Green and Ringwood 1969). The solidus for pyrolite containing 0.1 - 0.2% water will have the form shown in figure 2 (Green 1970). Within the amphibole stability field, the subsolidus assemblage will contain 10 - 15% amphibole, and the P_{H_2O} for this mineral assemblage will determine the solidus at some value below the solidus for $P_{H_2O} = 0$. At higher pressure, the solidus for the particular P_{H_2O} value will intersect the amphibole breakdown curve for the same P_{H_2O} value. The solidus will then migrate through a series of such intersections to the intersection of the amphibole breakdown and solidus curves for $P_{H_2O} = P_{total}$. If another fluid phase such as CO_2 is present in the upper mantle such that P_{H_2O} is some value $P_2 < P_{fluid} = P_{solid}$ then the solidus will only be depressed to the intersection of amphibole breakdown and solidus curves for $P_{H_2O} = P_2$.

If we consider a load pressure of 25 kb it is apparent from figs 1, 2 that partial melting of anhydrous pyrolite would require temperatures of around $1,450^\circ C$ and initial liquids would be in equilibrium with olivine + aluminous enstatite + aluminous clinopyroxene. At these temperatures the extent of mutual pyroxene solid solution is large so that the clinopyroxene in equilibrium with the liquid would be subcalcic ($\sim 11\% CaO$). In contrast, in the presence of 0.1 - 0.2% water in the source pyrolite, melting would begin at around $1,100^\circ C$ and initial liquids would be in equilibrium with olivine, enstatite ($\sim 3\% Al_2O_3$, fig. 1), clinopyroxene, and garnet. The amount of garnet present in the pyrolite composition would be $\sim 7\%$. The pyroxenes at the solidus would show a very much lower degree of mutual solid solution so that the clinopyroxene in equilibrium with early liquids would be calcium rich ($> 20\% CaO$).

An estimated geothermal gradient (Clark and Ringwood 1964, Ringwood 1966b) is illustrated in fig. 2. This does not intersect the anhydrous solidus for pyrolite but does intersect the solidus for pyrolite containing a small amount of water. Because the degree of melting at temperatures above the solidus is strongly controlled by the limited initial water content of the source rock (fig. 3), the amount of liquid present will remain small for a rather large temperature interval above the solidus. The effect of small quantities of water in the upper mantle is thus to cause a region beginning at depths of 80-100 kilometres in which there is

Temperature C°



normally present a small ($< 5\%$) melt fraction. It is considered that this region corresponds to the seismic low velocity zone, the small melt fraction producing low seismic velocity and high seismic attenuation. The lower boundary of the low-velocity zone may be caused either by decreasing water content or by entry of water into higher pressure silicate phases.

The discussion in this section has illustrated the very sensitive controls exerted by P , T and P_{H_2O} in determining the mineralogy and

solidus temperature of the model pyrolite composition. Attention in the following section is directed at the liquidus phases of basaltic magmas at high pressures since if these magmas are derived from the pyrolite composition by partial melting then the liquidus phases of the basalt and the residual phases in pyrolite at a given P , T and % melting must match one another. The study of the liquidus and near-liquidus phases of basalts at high pressure also shows the possible fractionation trends which a basalt magma, separated from its source peridotite, may follow at various depths in the upper mantle and crust.

4. CRYSTALLIZATION OF BASALTIC MAGMAS AT HIGH PRESSURE

(a) *Basalt nomenclature used* In studies of basalts where the chemical composition of a magma is the principal concern, it is convenient to adopt a normative (i.e. indirect chemical) classification rather than a modal (mineralogical) classification. The nomenclature used is as follows:

Tholeiite: basalt with normative hypersthene

Quartz tholeiite: basalt with normative hypersthene and quartz

Olivine tholeiite: basalt with normative hypersthene and olivine, hypersthene $> 3\%$

Olivine basalt: with normative olivine and with 0-3% normative hypersthene; no normative nepheline

Alkali olivine basalt: with normative olivine and nepheline; nepheline $< 5\%$

FIG. 2 (Green 1970). Diagram illustrating the amphibole stability limit at high pressure and the solidus for pyrolite composition containing approximately 0.1% H_2O . Also shown (dotted lines) are the solidus for anhydrous pyrolite and the boundary for appearance of garnet (from Green and Ringwood 1967a). The numbered dashed curves show the % liquid present above the solidus and the petrochemical character of these liquids is indicated. The geothermal gradient shown is the oceanic geotherm illustrated by Ringwood (1966b).

The positions of the curves on this diagram, other than the anhydrous boundaries, are estimated from reconnaissance experiments on the solidus and amphibole stability in pyrolite with 0.1-0.2% water and by experiments establishing the degree of depression of the liquidus temperatures of various basalts by addition of known amounts of water.

Ordinate — Temperature within the earth.

Abscissa — P_{total} = Load Pressure or Solid Pressure.

P_{H_2O} = equilibrium water pressure.

P_I = water pressure along solidus for pyrolite with 0.1% H_2O .

P_I varies along the solidus to P_I' and an arrow indicates the intersection of the solidus with a subsolidus breakdown curve for amphibole at $P_{H_2O} = P_I'$ also (amphibole \rightarrow pyroxenes + garnet + olivine + water).

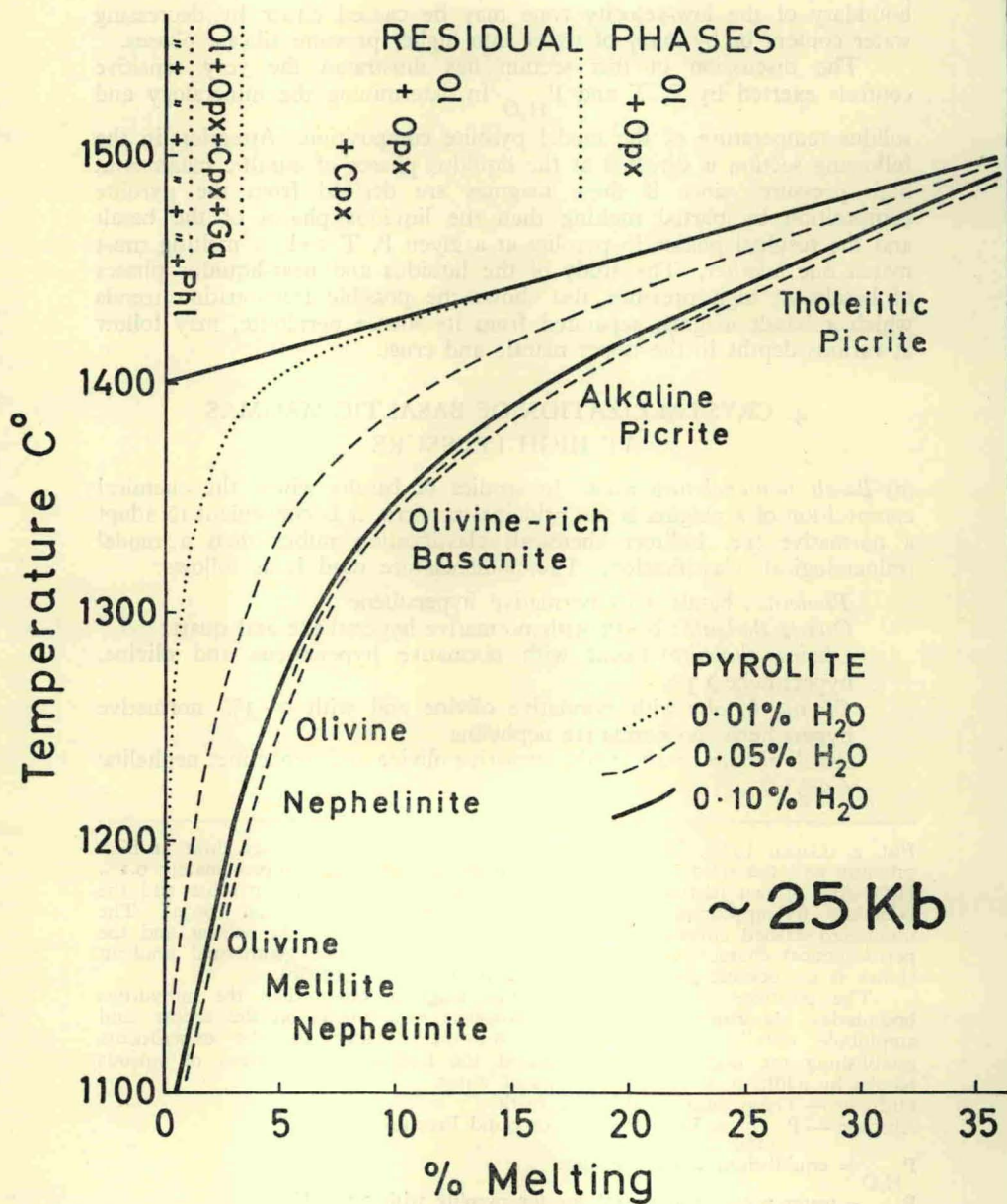


FIG. 3 (Green 1970). Diagram contrasting the melting behaviour of anhydrous pyrolite with melting in pyrolite compositions containing 0.01, 0.05 and 0.10% H₂O. The nature of the residual phases and the petrochemical character of the liquids derived are shown for example with 0.10% H₂O.

