Composition of Basaltic Magmas as Indicators of Conditions of Origin: Application to Oceanic Volcanism [and Discussion]

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V. PETROGENESIS AND DISCUSSION

Composition of basaltic magmas as indicators of conditions of origin: application to oceanic volcanism

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Basaltic magmas are formed by partial melting of a source rock of peridotitic composition (pyrolite) under upper mantle conditions. Experimental studies of the mineralogy of pyrolite and the melting relations of various basaltic magmas under high-pressure conditions are integrated in an attempt to present an internally consistent model of source composition, derived liquid compositions and residual mantle compositions. The role of a small (0.1 %) content of water in the upper mantle is treated in some detail. The presence of the low velocity zone in the upper mantle is attributed to a small ($< 5\%$) degree of melting of pyrolite containing approximately 0.1 % water. The small liquid fraction present in the low-velocity zone is highly undersaturated olivine nephelinite or olivine melilite nephelinite. Other magma types of direct upper mantle derivation ranging from olivine trachybasalt to olivine melilitite and to tholeiitic picrite are assigned to a genetic grid expressing the depth (pressure) of magma segregation, the degree of partial melting of the source pyrolite, the water content and approximate temperature of the magma. While this genetic model can account for variations in major element abundances and normative mineralogy among basalts, there are variations in abundances of the incompatible elements, particularly K, Rb, Ba, and the rare earths, which are inconsistent with a model invoking a constant source composition for all mantle-derived basalts. Additional factors producing source inhomogeneity, particularly in incompatible element abundances, include the possibility of two-stage melting and of chemical zoning within the low-velocity zone. It is suggested that vertical migration of a fluid or incipient melt phase, enriched in H_2O, CO_2 and incompatible elements, occurs within the low-velocity zone.

The evolution of continental and oceanic rift systems and of the Hawaiian volcanic province is discussed in relation to the depths and conditions of magma genesis derived from the models of magma genesis.

INTRODUCTION

Basaltic magmas are derived by partial melting of source rock of peridotitic composition and knowledge of possible magma and residue relations allows estimation of source compositions from observed natural basalts and peridotites. This is the rationale of the 'pyrolite' model composition for the Earth's upper mantle and it is emphasized that the method is a general one capable of revision and refinement as knowledge of magma/residue pairs is extended. At the present stage the model pyrolite composition of Ringwood (1966a) provides a suitable framework and reference composition to employ in the search for internally consistent models of primary mantle composition, derived liquids, refractory residues and magmatic accumulates.

The pyrolite composition crystallizes in four anhydrous mineral assemblages under near-solidus upper mantle conditions (Green & Ringwood 1967a, 1969). At temperatures near the anhydrous solidus, these four assemblages are:

(i) olivine + orthopyroxene + clinopyroxene + plagioclase + minor chromite
   $[P \rightarrow 9 \text{ kbar } (0.9 \text{ GN m}^{-2})]$;
(ii) olivine + orthopyroxene + clinopyroxene + minor spinel + minor plagioclase (pyroxenes moderately aluminous with 3–4 % Al_2O_3) $[P = 9\text{–}12 \text{ kbar } (0.9\text{–}1.2 \text{ GN m}^{-2})]$.
(iii) olivine + aluminous orthopyroxene + aluminous clinopyroxene (pyroxenes with 6–8 % Al_2O_3) $[P = 12\text{–}30 \text{ kbar } (1.2\text{–}3.0 \text{ GN m}^{-2})]$.

(iv) olivine + orthopyroxene + clinopyroxene + garnet [$P > 30$ kbar (3 GN m^{-2})].

In melting anhydrous pyrolite, the first drop of liquid formed at a given pressure will be in equilibrium with one of the mineral assemblages listed above and with increasing degree of melting, the residual phases will change in composition and be eliminated one by one, some possibly melting incongruently, e.g.

$$\text{clinopyroxene} + \text{liquid}_1 \rightarrow \text{olivine} \pm \text{orthopyroxene} + \text{liquid}_2.$$
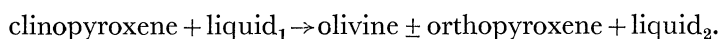
The composition of the partial melt will thus change with the degree of melting as the solid phases buffering it also change. Because of the olivine-rich nature of the source composition, all partial melts will be saturated with respect to olivine. Experimental studies on melting of pyrolite (Green & Ringwood 1967 a), on picritic and basaltic compositions, and consideration of the abundance of enstatite in the modal pyrolite mineralogy at pressures of 10 to 30 kbar (1–3 GN m^{-2}), show that enstatite will be the second last phase to disappear during progressive melting at pressures of 5 to 50 kbar (0.5–4.0 GN m^{-2}). Thus self-consistency arguments imply that basalts derived directly by partial melting of pyrolite composition will have olivine and enstatite and may have plagioclase, clinopyroxene, spinel or garnet as liquidus phases at appropriate pressure and temperature conditions.

Recent geophysical studies have confirmed the existence of a low velocity zone in the upper mantle, beginning at depths of 80 to 120 km and extending to around 200 km (Archambeau, Flinn & Lambert 1969). It has also been argued that the characteristics of the low-velocity zone indicate that this is a region in which a small percentage of interstitial liquid is present in the mantle pyrolite. Estimates of the continental or oceanic geothermal gradients (Clark & Ringwood 1964) and knowledge of the anhydrous pyrolite solidus show that the geotherm would lie well below the anhydrous solidus at this depth interval (see figure 1). Either there is considerable error in the estimates of average geothermal gradients or some other factor depresses the solidus of the mantle to lower temperatures in this depth interval. Recent studies of the role of water under upper mantle conditions show that small (0.1 to 0.2 %) quantities of water in upper mantle pyrolite provide a solution to this dilemma. Experimental studies of olivine nephelinite magmas showed that derivation of these extremely undersaturated magmas from pyrolite is possible provided these magmas were hydrous (2 to 10 % H$_2$O) and formed by partial melting of a pyrolite source rock at temperatures 150 to 250 °C below the anhydrous pyrolite solidus (Bultitude & Green 1968, Green 1969 a, b). Studies of the stability of the hydrous mineral amphibole in basaltic and peridotitic compositions (Green & Ringwood 1967 c; Lambert & Wyllie 1968; Green & Ringwood 1969) showed that this mineral was unstable relative to the assemblage garnet + pyroxene ($\pm$ olivine) + water at pressures above 25 to 30 kbar (2.5–3.0 GN m^{-2}), $P_\text{T} = P_{\text{H}_2\text{O}}$, $T = 1000$ °C. In particular, it has been shown that, in pyrolite composition containing 0.1 to 0.2 % H$_2$O, the olivine + amphibole + pyroxene(s) assemblage is stable at 1000 °C at pressures less than 28 to 29 kbar (2.8–2.9 GN m^{-2}) but is replaced by olivine + pyroxenes + garnet + minor phlogopite + water at higher pressures (Green & Ringwood 1969). The depth limitation on amphibole stability in the upper mantle has a large effect on the character of the pyrolite solidus. This is illustrated in figure 1 which contrasts the solidus for anhydrous pyrolite with the solidus for pyrolite containing approximately 0.1 % H$_2$O. At low pressures amphibole is present at the solidus and the latter is determined by some value of $P_{\text{H}_2\text{O}} = P_1 < P_\text{total}$. At higher pressures this solidus for $P_{\text{H}_2\text{O}} = P_1$ intersects the breakdown