Basanite: basalt with normative olivine, nepheline and albite and with nepheline $> 5\%$, albite $> 2\%$

Olivine nephelinite: basalt-like composition with major normative olivine and nepheline; albite $< 2\%$; normative orthoclase and/or leucite, but no normative larnite

Olivine melilite nephelinite: basalt-like composition with normative olivine, nepheline, leucite and larnite.

The principal variation between the above magma types is the degree of silica saturation. Most workers recognize continuity and transition in chemical composition between the various basalts. Silica content decreases from 47-49% in the olivine tholeiites to $< 38\%$ in olivine melilite nephelinite; alkalis ($\text{Na}_2\text{O} + \text{K}_2\text{O}$) increase and CaO also increases, particularly in the nephelinitic compositions. A further variation in basalt compositions is apparent in the recognition of distinctive high-alumina basalts (Tilley 1950; Kuno 1960). These are aphyric basalts with normative olivine and either normative hypersthene or low normative nepheline contents but with Al_2O_3 contents distinctly higher (at 16-20% Al_2O_3) than 'normal' olivine tholeiites or alkali olivine basalts of similar SiO_2 and $\text{Na}_2\text{O} + \text{K}_2\text{O}$ contents (Kuno 1960).

(b) *Selection of relevant basaltic magmas* The detailed studies of the active Hawaiian volcanoes convincingly demonstrate the importance of low pressure, near-surface crystal fractionation in determining the composition of the magma finally extruded at the surface. In selecting basalts for study under simulated upper mantle compositions, it is clearly necessary to seek those liquids which have been derived from the upper mantle without opportunity for crystal fractionation at shallow levels. One group of basalts obeying this restriction is the suite from alkali olivine basalts, through basanites, olivine nephelinites, olivine melilite nephelinites to olivine melilitites, which contain high density ($\rho \sim 3.3$) xenoliths and xenocrysts of high pressure mineralogy. These magmas clearly cannot have undergone any crystal fractionation at pressures less than those at which the xenoliths were picked up. As a group, these magmas are Mg-rich, rather high in normative olivine, and with SiO_2 contents ranging from $\sim 49\%$ SiO_2 (some olivine trachybasalts) to $\sim 36\%$ (olivine melilitites). There is of course no reason to suppose that this whole spectrum is related by crystal fractionation or by different degrees of partial melting at any one pressure (depth) but there clearly must be processes operative in the upper mantle capable of yielding each magma type by partial melting of pyrolite or by crystal fractionation of a higher temperature magma at some depth within the upper mantle. The experimental petrologist can directly evaluate the possible inter-relationships for these magmas.

In selecting an olivine tholeiite magma for experimental study at high pressures one cannot use the high-pressure inclusion criterion since this class of basalts does not contain such inclusions. However, the detailed studies of the 1959-60 Kilauea Iki eruption of Hawaii have provided evidence for the relatively rapid movement of a magma batch from depths of approximately 60 kilometres to the earth's surface, and eruption of that magma batch in episodic extrusions which ranged from very olivine-rich picrite to olivine-poor tholeiite. These episodic eruptions could be monitored for volume and magma composition, allowing an estimate of the mean composition of the lava lake in which the extrusions accumulated. The mean composition was estimated as an olivine-rich

	(Green and Ringwood, 1967b)				(Bultitude & G)	
	1	2	3	4	5	6
Olivine Tholeiite	46.95	47.05	45.51	45.39	44.63	44.50
	2.02	2.31	1.93	2.52	2.92	1.30
	13.10	14.17	12.44	14.69	11.67	12.00
	1.02	0.42	0.92	1.87	2.95	0.80
	10.07	10.64	8.67	12.42	9.39	10.60
	0.15	0.16	0.15	0.18	0.15	0.20
	14.55	12.73	18.79	10.37	13.85	16.70
	10.16	9.87	9.67	9.14	7.68	9.90
	1.73	2.21	1.64	2.62	3.65	2.80
	0.08	0.44	0.08	0.78	2.00	0.80
	0.21		0.20	0.02	1.03	0.40
	72.00	68.10	79.40	59.80	72.50	73.70
$\frac{\text{Mg}}{\text{Fe}^{++}}$	0.60	2.70	0.50	4.50	11.70	4.50
Norms	14.70	18.90	13.90	18.00	12.00	6.20
				2.20	10.20	9.40
	27.60	27.30	26.30	26.20	9.80	18.00
	17.00	17.60	16.50	15.70	18.70	22.80
	12.30	1.30	2.80			
	21.90	27.20	34.60	25.80	25.80	34.40
	3.80	4.40	3.70	4.80	5.60	2.40
	1.40	0.60	1.30	2.90	4.30	1.20
	0.50		0.40		2.00	1.00

tholeiite containing ~18% normative olivine (Macdonald and Katsura 1961) or 17% normative olivine (Murata and Richter 1966). Green and Ringwood (1967b) used an olivine tholeiite composition modelled closely on Macdonald and Katsura's estimate as a reasonable inferred composition for an olivine tholeiite magma derived from the upper mantle with little or no low pressure fractionation.

(c) *Crystal fractionation—olivine tholeiite to alkali olivine basalt* The compositions and CIPW norms of some of the basalt types which have been studied experimentally are listed in Table 1. The experimental results for olivine tholeiite (fig. 4) show olivine as the liquidus phase at low pressure, orthopyroxene at intermediate pressure, clinopyroxene at about 20 kb, and garnet at 27 kb. Electron probe analyses of the near-liquidus crystals and estimation of the percentage crystallization allow calculation of the derivative liquid compositions. In this way it has been shown that, at atmospheric pressure, parental olivine tholeiite will yield derivative quartz tholeiite liquids, at about 9 kb derivative liquids will be characteristically high-alumina olivine tholeiites, and at 13-18 kb derivative liquids will move rapidly into the nepheline-normative field. The validity of the major role of orthopyroxene in defining the trend from parental olivine tholeiite to derivative alkali olivine basalt was confirmed by experiments on an olivine basalt, an alkali olivine basalt, and on a picrite composition, the latter being saturated in olivine to higher pressures than the other three compositions but having orthopyroxene as the second phase to crystallize at 13-18 kb. It is emphasised that the calculation of crystal fractionation trends must be carried out on all oxide components of the basalt—plots of oxide trends, or projections into 'simple component planes' can be extremely misleading and should be used only for illustrative and not deductive purposes. In fig. 5, the molecular norms of the experimental compositions and of their calculated derivatives (Green and Ringwood 1967b) have been plotted as they would appear on a plane approximately parallel to the base of the 'basalt tetrahedron' (Qz-Ol-Cpx-Ne) of Yoder and Tilley (1962). The calculated fractionation trends are shown as solid lines—these show the possible fractionation behaviour of a magma batch of the chosen composition if cooled at the pressures indicated. Thus at 13.5 kb the olivine tholeiite (OT₁) crystallizes orthopyroxene alone near the liquidus and fractionation produces the trend illustrated. At the same pressure, a second olivine tholeiite (OT₂), prepared by adding 5% more olivine (FO₉₀) to the initial olivine tholeiite, crystallizes olivine alone (see fig. 4). However, at 13.5 kb it obviously cannot fractionate by olivine separation *through* the point OT₁. Thus before the liquid crystallizes 5% olivine, orthopyroxene appears and dominates the further crystal extract giving the trend shown as a dashed curve for 13.5 kb. A similar relationship is apparent between the picrite (PB) and olivine basalt (OB) at 13-18 kb. These compositions and their crystallization paths illustrate clearly that although olivine must appear (at appropriate P, T) with orthopyroxene as a liquidus phase for any liquid derived by partial melting of a pyrolite source rock, the olivine is a minor phase in the crystal accumulate formed if the magma separates and begins to crystallize at depths equivalent to 12-18 kb. The type of crystal accumulate formed would be an olivine orthopyroxenite. When clinopyroxene joins the precipitating phases, any olivine not removed from contact with the liquid may react with liquid. A reaction (olivine + liquid₁ → clinopyroxene + liquid₂) has been observed