curve for amphibole at $P_{H_2O} = P_1$ and the solidus migrates through a series of such intersections to the intersection of solidus and amphibole breakdown for $P_{H_2O} = P_{total}$. If some other fluid phase such as CO_2 is present then P_{H_2O} may only attain some value

$$P_{H_2O} = P_2 < P_{fluid} = P_{total}$$

and the solidus would be appropriate for this value of P_{H_2O}.

It is apparent from figure 1 that a geotherm such as that shown, which would not intersect the anhydrous pyrolite solidus, will intersect the solidus for pyrolite containing 0.1 % H_2O at a depth of approximately 80 km. In figure 1, the dashed curves illustrate the degree of melting at
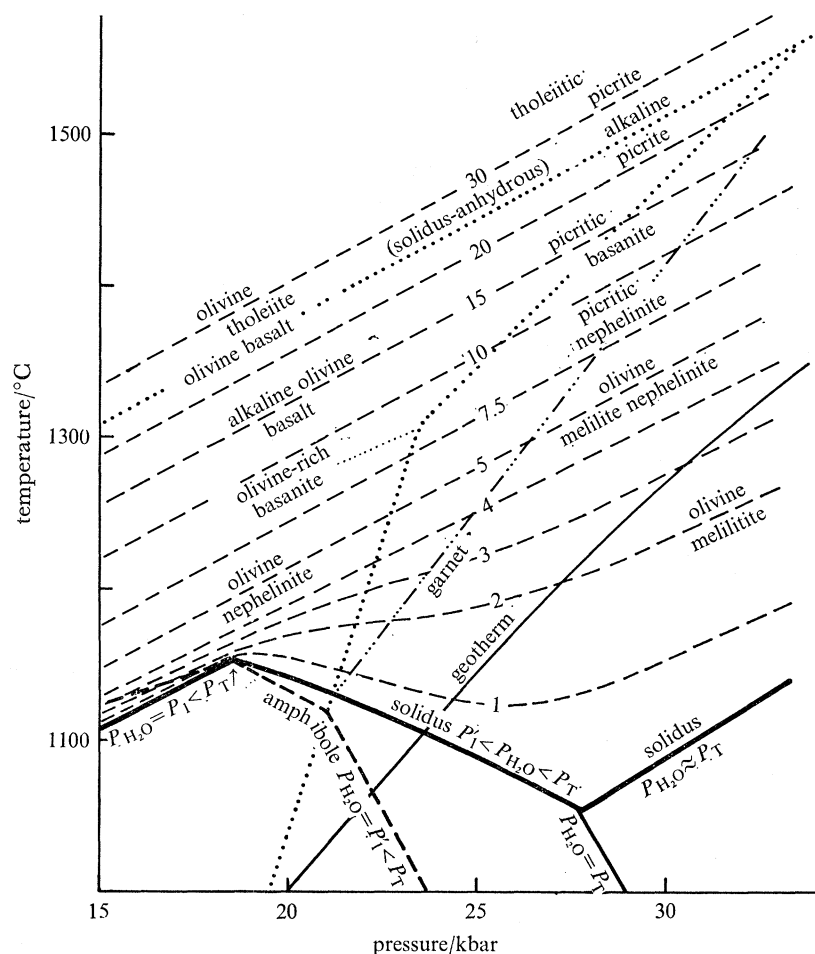

FIGURE 1. Diagram illustrating the amphibole stability limit at high pressure and the solidus for pyrolite composition containing approximately 0.1 % H_2O. Also shown (dotted lines) are the solidus for anhydrous pyrolite and the boundary for appearance of garnet (from Green & Ringwood 1967a). The numbered 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 to 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. P_{H_2O}, equilibrium water pressure; P_1, water pressure along solidus for pyrolite with 0.1 % H_2O.

P_1 varies along the solidus and an arrow indicates the intersection of the solidus with a subsolidus breakdown curve for amphibole at $P_{H_2O} = P_1'$ also (amphibole $\rightleftarrows$ pyroxenes + garnet + olivine + water).