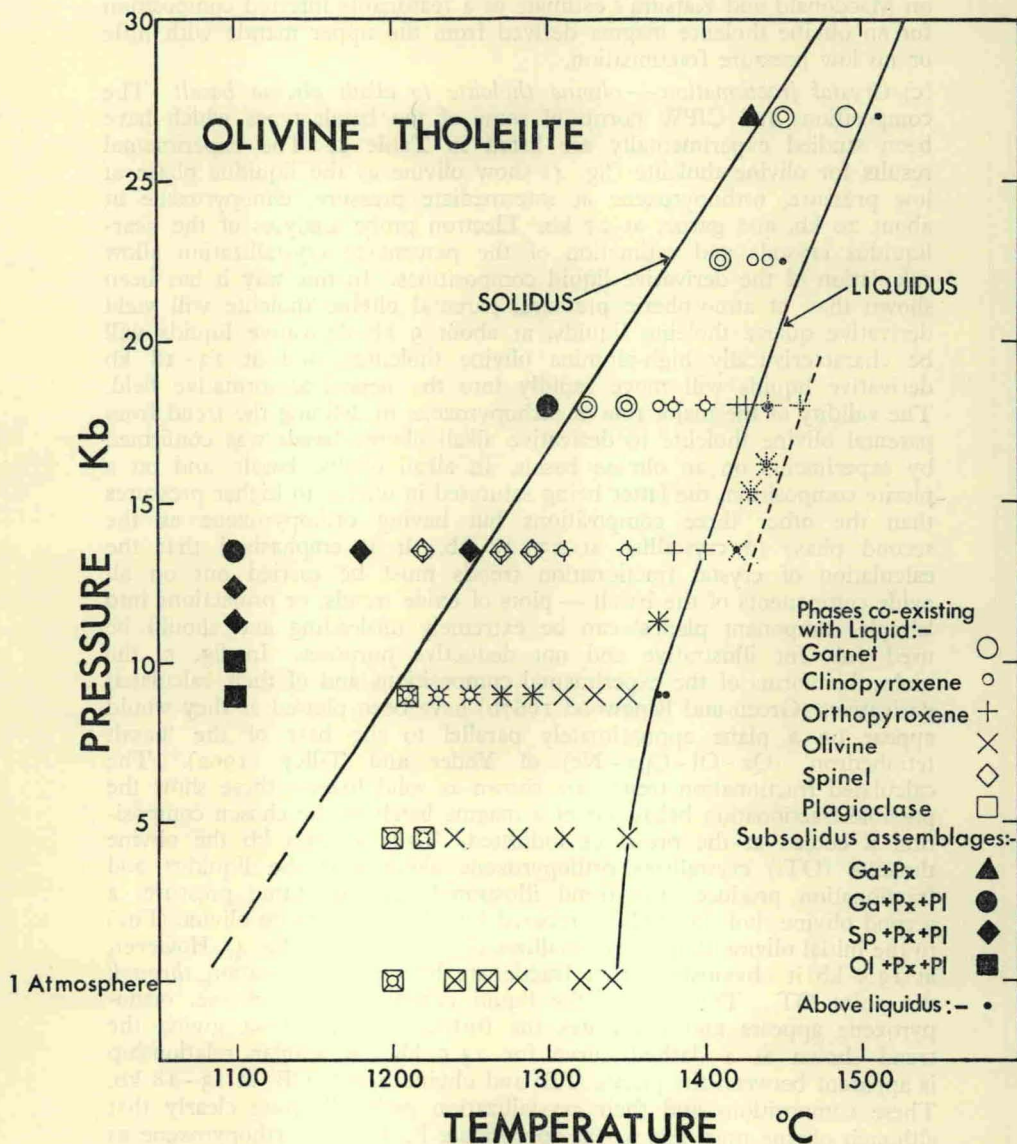


FIG. 4 (Green 1969b). Experimental crystallization of olivine tholeiite (Table 1) at various pressures. Data (dotted symbols) for an olivine-enriched tholeiite was prepared by addition of 5% olivine (Fo₉₀) to the olivine tholeiite of Table 1.

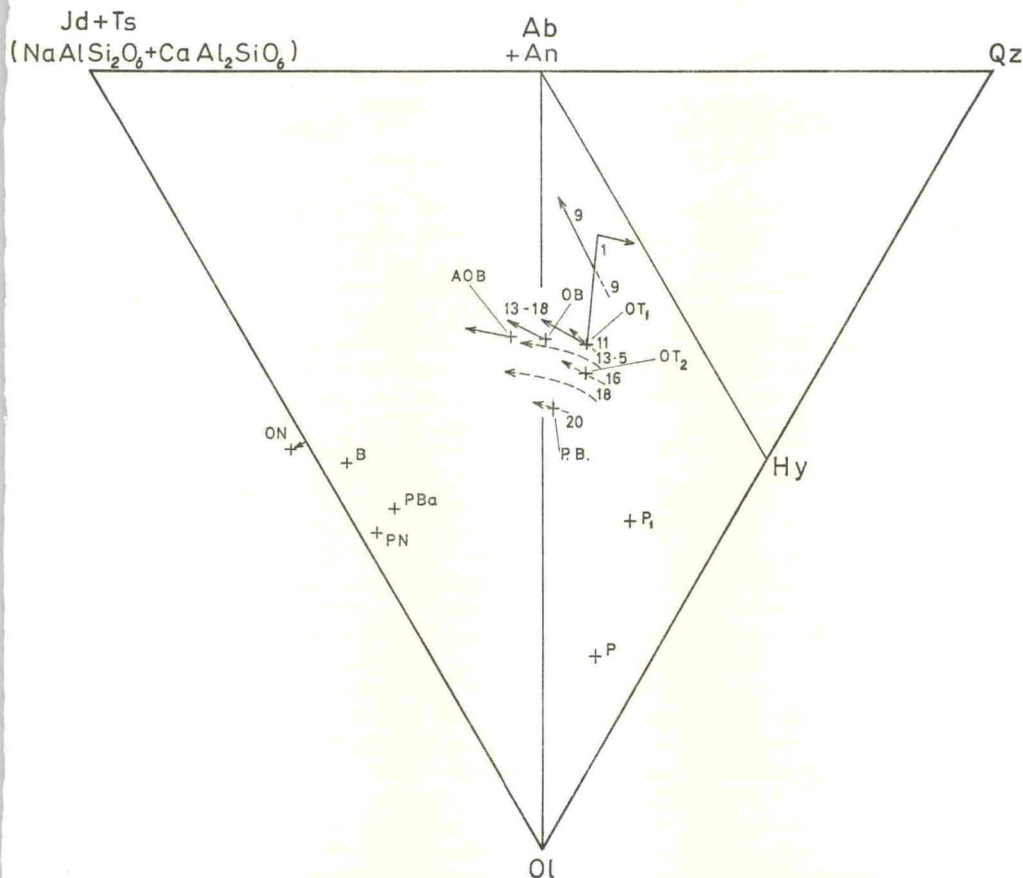


FIG. 5 Plot of experimental compositions of Table 1 and calculated fractionation trends, on the surface (Jadeite + Ca-Tschermak's silicate) — Olivine — Quartz. Co-ordinates calculated from the Ab, An, Ne, Hy, Ol components of the molecular norm as follows:

$$\text{Jd} + \text{Ts} = \text{Ab} + \text{An} + \text{Ne}$$

$$\text{Qz} = \text{Ab} + \text{An} - \text{Ne} + \frac{\text{Hy}}{2}$$

$$\text{Ol} = \text{Ol} + \frac{\text{Hy}}{2}$$

OT₁ = Olivine tholeiite

OT₂ = Olivine tholeiite (OT₁) + 5% olivine Fo₉₀

OB = Olivine basalt

AOB = Alkali olivine basalt

PB = Picrite

PBa = Picritic basanite

B = Olivine-rich basanite

ON = Olivine nephelinite

PN = Picritic nephelinite

P = Pyrolite

P₁ = Pyrolite — 40% Olivine

Solid lines: calculated fractionation trends (Green and Ringwood 1967b) at various pressures (9kb, 13-18 kb).

Dotted lines: estimated trends for liquids in equilibrium with olivine and orthopyroxene at various pressures (numbers indicate the pressures).

in some undersaturated compositions at pressures up to 31.5 kb (Bultitude and Green 1970). Accumulates lacking primary olivine and consisting only of aluminous orthopyroxene + aluminous clinopyroxene may thus be formed, and at lower temperatures the disappearance of orthopyroxene and later appearance of spinel or garnet may yield clinopyroxenite, spinel clinopyroxenite or garnet clinopyroxenite accumulates.

(d) *Crystal fractionation — basanite to olivine nephelinite* The compositions of some highly undersaturated basaltic magmas which have been studied experimentally are given in Table 1 and plotted on Fig. 5. The compositions were selected from amongst basalts containing high pressure lherzolite xenoliths, and thus should closely approach the compositions of magmas existing in the upper mantle. All are olivine-rich and, under dry conditions, olivine is the liquidus phase up to 20 kb or to 32 kb depending on the specific composition. Clinopyroxene is the second phase to crystallize over a large pressure range and becomes the liquidus phase over a small pressure interval in the olivine nephelinite and olivine-rich basanite compositions. At higher pressure, clinopyroxene is either joined by garnet on the liquidus or garnet becomes the liquidus phase, joined by clinopyroxene slightly below the liquidus. Details of the experimental studies, with analytical data on the near-liquidus phases are given by Bultitude and Green (1970). In the picritic basanite composition, orthopyroxene appears in the crystallization sequence at 18 kb over a very small temperature interval after the crystallization of 5 - 8% olivine and prior to appearance of clinopyroxene. In the picritic nephelinite composition at 22.5 kb, orthopyroxene again makes a very brief appearance in the crystallization interval, after the precipitation of some olivine and clinopyroxene. Orthopyroxene does not appear in the crystallization sequences for the olivine nephelinite or olivine-rich basanite at any pressure under dry conditions.

The fractionation trends at various pressures have been calculated from the analytical data on the liquidus phases. Low-pressure fractionation is dominated by olivine and yields derivative liquids of higher silica content and rapid iron enrichment. At pressures of 18 - 27 kb, fractionation of these particular liquids is dominated by clinopyroxene with minor olivine or garnet. Derivative liquids become more undersaturated but there is marked calcium depletion accompanying the silica depletion. At pressures greater than 27 kb crystal fractionation is controlled by garnet + clinopyroxene separation. Evaluation of fractionation controlled by these phases is particularly important since O'Hara (1968) considers that eclogite (i.e. garnet + clinopyroxene) fractionation is the process by which magmas such as alkali olivine basalts, basanites, nephelinites, etc. become highly enriched in 'incompatible elements' (such as K, P, Ti, Rb, Sr, U, Th, etc.) relative to olivine tholeiites. O'Hara (1968 pp. 117 - 118) suggests 50% eclogite fractionation will be important in trace element enrichment but will have little effect on the major element composition of the basalts. O'Hara and Yoder (1967, p. 110) similarly invoke eclogite fractionation to account for enrichment in trace elements, but also invoke eclogite fractionation from parental tholeiitic picrite to produce residual liquids with abundant olivine, nepheline, larnite, leucite and kalsilite in the CIPW norm. The quantitative evaluation of fractionation by garnet + clinopyroxene separation at pressures of 27 - 36 kb from the picritic basanite and nephelinite compositions by Bultitude and Green (1970) shows that large

degrees of crystallization are required to produce small changes in silica content and in degree of undersaturation. Such fractionation is accompanied by depletion in calcium and enrichment in iron relative to magnesium. These effects are inconsistent with characteristics on natural magmas of mantle derivation which show increasing Ca-contents with increasing degree of undersaturation and demonstrate similar or higher

$\frac{\text{Mg}}{\text{Mg} + \text{Fe}^{++}}$ values for the very undersaturated magma types, relative to olivine tholeiites. Separation of garnet and clinopyroxene under upper mantle conditions undoubtedly occurs in specific cases (xenoliths of garnet clinopyroxenite in the Delegate or Salt Lake Crater tuffs are in part probably examples of such accumulates) but it is concluded that it is not the process responsible for the natural magma series from olivine-rich tholeiite to olivine nephelinite and olivine melilitite.

Experiments on the effect of water on the liquidus phases of the basanite and nephelinite compositions of table 1 (Bultitude and Green 1968, Green 1969a, b) show that whereas olivine and clinopyroxene or clinopyroxene alone may be the liquidus phases under dry conditions (liquidus 1,400 - 1,450° C, 20 - 30 kb), if the liquidus is depressed to 1,200 - 1,300° C by the addition of 2 - 5% water, then olivine and orthopyroxene, or olivine, orthopyroxene and clinopyroxene may occur at or near the liquidus. This effect has been substantiated for the olivine-rich basanite composition and an alkali olivine basalt composition with experiments using sealed capsules and known quantities of water (Green 1969a, b, Green and Hibberson 1969). It should be noted that the near-liquidus clinopyroxene under dry conditions at 1,400 - 1,450° C is rather sub-calcic, and with the rapid widening of the pyroxene miscibility gap at lower temperature, such a composition would be represented by orthopyroxene and clinopyroxene. Thus depression of the liquidus by the addition of water may bring in orthopyroxene as a liquidus phase, not as a consequence of any major compositional shift of the cotectic between olivine and orthopyroxene (cf. Kushiro 1969), but as a direct consequence of the sensitive temperature dependence of the pyroxene solid solutions.

The important near-liquidus role of orthopyroxene in water-bearing basanitic and nephelinitic magmas at high pressure may permit fractionation of magmas through olivine-rich basanites to olivine nephelinites at 20 - 25 kb and through picritic basanites and picritic nephelinites to olivine melilitite nephelinites at ~ 27 kb.

5. APPLICATION OF BASALT CRYSTALLIZATION STUDIES TO PARTIAL MELTING

If we consider an olivine tholeiite magma then the experimental studies (Green and Ringwood 1967b) have shown that by separation of approximately 15% crystals at 13 - 18 kb we may produce an olivine basalt magma type. The crystal extract would be dominantly orthopyroxene with very minor olivine. Further crystallization (~15%) of orthopyroxene, now joined by sub-calcic clinopyroxene and possibly still accompanied by very minor olivine, will produce a liquid of alkali olivine basalt type. If crystals have remained in contact with the liquid, then we have a chemical system in which an alkali olivine basalt liquid is in equilibrium with minor olivine, minor sub-calcic clinopyroxene and major orthopyroxene. The proportion of crystals to liquid is approximately 30:70 but it will not

affect the composition of either liquid or crystals nor the equilibrium between them if we grossly change the proportions of the crystals adding a very large amount of olivine, a little more orthopyroxene but no more clinopyroxene so that the proportion of crystals to liquid becomes approximately 80:20. We are now considering a total composition which is peridotitic, and have a model of partial melting or batch melting of that total composition. By increasing the temperature on the peridotitic bulk composition we can envisage all the clinopyroxene, major orthopyroxene and minor olivine entering the melt and changing the melt composition back through olivine basalt to reach olivine tholeiite at a stage of about 30% melting.

This discussion illustrates the way in which the experimental studies on basalt crystallization can be applied to a model of partial melting. This reasoning can *only* be applied if the observed liquidus phases of the series of basaltic liquids are the same as the observed phases in the postulated source composition at the same P and T. This particularly applies to

$\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ ratio — a basalt with olivine of Fo₈₀ as the liquidus phase could not be in equilibrium with a peridotite having olivine Fo₈₀ under any conditions. In this example a prediction that a basalt with similar normative mineralogy but suitably higher $\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ ratio would have olivine of Fo₈₀

as liquidus phase at the same pressure, is probably valid. The situation is more complex with the pyroxenes, however, for the extent of mutual solid solution of calcium-rich and calcium-poor pyroxenes is very dependent

on both temperature and $\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ ratio. Thus if a particular basalt had a clinopyroxene of $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} = 80$ as the liquidus phase at a particular

pressure it is difficult to predict whether a more magnesian basalt of the same normative mineralogy will have orthopyroxene or clinopyroxene or even olivine as the liquidus phase at the same pressure, since the pyroxene solid solution limits will be considerably different (less extensive) for the more magnesian pyroxenes. This discussion illustrates the caution that is necessary when attempting to extrapolate from simple iron-free systems to natural basalts, and emphasizes the limitations of projections in which Mg and Fe are equated and on which basalts, peridotites, and mineral

compositions of very different $\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ ratio are plotted with attempts

to derive generalized liquidus phase fields and cotectics (e.g. O'Hara 1968). Such techniques may have illustrative value if applied to closely coherent groups of compositions, but remain qualitative and invalid for deductive purposes unless augmented by treatment of the total chemical composition.