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temperatures above the solidus and this is further illustrated and contrasted with the anhydrous case in figure 2. It is apparent that melting in a pyrolite source with a low and limited amount of water produces a condition where the amount of liquid formed remains small over quite large temperature and depth variations, i.e. small temperature perturbations in the low-velocity zone or small changes in depth within the zone do not produce extremely large variations in the percentage of melt and thus in the seismic properties. If the low-velocity zone is itself compositionally zoned (see later sections) so that the amount of water decreases with depth below 120 to 150 km then this will limit or prevent melting at deeper levels.
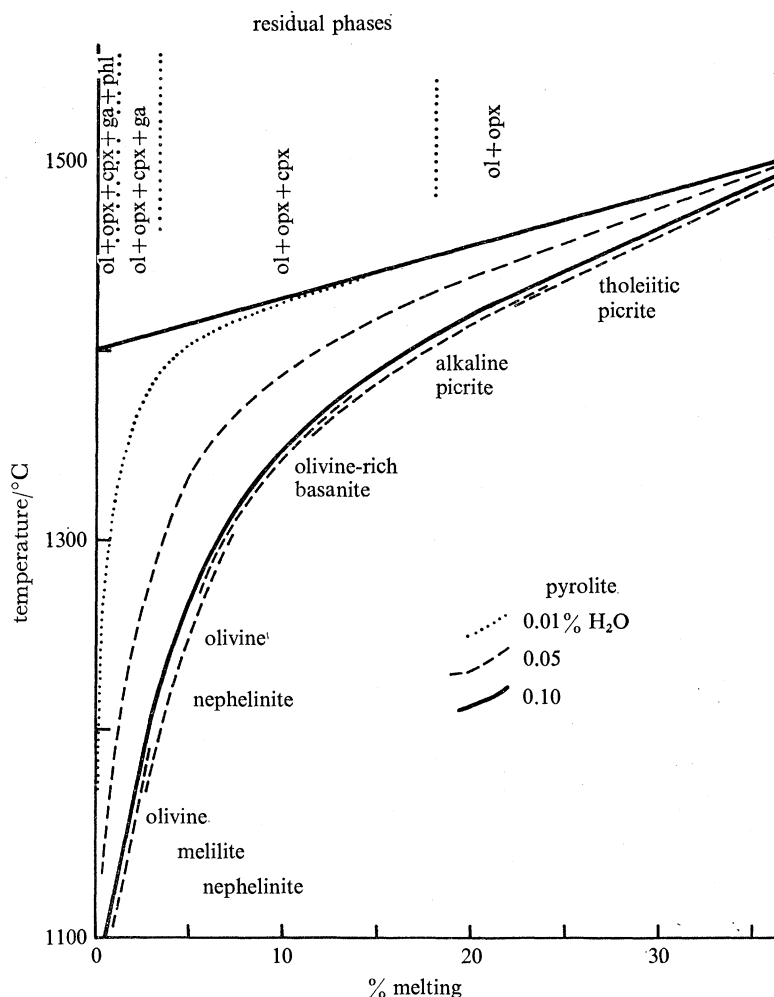

FIGURE 2. Diagram contrasting the melting behaviour of anhydrous pyrolite with melting in pyrolite compositions containing 0.01, 0.05 and 0.10 % H_2O. The nature of the residual phases and the petrochemical character of the liquids derived are shown for the example with 0.10 % H_2O.

The hypothesis that melting occurs in a hydrous pyrolite source has important implications in that pyrolite contains both orthopyroxene and clinopyroxene and the mutual solid solution between these phases is very temperature dependent. Similarly, the Al_2O_3 solubility in pyroxenes is both temperature and pressure dependent. Thus a liquid produced with a very small degree of melting of anhydrous pyrolite at 1400 °C, 25 kbar (2.5 GN m^{-2}) would be in equilibrium with olivine, aluminous enstatite ($\sim$ 6 % Al_2O_3, 2.2 % CaO) and aluminous

subcalcic clinopyroxene ($\sim 8 \%$ Al_2O_3, 11% CaO), whereas the liquid produced with a similarly small degree of melting at 25 kbar, 1150 °C (hydrous pyrolite source) would be in equilibrium with olivine, enstatite ($\sim 3 \%$ Al_2O_3, $< 1.5 \%$ CaO) clinopyroxene ($\sim 5 \%$ Al_2O_3, $> 20 \%$ CaO) and garnet (or spinel) (Green & Ringwood 1967a, 1969).

The discussion in this section has centred on the choice of a model composition for the upper mantle acting as source for basaltic magmas and on illustrating the very sensitive controls exerted by P, T and P_{H_2O} in determining the mineralogy and solidus temperature of this composition. In the following section, the liquidus phases of basalts at high pressures are briefly reviewed in the context of self-consistency with derivation from the pyrolite model.

LIQUIDUS MINERALS OF BASALTIC COMPOSITIONS AT HIGH PRESSURES

In seeking to establish models in which basaltic magmas are derived by partial melting of a peridotite source rock, a principal constraint, imposed by self-consistency, is that the liquidus phases of the postulated melt must be the same as the residual phases of the postulated source rock at the appropriate P, T, P_{H_2O} conditions. Experimental study of basaltic compositions at high pressures, both dry, and with various water contents, serves to identify the nature of the liquidus phases which can then be compared with the data on solidus and near-solidus mineralogy of pyrolite to determine whether the basalt studied is a possible derivative liquid from the pyrolite source at some P, T, P_{H_2O}. It must be emphasized that the feasibility of using this method depends entirely on obtaining accurate chemical compositions of the mineral phases—comparison must include degree of pyroxene solid solution, Al_2O_3 content of pyroxenes and, most importantly, the $Mg/Mg + Fe$ ratios of the mineral phases. The combined use of experimental high-pressure crystallization techniques utilizing complex natural basalts and electron microprobe analyses of the mineral phases synthesized is essential in this type of study.

Two examples of basalt suites studied experimentally are briefly summarized below. Studies of olivine tholeiite and high alumina olivine tholeiite (Green & Ringwood 1964, 1967b; Green, Green & Ringwood 1967; Green 1969b; Tilley & Yoder 1964) under *dry* conditions show that these liquids may have magnesian olivine and aluminous orthopyroxene occurring together on the liquidus at a specific pressure. At 9 kbar (0.9 GN m^{-2}), the liquid with olivine, enstatite and clinopyroxene together on the liquidus has about 10 to 15 % normative olivine, 10 to 15 % normative hypersthene and has high Al_2O_3 content (15 to 16 %). Liquids with olivine and enstatite only on the liquidus at 9 kbar have higher olivine and hypersthene contents and lower Al_2O_3 contents (Green & Ringwood 1967b, Green et al. 1967). At 12 kbar (1.2 GN m^{-2}), a liquid with olivine and enstatite together on the liquidus contains 20 % normative olivine, 12 to 14 % hypersthene and lower Al_2O_3 content (13 to 14 %) (Green & Ringwood 1967b). At 15 to 16 kbar (1.5 to 1.6 GN m^{-2}), a liquid with olivine and enstatite together on the liquidus has 26 % normative olivine, 12 % hypersthene, and 12 to 13 % Al_2O_3. At 18 kbar (1.8 GN m^{-2}), the tholeiitic liquid would contain 28 % olivine, 12 % hypersthene and 11 to 12 % Al_2O_3 (Green 1969b). These experiments illustrate the variation in the nature of tholeiitic liquids which can be produced at various depths from the pyrolite source rock, provided the degree of melting is sufficient to eliminate all the plagioclase, and most of the clinopyroxene from the source pyrolite at 9 kbar, all the spinel and clinopyroxene at 12 kbar and all the clinopyroxene together with appreciable olivine and aluminous enstatite at 15 to 18 kbar.