(a) *Pyrolite as a source rock for the experimentally studied basalts* In the discussion of basalt crystallization and fractionation it was suggested that by about 30% crystallization of olivine tholeiite at 13 - 18 kb, derivative alkali olivine basalt magmas could be obtained. The accumulate would be dominated by aluminous orthopyroxene, with minor sub-calcic clinopyroxene and minor olivine. Referring to the detail of the analyzed basalts (Green and Ringwood 1967b), the olivine tholeiite and olivine basalt compositions (table 1) had liquidus olivine and orthopyroxene with

$\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} \sim 90$, whereas the more iron-rich alkali olivine basalt had liquidus olivine (9 kb) and orthopyroxene (13.5 kb) with $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} = 83$. In general terms, liquids which could be in equilibrium with the pyrolite ($\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} = 89$) or residual assemblages from the pyrolite

must be relatively magnesian basalts i.e. $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} \sim 68 - 73$. In this respect the alkali olivine basalt composition of Table 1 is unsuitable for direct derivation from a pyrolite source, but a more magnesian composition of similar normative mineralogy is predicted to show similar crystallization behaviour with a slightly higher liquidus temperature and possibly slightly enhanced role of orthopyroxene crystallization at > 13 kb.

Referring to fig. 1, it is apparent that the phases which were invoked in the postulated crystal fractionation of olivine tholeiite to alkali olivine basalt are the same as the phases at the anhydrous pyrolite solidus within the pressure range 10 - 13 kb. This applies to a more detailed comparison of

$\frac{\text{Mg}}{\text{Mg} + \text{Fe}}$ ratios and of alumina contents of the pyroxenes (Green and Ringwood 1969). Furthermore, the limited partial melting data on the pyrolite (Green and Ringwood, 1967a) shows that clinopyroxene rapidly enters the liquid and disappears from the residue leaving olivine and enstatite only. The studies on basalt crystallization and mineralogy of pyrolite at high pressure may be integrated to suggest a model of partial melting at depths of 13 - 18 kb. At these depths, if the pyrolite source rock is anhydrous, then the first liquid at the solidus will be in equilibrium with olivine, aluminous orthopyroxene, and aluminous clinopyroxene — a liquid saturated with these phases will be nepheline-normative, i.e. an alkali olivine basalt or, at low degrees of melting at about 18 kb, an olivine-rich basanite (Bultitude and Green 1970). Up to 15 - 20% melting, clinopyroxene will remain among the residual phases and the liquids will remain nepheline-normative, though decreasingly so with increasing temperature. Since all the phases existing in the source rock are Fe-Mg solid solutions, it is considered that melting with increasing temperature will be nearer to a linear process (as illustrated in fig. 3) than a eutectic and 'stepped' process (O'Hara 1968) although different 'gradients' of the various solid-solution cotectics may produce some non-linearity in the % melting *vs.* temperature curve. For degrees of melting greater than 15 - 20%, clinopyroxene will be absent from the residual phases and the liquid, saturated only with olivine and aluminous enstatite, will become increasingly hypersthene normative changing through olivine basalt to olivine-rich tholeiite ($\sim 22\%$ olivine, 15% hypersthene, at 13 kb; $\sim 27\%$ olivine, 13% hypersthene at 18 kb).

At lower pressures (~ 9 kb) the experimental studies permit a similar deduction that liquids developed at 15-20% melting will be in equilibrium with olivine, enstatite (lower Al_2O_3 content), and minor clinopyroxene and such liquids will be high-alumina olivine tholeiites (10-15% normative olivine). With higher degrees of melting, clinopyroxene, orthopyroxene and minor olivine enter the melt changing it to olivine tholeiite of low Al_2O_3 type with high normative olivine and hypersthene contents. At higher pressures,

~ 22 kb, initial melts may be olivine-rich or picritic basanite, and change through alkali picrites to tholeiitic picrites with increasing melting.

The preceding discussion has considered partial melting in a source pyrolite composition which is anhydrous. However, it has been inferred in previous sections that some mantle derived magmas, such as olivine nephelinite or olivine melilitite, are hydrous magmas at their depth of origin implying that the mantle source region is itself water bearing. The effects on the near-solidus mineralogy and solidus temperature of water contents of 0.1 - 0.2% have been discussed in earlier sections, and now must be integrated with the basalt melting data under both wet and dry conditions.

(b) *Partial melting of a hydrous pyrolite mantle: andesitic or nephelinitic magmas?* Turning to the nature of liquids derived by partial melting of a hydrous peridotite source, there are two hypotheses in the current literature which could hardly be more diametrically opposed or mutually exclusive. Each is based on experimental data which are probably correct. Kushiro *et al.* (1968) and Kushiro (1969) have studied melting relations in the systems enstatite + water, and forsterite + diopside + silica + water, and inferred that enstatite melts incongruently to forsterite + quartz normative liquid under high water pressures (up to 30 kb water pressure). A similar incongruent melting relationship for enstatite is inferred for water undersaturated conditions i.e. $P_{\text{H}_2\text{O}} < P_{\text{total}}$. These results are

extrapolated to the melting of more complex olivine + enstatite + clinopyroxene-bearing assemblages in the presence of a minor water-rich fluid phase in the upper mantle, and lead to the hypothesis that liquids developed from such a source will be silica-oversaturated, specifically, quartz tholeiites or andesites. Such silica-rich magmas ($\text{SiO}_2 > 50\%$) might also form from parental olivine tholeiite by crystal fractionation at high water pressures (Kushiro *et al.* 1968).

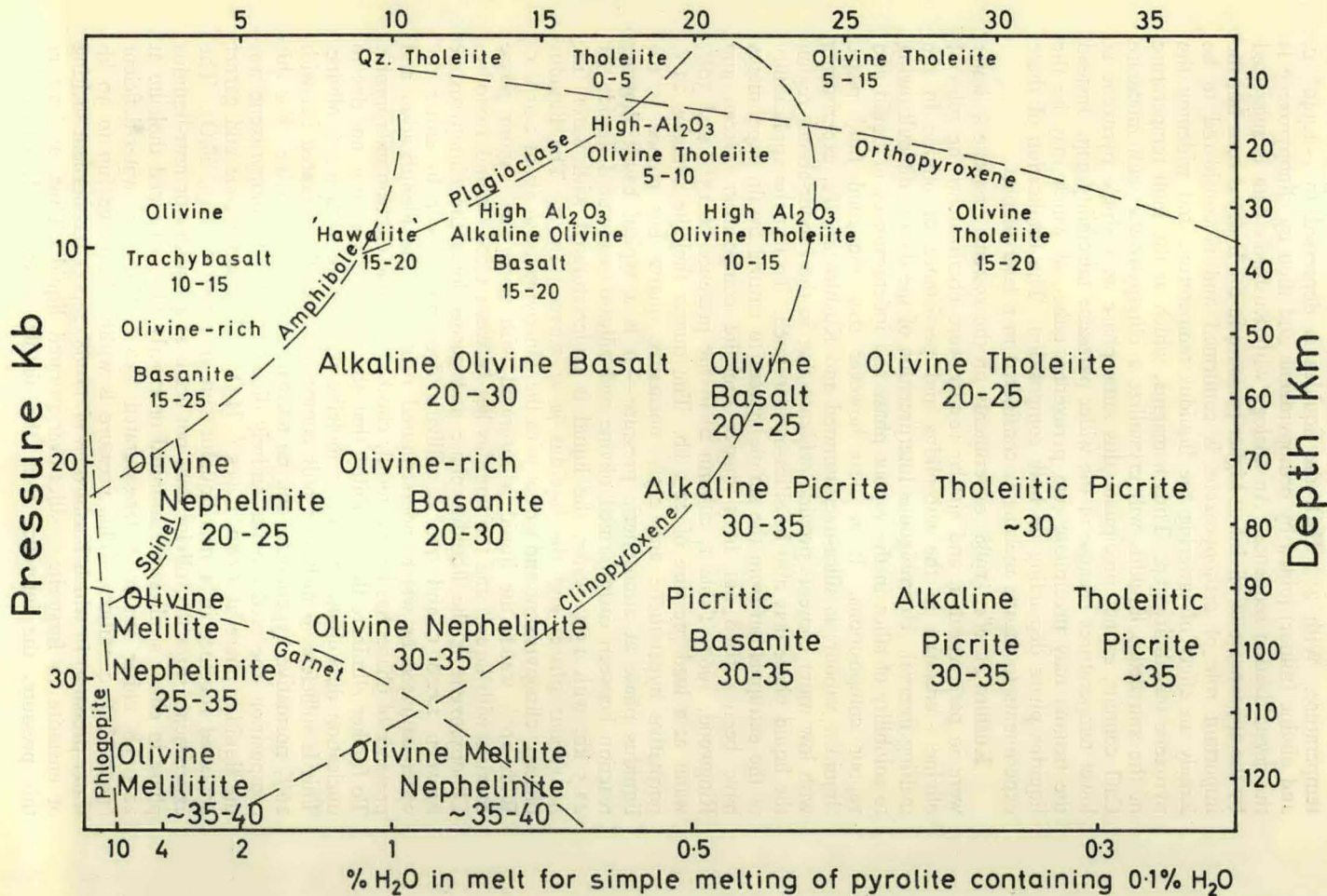
In contrast, Bultitude and Green (1968) had experimented on the role of water in depressing the liquidus of very undersaturated natural magmas (olivine nephelinites, Table 1), and reported a major role for orthopyroxene as a near-liquidus phase at 20 - 30 kb, $P_{\text{H}_2\text{O}} < P_{\text{total}}$, temperature 150 - 250° C below the anhydrous solidus. From these studies it was inferred that highly undersaturated olivine melilitite nephelinite, olivine nephelinite, and olivine-rich basanite magmas could form by partial melting of a pyrolite source (i.e. be in equilibrium with residual olivine, orthopyroxene, ± clinopyroxene, ± garnet) at pressures of 20 - 30 kb. It was inferred that parental olivine tholeiite could fractionate to these silica-poor magmas ($\text{SiO}_2 \sim 42\%$) by crystal fractionation in the presence of small but essential water contents.

Kushiro (1969) considered that Bultitude and Green's data were in error due to the experiments being carried out in crimped (rather than welded) Pt capsules with the result that the sample composition had altered during the run. Precautions against this had been taken by Bultitude and Green, and the use of the microprobe to analyze charges after runs effectively restricted any such sample leaching or contamination to very selective migration of alkali elements. Confirmation of the effect, using sealed platinum capsules and known quantities of water, was obtained for olivine-rich basanite composition (Table 1, column 5, reported in Green 1969a (footnote) and 1969b). In anhydrous runs on this composition, clinopyr-

oxene is the liquidus phase at 1,470° C and 27 kb, joined by garnet at lower temperature. With 2% H₂O, the liquidus is depressed to ~1,340° C, and olivine (minor) joined by orthopyroxene and then by clinopyroxene is the crystallization sequence. An analogy with Kushiro's data is suggested by the appearance of olivine at higher pressure in the hydrous runs but the important role of orthopyroxene is confirmed and is considered to be largely an effect of lowering the liquidus temperature and widening the pyroxene miscibility gap. Thus a magma, which at the high temperature of the anhydrous liquidus will crystallize a clinopyroxene with moderate CaO content, cannot precipitate this same phase as a single pyroxene at lower temperatures because of the wider pyroxene miscibility gap. Instead the magma may precipitate two pyroxenes, either of which may be the liquidus phase depending on bulk composition. The implication of these experiments to mantle melting is considered in a later section.

Kushiro *et al.*'s. (1968) experiments on the systems enstatite + water were in part repeated, and similar results were obtained on the roles of olivine + enstatite in the subsolidus and persistence of olivine in the melting interval. Unambiguous interpretation of the data is difficult owing to solubility of silica in the vapour phase and indeterminacy of liquid and vapour compositions. It is clear however that the total fluid phase (liquid + vapour) is silica-oversaturated and Kushiro *et al.*'s. experiments with low water contents provide evidence that in the vapour-absent region the liquid phase may also be silica-oversaturated. To test the applicability of the extrapolation from the simple system to natural basalt, experiments have been carried out in a quartz tholeiite composition (Green and Ringwood, 1967, Table 1, column 2) in the presence of 15% and 30% water at a load pressure of 22.5 kb. The quartz tholeiite contains 21% normative hypersthene and 2.5% normative quartz but olivine is the liquidus phase at atmospheric pressure — it is a typical basalt showing reaction between early-formed olivine and silica-oversaturated liquid. At 22.5 kb, with 15% water, the liquid is undersaturated with water (i.e. no vapour phase) and the liquidus is at 1,070 ± 15° C. The liquidus phase is clinopyroxene and garnet joins the clinopyroxene at about 1,050° C. With 30% water, the liquid is saturated with water (i.e. vapour phase coexists with liquid) and the liquidus lies between 1,000° C and 1,050° C. Clinopyroxene is the liquidus phase and garnet joins the clinopyroxene between 1,000° C and 970° C. Olivine does not appear in either the water-saturated or water-undersaturated runs, and the identification and presence of phases has been carefully checked with the electron microprobe. To further eliminate the possibility that absence of olivine was not due to nucleation difficulties, the quartz tholeiite was seeded with 6.4% olivine. This is sufficient to make the bulk composition exactly saturated i.e. with 26% normative hypersthene and no normative quartz or olivine. For this composition, the seed olivines entirely disappeared and clinopyroxene was the liquidus phase at 1,080° C (15% H₂O) and clinopyroxene and garnet appeared together in a near-liquidus run at 1,000° C (30% H₂O). The experiments show that olivine does not occur as a liquidus or near-liquidus phase in either a silica-oversaturated or exactly silica-saturated tholeiite at 22.5 kb either under water-saturated (30% H₂O) or water-deficient (15% H₂O) conditions. This pressure is within the range (up to 30 kb water pressure) in which Kushiro *et al.* (1968) report incongruent melting of enstatite to forsterite + silica-oversaturated liquid. Thus, at least at this pressure, the inference from the simple system study cannot be

% MELTING



extrapolated and applied to a typical basalt. It is considered most improbable that a quartz-normative tholeiitic or andesitic liquid can coexist with olivine (i.e. have olivine as its liquidus phase) under upper mantle conditions ($P > 10$ kb) even at high water pressures.

Thus the hypothesis that quartz tholeiitic or andesitic liquids may be derived by direct partial melting of a source peridotite composition in the presence of water in the upper mantle is rejected, and the contrasted hypothesis that such wet melting produces extremely undersaturated magmas from the pyrolite source is pursued in the following section.

(c) *Nephelinitic magmas from partial melting of hydrous pyrolite*

Experiments referred to in previous sections have shown that very undersaturated basanite and olivine nephelinites may form in equilibrium with residual magnesian olivine, orthopyroxene and calcic clinopyroxene at pressures of 18 - 35 kb, provided temperatures of melting are depressed 100 - 250° C below the anhydrous solidus by the presence of water. Olivine nephelinite magmas may form at 18 - 22 kb in equilibrium with residual olivine, enstatite, and clinopyroxene. The presence of garnet in the residual phases with olivine, enstatite, and clinopyroxene at pressures of ~ 27 kb will produce early liquids of low alumina and high calcium content (i.e. olivine melilite nephelinite and olivine melilitite). It may be noted that the temperature—pressure regime under discussion (i.e. $T \sim 1,200^\circ \text{C}$, $P > 18$ kb) is largely outside the stability field of amphibole, so that this phase does not occur at the solidus, particularly at pressure > 25 kb. For conditions ~ 18 kb amphibole will appear at the solidus at sufficiently high water pressure—liquids in equilibrium with amphibole at low degrees of partial melting ($P = 10 - 18$ kb) are probably potassium-rich and range from basanites near 18 kb to olivine trachybasalts near 9 kb.