Experiments on the effect of water on the liquidus phases of undersaturated basalts (Bultitude & Green 1968, Green 1969a, b) show that whereas olivine and clinopyroxene or clinopyroxene alone may be the liquidus phases under dry conditions (liquidus 1400 to 1450 °C, 20 to 30 kbar (2 to 3 GN m^{-2}), if the liquidus is depressed to 1200 to 1300 °C by the addition of 2 to 5 % water, then olivine and orthopyroxene, or olivine, orthopyroxene and clinopyroxene may occur at or near the liquidus. This effect has been substantiated with experiments using sealed capsules and known quantities of water (Green 1969a, b; Green & Hibberson 1969). It should be noted that the near-liquidus clinopyroxene under dry conditions at $\sim$ 1400 to 1450 °C is rather subcalcic and with the rapid widening of the pyroxene immiscibility gap at lower temperature, such a composition would be represented by orthopyroxene and clinopyroxene. Thus the 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 experiments show that very undersaturated basanites and olivine nephelinites may form in equilibrium with residual magnesian olivine, orthopyroxene and calcic clinopyroxene at pressures of 18 to 35 kbar (1.8 to 3.5 GN m^{-2}) provided the solidus of the source rock is depressed 100 to 250 °C below the anhydrous solidus, with consequent changes in source-rock mineralogy and in the nature of liquids formed at or near the solidus. It may be noted that the temperature, pressure régime under discussion, i.e. $T \sim$ 1200 °C, $P \geqslant$ 18 kbar (1.8 GN m^{-2}) is largely outside the stability field of amphibole so that this phase does not occur at the solidus, particularly at pressures > 25 kbar (2.5 GN m^{-2}). Phlogopite remains stable, and may be a minor but very significant phase in the subsolidus mineralogy.

SUMMARY OF BASALT MAGMA AND PYROLITE RELATIONS

Figures 1 to 3 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 those magmas of direct mantle derivation as evidenced by the presence within them of xenoliths or xenocrysts of high-pressure mineralogy or in the case of tholeiitic magmas, as deduced from geophysical data. Processes of crystal fractionation at crustal levels may modify these magma types producing other magmas classifiable as basanite, trachybasalt or hawaiite but differing chemically and mineralogically from the types represented in figures 1 to 3.

In figures 2 and 3, tholeiitic magmas are produced with high degrees of partial melting ($\sim$ 30 %) and only olivine and enstatite remain in the residue. The water contents of such magmas, if developed from a source rock with 0.1 % H$_2$O, would be only 0.3 to 0.4 % H$_2$O. It is to be noted that such magmas would form only 30 to 50 °C below the temperature at which tholeiitic magmas would form from anhydrous pyrolite and little compositional difference would be anticipated between tholeiites developed from an anhydrous pyrolite and those from a hydrous pyrolite source.

The curve for the appearance of garnet in the anhydrous pyrolite composition is shown in figure 2 and this intersects the anhydrous solidus at > 30 kbar (3 GN m^{-2}). If the source rock contains 0.1 % H$_2$O then the lower part of this curve lies within the amphibole stability field and garnet may not appear until amphibole disappears at the solidus with a reaction possibly represented by amphibole $\rightleftharpoons$ olivine + garnet + pyroxene + liquid.

The curve for the appearance of garnet *above* the solidus for 'wet' pyrolite is shown schematically. Contours in figure 2 illustrate the variation in degree of melting at temperatures above the solidus and the petrologic character of the liquids for various degrees of melting is also shown. Since the source is regarded as having a fixed water content and all of this is partitioned into the liquid at the solidus (except below $\sim$ 18 kbar (1.8 GN m^{-2}) where amphibole has a minor role immediately above the solidus) the water content of the various magma types can also be read from figure 2.
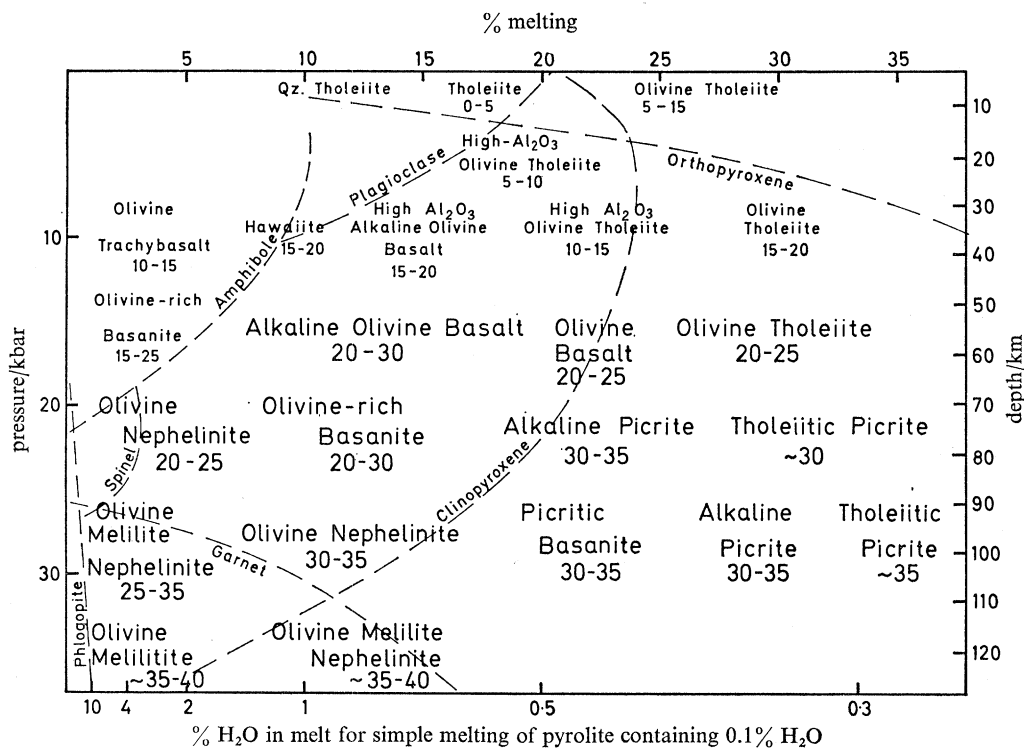

FIGURE 3. A petrogenetic grid for mantle-derived basaltic magmas. Various basalt magma types are assigned to a % melt, pressure grid (implying also specific % H_2O and temperature of melting, see figures 1 and 2) in which they are regarded as partial melting products of a pyrolite composition containing 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 of this figure.