6. SUMMARY OF PARTIAL MELTING RELATIONSHIPS

Figures 2, 3 and 6 present, in diagrammatic form, an integrated model for derivation of basalt types ranging from olivine trachybasalt and hawaiite to olivine melilitite and to tholeiitic picrite. It is emphasized that the magma types shown are restricted to magmas of direct mantle derivation. Processes of crystal fractionation at crustal levels or at deeper levels may modify these magma types producing other magmas not represented in these figures. Also it is *not suggested* that *all* rocks classifiable as 'hawaiite', quartz tholeiite or alkali olivine basalt etc. are derived as partial melts from pyrolite in the manner indicated. Rather the figures illustrate the petrochemical characteristic of the particular liquid formed *if partial melting of pyrolite occurs* at the designated conditions. In general terms, liquids formed in equilibrium with pyrolite source rock must be very

magnesian ($\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}} = 68 - 73$), whereas if any liquid segregates from

FIG. 6 (Green 1970). A petrogenetic grid for mantle-derived basaltic magmas. Various basalt magma types are assigned to a % Melt, Pressure grid (implying also specific % H_2O and temperature of melting, see figs. 2 and 3) in which they are regarded as partial melting products of a pyrolite composition containing 0.1% H_2O . The numbers placed with each basalt type refer to the normative olivine content of this liquid at its depth of origin—because of the expansion of the olivine crystallization field at low pressure most basalts will precipitate olivine before other phases if fractionation occurs at *lower* pressures. The dashed boundaries marked with a mineral name show that this mineral will occur among the residual phases remaining after extraction of magma types to the left of the boundary. Olivine is present in equilibrium, i.e. is a residual phase in the pyrolite composition for all the magma types, cf. fig. 3.

the source rock and begins to cool and crystallize, then separation of olivine, pyroxenes, garnet, or amphibole will all produce iron-enrichment in the derived liquids. The ratio $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}}$ for a basalt is thus a sensitive indicator of the prehistory of a particular magma, but there is a practical difficulty in that any oxidation of iron in the magma or rock will produce falsely high $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}^{++}}$ ratios.

In figs. 2, 3 and 6, tholeiitic magmas are produced with high degrees of partial melting, with olivine and enstatite remaining as residual phases. The water contents of these magmas, if they have developed from a source rock with 0.1% H₂O, would be only 0.3 to 0.4% H₂O. This water content would depress the tholeiite liquidus to only 30 - 50° C below the anhydrous liquidus. Tholeiitic magmas would form from a pyrolite source with about 0.1% H₂O at temperatures only 30 - 50° C below the temperature at which similar melts would form from *anhydrous* pyrolite and little difference in chemical composition would be expected between the two tholeiites. This is very different from the situation when very low degrees of partial melting are considered (fig. 3) for then the relatively water-rich magmas form at very much lower temperatures than anhydrous melts and the liquid-residue equilibria may be significantly different.

The conclusions on basalt petrogenesis outlined in figs. 2, 3 and 6 will undoubtedly require refinement and modification as further detailed studies proceed. However, these figures are intended to give a framework in which the increasingly detailed studies of natural basalt provinces can be viewed with greater insight as to possible magma source conditions and magma inter-relationships. Finally the acceptance or rejection of petrogenetic models such as figs. 2, 3 and 6 will be based on the natural basalts themselves, and there is a great deal of information still to be obtained. A specific example of how the P, T history of a particular magma (an olivine basalt from Auckland Is.) can be deduced is presented by Green and Hibberson (1969), and the same techniques can be applied to a very large range of natural, 'xenocryst'-bearing magma compositions. In the example quoted, it was shown that olivine, orthopyroxene, and clinopyroxene closely matching the compositions of 'xenocrysts' in the olivine basalt, were liquidus and near-liquidus phases for the host magma composition at 14 - 16 kb and 1,200 ± 20° C. For the basalt liquidus to occur at that P, T, then the basalt must have contained 2% water.

Furthermore, since the liquidus phases had $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} = 86 - 87$, the liquid could not have been a direct partial melt from pyrolite with $\frac{100 \text{ Mg}}{\text{Mg} + \text{Fe}} = 89$, but had probably already fractionated some way along the path towards alkali olivine basalt by separation of the observed orthopyroxene, clinopyroxene, and minor olivine at depths of about 50 - 55 kilometres.

7. TRACE ELEMENTS IN BASALTS — WALL-ROCK REACTION AND SOURCE INHOMOGENEITY

In the preceding discussion, the emphasis has been on the major element composition of basaltic magmas, since this determines the mineralogy and forms the basis of classification of basaltic rocks. In attempting to relate olivine tholeiitic and alkali olivine basalt magmas by crystal fractionation or different degrees of melting of a pyrolite source, Green and Ringwood (1967b) noted that there was a group of elements including K, Ti, P, Ba, Sr, Rb, Zr, Hf, U, Th, Pb, and the lighter rare earth elements, which show enrichment factors in Hawaiian alkali olivine basalts which are greater than would be produced by 30% crystallization of a parental olivine tholeiite or the difference between 20% melting (alkali olivine basalt) and 30% melting of the pyrolite source rock. These highly enriched elements ("incompatible elements") are those which do not substitute readily in the major minerals of the upper mantle, but probably form rare accessory phases. These elements are expected to be strongly partitioned into any melt fraction, so that for partial melting in excess of about 5% their concentration should closely reflect the %-melting and source composition only, and be essentially independent of the relative proportions of olivine, enstatite and clinopyroxene remaining in the residue.

Green and Ringwood (1967b) suggested an additional mechanism of concentrating incompatible elements in a fractionating magma by a process called 'wall-rock reaction'. In this process, it was envisaged that a body of magma could, under some conditions, cool by reaction with, and solution of, the lowest melting fraction of any wall-rock material with which it was in contact. This process was considered to be an important one for bodies of magma in a region of little temperature contrast between magma and wall-rock (i.e. a magma body near its depth of segregation from source pyrolite or for a large magma body with a relatively long residence time in the crust and undergoing crystal fractionation at low pressures). The latter aspect of wall-rock reaction was applied particularly to the fractionated and 'incompatible element enriched' quartz tholeiites of Tasmania and Antarctica.

With the recognition that the genesis of the undersaturated basanites, nephelinites, etc. (the magma with the highest enrichment factors for the incompatible elements) requires the presence of water, a further factor in selective element enrichment was suggested (Green 1969a, b). If hydrous accessory minerals such as phlogopite, amphibole, apatite, etc. are stable in the upper mantle then there exist conditions with $P_{\text{H}_2\text{O}}^{\text{wall-rock}} < P_{\text{fluid}}^{\text{wall-rock}} \leq P_{\text{solid}}$. A body of magma moving through such an environment may be very undersaturated in water and $P_{\text{H}_2\text{O}}^{\text{magma}}$ may be less than $P_{\text{H}_2\text{O}}^{\text{wall-rock}}$. Hydrous phases in the wall rock may break down under the influence of the neighbouring magma and water may migrate into the magma. The concept of wall rock reaction was modified to include the movement of a fluid phase containing incompatible element concentrations or diffusion of incompatible elements through a fluid phase from wall-rock to magma.

In an important paper, Gast (1968) discussed the trace-element enrichment in basalt magmas derived by partial melting, and showed that the strongly enriched 'incompatible element' contents of alkali olivine basalts, basanites, etc. would result if these magmas were produced by

very low degrees of melting (3-7%) of their source peridotite. Olivine tholeiites from the mid-oceanic ridge environments, with very low potassium and other incompatible element contents, were regarded as derived by 20-30% melting of a source rock which had previously suffered partial melting. Hawaiian olivine tholeiites, enriched in incompatible elements when compared with many oceanic ridge olivine tholeiites, were grouped with the alkali olivine basalts as products of smaller degrees of partial melting. This suggested relationship is inconsistent with figs. 4 and 6. Fig. 6 could reasonably account for the differences between Hawaiian magma types on the basis of different degrees of partial melting (provided alkali olivine basalts are in general formed by up to 20% melting, and commonly much less than this) of a pyrolite source rock, but the same source rock could not also yield the oceanic tholeiites (high-alumina olivine tholeiites) with their characteristically depleted incompatible element contents, by 15-20% melting in a single-stage process.