In figure 3, the arguments of figures 1 and 2 are synthesized to present a pressure (or depth) against % melting grid for a source pyrolite with 0.1 % H_2O, in which characteristic natural magma types are assigned to a pigeon-hole specifying their depth of magma segregation, the degree of partial melting (with implication of temperature as in figures 1 and 2) their water content and the nature of mineral phases with which they were in equilibrium at the point of magma segregation and which will remain in the more refractory peridotitic residue. The magma types appearing on figure 3 include only these liquids developed in equilibrium with residual peridotite—these magmas may in turn be modified by crystal fractionation as they move from the depth of magma segregation to eruption at the Earth's surface.

The conclusions on basalt petrogenesis summarized in figures 1 to 3 will undoubtedly require modification and refinement as further detailed studies proceed—the olivine trachybasalt to high alumina olivine basalt series and the olivine nephelinite to alkaline picrite series are currently under more intensive study. However, the relation outlined in figures 1 to 3 are considered sufficiently well founded to justify an attempt to use this petrogenetic grid to make deductions about the nature of melting processes responsible for rift valley, oceanic ridge and oceanic island volcanism.

MINOR AND TRACE ELEMENTS IN BASALTS—SOURCE INHOMOGENEITY

The relations expressed in figures 1 to 3 are considered to account for differences between basalt types in major element chemical composition, normative and modal mineralogy and in ratios such as $Mg/(Mg+Fe^{2+})$, $Na/(Na+Ca)$. However, there is a group of elements, referred to as incompatible elements by Green & Ringwood (1967b) and including K, Ti, P, Ba, Sr, Rb, Zr, Hf, U, Th, Pb and the lighter rare earth elements, which show variations in basalts that cannot be accounted for by the hypothesis of a constant source (pyrolite) composition and single stage melting of the character illustrated in figure 3. These elements are those which would be so strongly partitioned into the basaltic liquid that for partial melting in excess of about 5 % their concentration should closely reflect the % melting and source composition only. Their concentration would be largely independent of relative proportions of olivine, orthopyroxene and clinopyroxene in the residual phases. On this basis and from figure 3, high-alumina olivine tholeiites such as occur commonly in the mid-oceanic ridge environment, should have K contents intermediate between those of alkali olivine basalts and parental low-alumina tholeiites. Comparison of K contents of Hawaiian alkali olivine basalts and Hawaiian low-alumina olivine tholeiites shows that these magma types are roughly consistent with magma genesis at 50 to 60 km according to the relations of figure 3. However, the same source rock could not produce the very low-potassium tholeiites (of both low-alumina and high-alumina type) with the similar degrees of melting implied by figure 3. Differences in rare-earth element fractionation patterns between Hawaiian tholeiitic basalts and low-potassium tholeiitic rocks from the oceanic ridges, and between low-potassium and high-potassium basalts from oceanic ridges, further illustrate the dilemma (Green & Ringwood 1967b; Gast 1968; Green 1969b; Kay, Hubbard & Gast 1970). Green & Ringwood (1967b) called on processes of wall rock reaction to account for selective enrichment of the incompatible elements in the more under-saturated rocks but this will not account for differences between various olivine tholeiitic magmas. Gast (1968) clearly delineated the problem and, considering the trace element data of paramount importance, interpreted alkaline magmas such as the Hawaiian alkali magma series as products of very small degrees (3 to 7 %) of partial melting of the peridotite source. The low-potassium olivine tholeiites of the mid-oceanic ridges were interpreted as products of 20 to 30 % melting. Olivine tholeiites of Hawaiian type, which have higher K_2O contents and r.e. abundances showing enrichment in the light r.e., were grouped with the alkaline basalts on the basis of their trace element contents. The inference that parental Hawaiian olivine tholeiite is produced by very small degrees of partial melting is inconsistent with the phase equilibrium studies on the roles of olivine, enstatite and clinopyroxene as liquidus phases and mantle residual phases for these different magma types. Gast (1968) also advocated a two-stage melting process. The source rock for the low potassium olivine tholeiites was regarded (p. 1077) as having

previously been depleted, particularly in incompatible elements, by an earlier partial melting event in which a very small magmatic fraction was extracted.

Further geochemical data on oceanic basalts, particularly the studies of r.e. element distributions by Kay *et al.* (1970), Frey, Haskin, Poetz & Haskin (1968) and Schilling (1969), lends support to the concept that differences in incompatible element contents reflect source characteristics. Muir & Tilley (1964) and Nicholls (1964) both describe high-alumina olivine basalts transitional between olivine tholeiite and high-alumina alkali olivine basalt. Both papers describe dredged samples from the crestal region of the oceanic ridge but those described by Muir & Tilley have high K_2O contents (0.4 to 0.6 % K_2O), whereas that described by Nicholls has $K_2O = 0.09$ %. The abundance of r.e. elements in at least one of Muir & Tilley's samples is enriched in lighter rare earths (Frey *et al.* 1968; Kay *et al.* 1969), whereas Nicholls's sample would be expected to show the depletion of light r.e. elements characteristic of other low-potassium oceanic basalts (Kay *et al.* 1970).

If basalts with marked differences in incompatible element contents, but with similar major element composition were randomly distributed in space and time then one might conceive of a mantle source region which is irregularly and randomly inhomogeneous due to effects of variable degrees of magma extraction at various times in the past. However, several authors, notably Gast (1968), Schilling (1969) and Kay *et al.* (1970) have argued that the regular appearance of the low-potassium tholeiite at mid-oceanic ridge crests suggests that this magmatic process taps a widespread and relatively homogeneous source region of the mantle. Basalts with high K_2O contents and high contents of other incompatible elements relative to the low-potassium tholeiites do occur at abyssal depths in the ocean basins. Commonly they occur on the flanks of the ridges rather than within the median valley (Aumento 1967, 1968; Kay *et al.* 1970) and rock types range fom high-alumina alkali olivine basalts to olivine basanite (Melson, Jarosewich, Cifelli & Thompson 1967). The occurrence of high-potassium basalts is particularly characteristic of seamounts which locally are symmetrically developed on either side of the median valley (Aumento 1967, 1968) or from chains moving away from the ridge crest in the direction of oceanic plate movement (Menard 1969). These occurrences show that an oceanic crustal locality, which taps a source of low-potassium tholeiite when at the ridge crest, may later tap a second source region as it moves away from the ridge crest. The second source region characteristically provides magmas relatively enriched in incompatible elements.

It appears that there are two independent controls on the chemical characteristics of major oceanic basalt magma types. These are the conditions (P, T, P_{H_2O}) of melting and the nature of the source composition. The source composition is envisaged as relatively constant in major element content and thus in mineralogy but variable in content of incompatible elements and accessory phases such as phlogopite or apatite. It is suggested that the upper mantle is vertically zoned in its content of incompatible elements and that this chemical zoning results from the upward migration within the low-velocity zone of a separate fluid phase. This is envisaged as water-rich, possibly CO_2-rich and may be a very small water-rich silicate melt fraction. This concept is illustrated in figures 4 and 6. The movement of incompatible elements within or through a fluid phase has been discussed by Green (1969b) and Frey (1969) gives an excellent illustration of the operation of this process as regards r.e. elements. Frey shows that selective depletion of light rare earths occurs in a dehydration reaction (amphibole → pyroxene) in a metamorphic aureole and that enrichment in light rare earths occurs in neighbouring rocks (highly magnesian peridotite) in which amphibole is stable and crystallizes under the ambient

conditions at the expense of magnesian pyroxenes and olivine. A similar effect on a large scale and over long periods of time is envisaged within the low-velocity zone. The presence of a free fluid phase ($H_2O \pm CO_2$) in this region permits the upward migration of incompatible elements which are strongly rejected by the major minerals in the low-velocity zone (olivine, aluminous pyroxenes, garnet) until they are 'fixed' by entering a small silicate melt fraction in the upper part of the low-velocity zone or possibly by entering minor hydrous phases (particularly amphibole) at the base of the lithosphere. Referring to the r.e. elements, the presence of garnet in the low-velocity zone and its increasing abundance at deeper levels (Green & Ringwood 1967a) would strongly influence the r.e. distribution pattern as the heavy rare earths would be strongly partitioned into the garnet (Gast 1968), whereas the light rare earths would strongly partition into the fluid phase relative to garnet, clinopyroxene or enstatite (Frey 1969). In this way contrasted r.e. distribution patterns would be developed within the low-velocity zone. To summarize, it is suggested that the upper part of the low-velocity zone has relative abundances of incompatible elements matching those of olivine-rich tholeiite of Hawaiian type (e.g. average Kilauea Iki lava lake), whereas the lower part of the low-velocity zone has the relative abundances of incompatible elements of an oceanic ridge low-potassium olivine tholeiite. In both regions, the actual concentrations of the incompatible elements in the source rocks, are one-fifth to one-third those of the characteristic tholeiitic magmas.

The mantle lithosphere overlying the low-velocity zone is considered to be inhomogeneous on a smaller, local scale reflecting diapiric emplacement of mantle from the low-velocity zone, local magma segregation and magma channels, and local accumulates from basaltic

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LEGEND TO FIGURE 4

FIGURE 4. Illustration of possible processes relating rift valley volcanism to continental rifting and the creation of new oceanic crust with a typical mid-oceanic rift system. The mantle lithosphere is in part refractory, residual peridotite in which minor amphibole may occur if water is available. The low-velocity zone is of pyrolite composition containing, in the upper part at least, about 0.10 % H_2O. Amphibole is unstable in the low-velocity zone and this results in a region of partial melting with < 5 % of highly undersaturated and volatile-enriched magma in the upper part of the low-velocity zone. Vertical migration of a H_2O-rich fluid phase is considered to cause chemical zoning within the low-velocity zone, the lower part becoming depleted, and the upper part enriched, in incompatible elements.

Stage I. Local tapping of magmas present and potentially available within the low-velocity zone—parental magma types are olivine nephelinites and olivine melilitites, characteristic of small degrees of partial melting at depths of 80 to 120 km. A zone of partially depleted pyrolite is produced within the low-velocity zone.

Stage II. With appreciable crustal thinning under tension, the upwelling of material from the low-velocity zone becomes important leading to both lateral and vertical movements within the low velocity zone. Volcanism of the characteristic stage I type may continue with the depth of magma segregation now near 60 to 70 km (olivine-rich basanite) but also with the diapiric upward movement of material from the low-velocity zone (Green & Ringwood 1967b, figure 12) magmas produced by > 5 % partial melting and with segregation at ≤ 60 km may appear (alkali olivine basalts, olivine basalts, olivine tholeiites). The source of the pyrolite diapirs is considered to be the upper part of the low-velocity zone.