Consideration of incompatible element variations between rocks which are essentially similar in major element content, leads to the concept of inhomogeneity in incompatible element contents in the source region. Gast's (1968) suggestion of 2-stage melting, the first extracted melt being of small volume but highly enriched in incompatible elements, may be important. An additional suggestion has been made that the upper mantle source region may be vertically zoned in incompatible element concentrations (Green 1970) due to upward migration within the low velocity zone of a fluid phase. This fluid phase is envisaged as water-rich and possibly CO₂-rich, or possibly, a very small, water-rich, silicate melt fraction. The presence of a free fluid phase within the low velocity zone may permit the upward migration of incompatible elements, which do not readily substitute in olivine, pyroxenes, or garnet, until they are 'fixed' by entering a small silicate melt fraction in the upper part of the low velocity zone. In this way, source regions with 'depleted' and 'enriched' patterns of incompatible element abundance could be developed in the lower or upper part respectively of the low velocity zone. The particular tectonic or dynamic mechanism of partial melting (i.e. mantle diapirism, 'fault' tapping of the low velocity zone etc.) would determine which source region was involved in magma production.

8. CONCLUSIONS

The understanding of basalt magma genesis and of upper mantle composition are mutually dependent problems and both require a close integration of studies of geochemical and geophysical aspects of natural basalts, and experimental studies under simulated (P, T, P_{H₂O}) upper mantle conditions. I have, in this lecture, developed a model of basalt genesis in which the source composition (pyrolite) is regarded as essentially homogeneous in major elements but inhomogeneous or chemically zoned in content of incompatible elements and accessory minerals. The low velocity zone of the upper mantle is interpreted as a region in which, due to the instability of amphibole in pyrolite containing 0.1-0.2% H₂O at depths in excess of 80-100 kilometres, there is a very small (<5%) degree of partial melting. The nature of this liquid fraction is highly undersaturated olivine nephelinite or olivine melilitite. Ascent and extrusion of basaltic magmas may occur when the low velocity zone is tapped rather directly

by major rift or fault systems. More voluminous magmas, which are also usually products of much greater degrees of partial melting, form when there is diapiric upwelling of material from the low velocity zone to higher levels in the mantle or lower crust (Green and Ringwood 1967; Green 1970).

Magma types ranging from tholeiitic picrite to olivine melilitite and olivine trachybasalt are assigned to a petrogenetic grid in which the depth (pressure) of magma segregation, the degree of partial melting of the source rock, the water content, and approximate temperature of the magma, are specified. While this model can account for variations in major element abundances and normative mineralogy among basalts, there are variations in abundances of K, Rb, Ba, the rare-earths, etc. (incompatible elements), which are inconsistent with a model involving a constant source composition for all mantle derived basalts. Abundances of incompatible elements in the magmas are partly determined by the source composition at magma segregation but may be further enriched by wall-rock reaction processes in the upper mantle or in the continental crust. Source inhomogeneity appears to be required to account for variations in incompatible element abundances and such inhomogeneity may be a characteristic feature within the low velocity zone, the upper part of which is considered to have 'enriched' and the lower part a 'depleted' pattern of incompatible element contents.

REFERENCES

- BOYD, F. R., 1967. Electron probe study of diopsidic pyroxenes from kimberlite. *Carnegie Inst. Washington Year Bk.*, **65**, 252-260.
- BULTITUDE, R. J. and GREEN, D. H., 1968. Experimental studies at high pressures on the origin of olivine nephelinite and olivine melilitite nephelinitic magmas. *Earth Planetary Sci. Lett.*, **3**, 325-337.
- BULTITUDE, R. J. and GREEN, D. H., 1970. Experimental study of crystal-liquid relationship at high pressures in olivine nephelinite and basanite compositions. *J. Petrol.* (In press).
- CARSWELL, D. A. and DAWSON, J. B., 1970. Garnet peridotite xenoliths in South African kimberlite pipes and their petrogenesis. *Contr. Mineral. Petrol.*, **25**, 163-184.
- CLARK, S. P. and RINGWOOD, A. E., 1964. Density distribution and constitution of the mantle. *Rev. Geophys.*, **2**, 35-88.
- COOPER, J. A. and GREEN, D. H., 1969. Lead isotope measurements on lherzolite inclusions and host basanites from Western Victoria, Australia. *Earth Planetary Sci. Lett.*, **6**, 69-76.
- FREY, F. A., 1969. Rare earth abundances in a high temperature peridotite intrusion. *Geochim. Cosmochim. Acta.*, **33**, 1429-1447.
- GAST, P. W., 1968. Trace element fractionation and the origin of tholeiitic and alkaline magmas types. *Geochim. Cosmochim. Acta.*, **32**, 1,057-1,086.
- GREEN, D. H., 1969a. The origin of basaltic and nephelinitic magmas in the earth's mantle. *Tectonophysics*, **7**, 409-422.
- GREEN, D. H., 1969b. A review of experimental evidence on the origin of basaltic and nephelinitic magmas. *Phys. Earth Planetary Interiors*, **3**, 221-235.
- GREEN, D. H., 1970. Compositions of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism. *Phil. Trans. Roy. Soc. London*, (In press).
- GREEN, D. H. and HIBBERSON, W., 1969. Experimental duplication of conditions of precipitation of high pressure phenocrysts in a basaltic magma. *Phys. Earth Planetary Interiors*, **3**, 247-254.
- GREEN, D. H. and HIBBERSON, W., 1970. Instability of plagioclase in peridotite at high pressures. *Lithos*, **3**, 209-221.
- GREEN, D. H. and RINGWOOD, A. E., 1967a. The stability fields of aluminous pyroxene peridotite and garnet peridotite and their relevance in upper mantle structure. *Earth Planetary Sci. Lett.*, **3**, 151-160.

- GREEN, D. H. and RINGWOOD, A. E., 1967b. The genesis of basaltic magmas. *Contr.Mineral.Petrol.*, **15**, 103 - 190.
- GREEN, D. H. and RINGWOOD, A. E., 1967c. An experimental study of the gabbro to eclogite transformation and its petrological applications. *Geochim. Cosmochim.Acta.*, **31**, 767 - 833.
- GREEN, D. H. and RINGWOOD, A. E., 1969. Mineralogy of peridotitic compositions under upper mantle conditions. *Phys.EarthPlanetaryInteriors*, **3**, 359 - 371.
- JACKSON, E. D. and WRIGHT, T. L., 1970. Xenoliths in the Honolulu Volcanic Series, Hawaii. *J.Petrol.*, **11**, 405 - 430.
- KUNO, H., 1960. High-alumina basalt. *J.Petrol.*, **1**, 121 - 145.
- KUSHIRO, I., 1969. Discussion of paper "The origin of basaltic and nephelinitic magmas in the earth's mantle". *Tectonophysics*, **7**, 427 - 436.
- KUSHIRO, I., YODER, H. S. and NISHIKAWA, M., 1968. Effect of water on the melting of enstatite. *Geol.Soc.Amer.Bull.*, **79**, 1,685 - 1,692.
- MACDONALD, G. A. and KATSURA, T., 1961. Variations in the lava of the 1959 eruption in Kilauea Iki. *Pacific Sci.*, **15**, 358 - 369.
- MURATA, K. J. and RICHTER, D. H., 1966. Chemistry of the lavas of the 1959 - 60 eruption of Kilauea volcano, Hawaii. *U.S.Geol.Surv.Profess.Papers*, **537-A**, 1 - 26.
- NOCKOLDS, S. R., 1954. Average chemical compositions of some igneous rocks. *Geol.Soc.Amer.Bull.*, **65**, 1,007 - 1,032.
- O'HARA, M. J., 1968. The bearing of phase equilibria studies in synthetic and natural systems on the origin and evolution of basic and ultrabasic rocks. *Earth Sci.Rev.* **4**, 69 - 133.
- O'HARA, M. J. and YODER, H. S., 1964. Formation and fractionation of basic magmas at high pressures. *Scottish J.Geol.*, **3**, 67 - 117.
- RINGWOOD, A. E., 1962. A model for the upper mantle. *J.Geophys.Res.*, **67**, 857 - 867.
- RINGWOOD, A. E., 1966a. The chemical composition and origin of the earth. p. 287 - 356 in *Advances in Earth Science*, ed. P. M. Hurley. M.I.T. Press, Cambridge, Mass.
- RINGWOOD, A. E., 1966b. The mineralogy of the mantle. p. 357 - 417 in *Advances in Earth Science*, ed. P. M. Hurley. M.I.T. Press, Cambridge, Mass.
- TILLEY, C. E., 1950. Some aspects of magmatic evolution. *Quart.J.Geol.Soc., London*, **106**, 37 - 61.

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