Stage III. With rifting apart of the continental crust and mantle lithosphere, pyrolite upwelling to shallow depths (∼ 30 km or less) will occur. This may bring partially depleted pyrolite (the source region for stage I magmas) to depths of ∼ 30 km causing a second melting stage—20 to 25 % melt will yield high-alumina olivine tholeiites and the two-stage melting will produce very low contents of incompatible elements in these magmas. Continuation of the rifting process will lead to upwelling of pyrolite from deeper within the low velocity zone and having the incompatible element contents typical of this zone. It is conjectured that the upper part of the low velocity zone moves with the overlying plate, the rate of such movement decreasing with depth and reversing at some depth to flow towards the ocean-floor spreading centre. Tectonic activity within the lithosphere plates on either side of the spreading centre may tap magmas from the low-velocity zone yielding volumetrically small extrusions of extremely undersaturated magma types—such centres would move with the lithosphere plate.
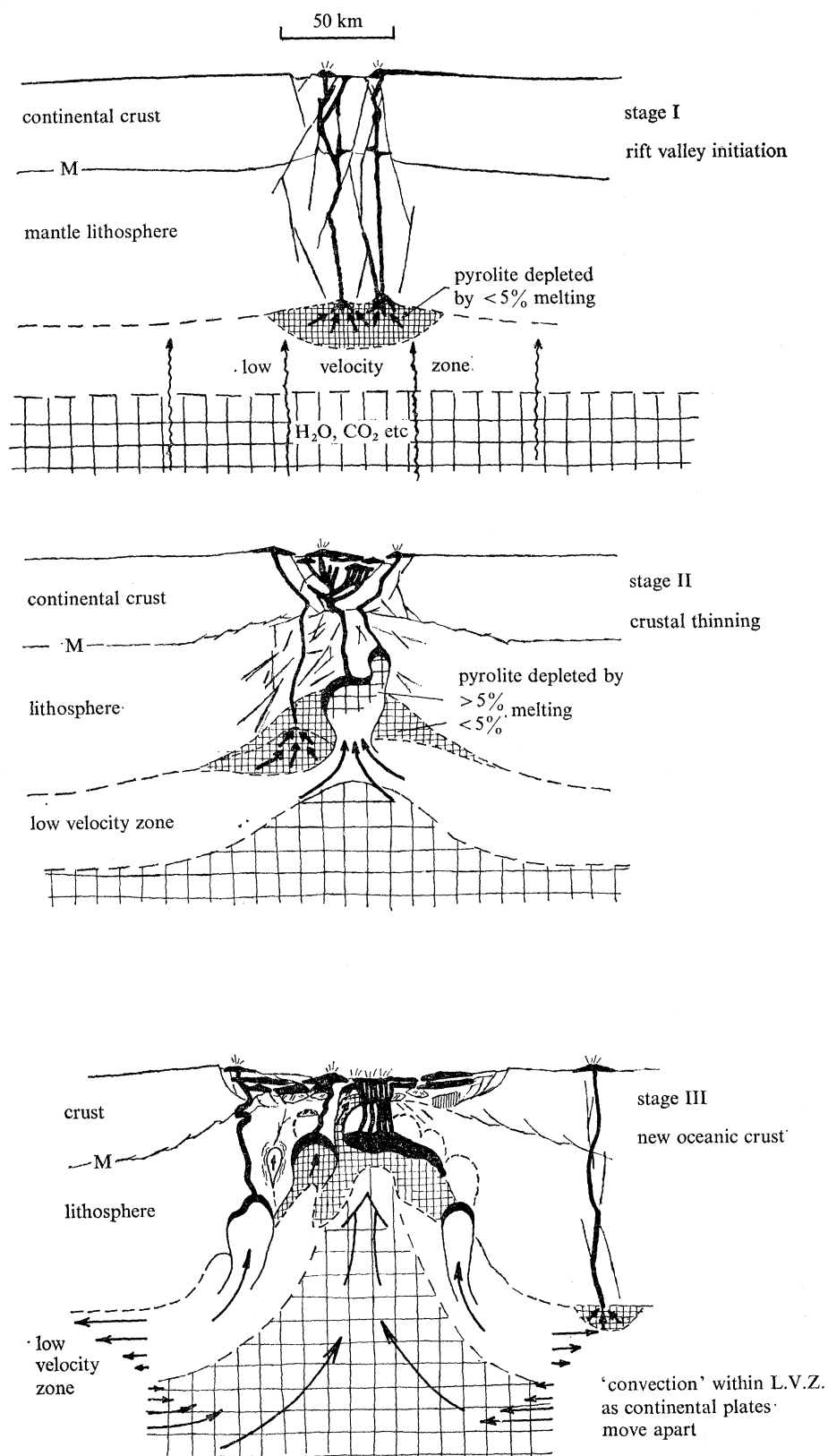
FIGURE 4. For legend see facing page.

magmatism. It is considered to range from refractory harzburgite to near-primitive pyrolite. This is the region sampled by explosive eruption of hydrous magmas from the low velocity zone and the detailed geochemistry of lherzolite inclusions is currently revealing the variation and complex history of this mantle region (Leggo & Hutchison 1968; Kleeman, Green & Lovering 1969; Cooper & Green 1969).

recent tholeiitic centres
recent alkali olivine basalt→trachyte centres
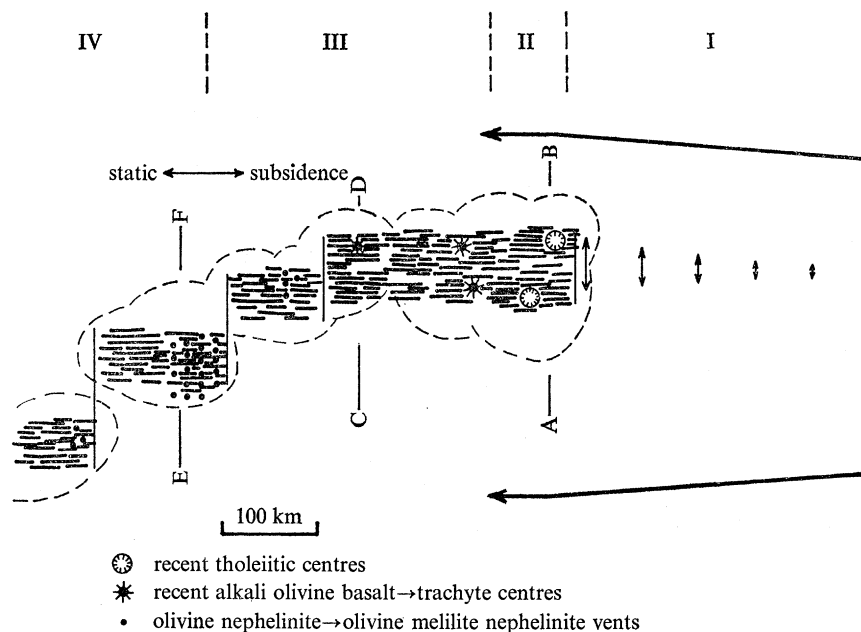olivine nephelinite→olivine melilite nephelinite vents

FIGURE 5. Hypothesis outlining the origin of a chain of volcanic islands on a moving oceanic plate. The diagram represents an idealized plan of the Hawaiian Islands. It is suggested that divergence of flow vectors, such as could be produced by a plate moving over an imperfect sphere, leads to a zone of tension. Intermittent failure of the lithosphere leads to rapid upwelling from the low velocity zone, a short-lived stage of active volcanism with creation of new oceanic crust, subjacent dyke swarms (illustrated diagrammatically) and a broad carapace of extruded lavas.

Stage I. Zone of stress build-up in lithosphere due to divergent plate movement vectors.

Stage II. Failure of lithosphere, stress release. Active volcanism with formation of new crust and growth of individual tholeiitic dome in 1 to 2 Ma.

Stage III. Decay of volcanism with source now emplaced within and moving with the oceanic plate; consolidation of new crust and lithosphere.

Stage IV. Local rejuvenation of volcanism with transverse tensional faulting tapping magmas directly from the low-velocity zone. This has been correlated with a hinge zone between a region of crustal loading and subsidence and a static zone (Jackson & Wright 1970).

Diagrammatic sections illustrating the volcanism are given in figure 6. AB represents a section approximately through Kilauea and Mauna Loa, Hawaii. CD represents a section through East Maui. EF represents a section through Oahu.

APPLICATION TO RIFT VALLEY AND OCEANIC VOLCANISM

In previous sections detailed models of basalt petrogenesis and of upper mantle chemical variation have been presented. These models are now applied to specific natural volcanic provinces in figures 4 to 6. These figures and their detailed legends are largely self-explanatory. Although figure 4 differs in some respects, it owes much to models of oceanic ridge volcanism developed by Aumento (1967), Oxburgh & Turcotte (1968), Gast (1968), Cann (1968), Vogt, Schneider & Johnson (1969), Ringwood (1969) and Gass (1970). It is drawn with the African rift system, the Afar depression and the Red Sea and Gulf of Aden areas as natural parallels to
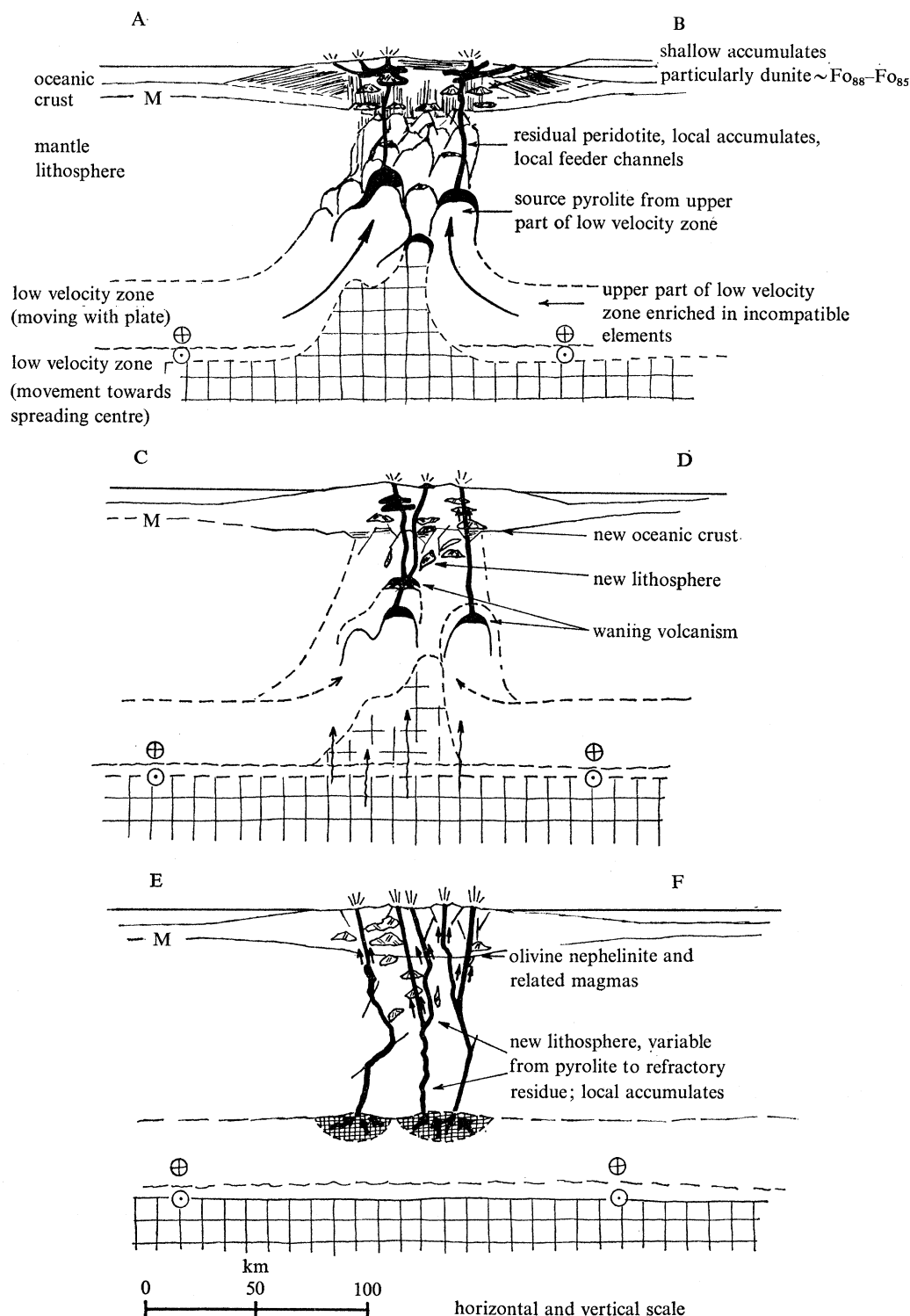
FIGURE 6. Evolutionary stages of oceanic volcanic islands with particular reference to the Hawaiian Islands. See figure 5 for location of sections AB, CD and EF.

AB. Stage of rapid cone building, development of new oceanic crust and new lithosphere from very rapid upwelling of pyrolite from the upper part of the low-velocity zone. Magmas of olivine tholeiite type formed by around 30 % melting at a depth of magma segregation of 40 to 60 km. In the later stages of cone formation, residence of magma batches within the volcanic pile becomes common with consequent precipitation of low pressure accumulates, particularly dunite with olivine of composition Fo_{88}–Fo_{85}.

CD. Stage of waning volcanism. The re-establishment of the lithosphere/low-velocity zone boundary and movement of the newly created lithosphere with the overlying plate. Magmas produced from smaller degrees of partial melting (< 20%) at depths of magma segregation of 40 to 70 km or from high-pressure crystal fractionation of magma batches relict from tholeiitic volcanism. Magmas of alkali olivine basalt series.

EF. Faulting rejuvenates local minor volcanism. Magma tapped directly from upper part of the low-velocity zone (cf. figure 4, stage I) and is volatile-enriched, olivine-rich basanite, olivine nephelinite or olivine melilite nephelinite. Explosive eruption from various depths entrains xenoliths of lower crust and upper mantle rock types.

the three stages of the model. In stage III, the model illustrates an example of two-stage melting in which a mantle volume which earlier yielded a small olivine nephelinite melt fraction, later undergoes a second melting at much shallower levels to produce low-potassium olivine tholeiites. While this can account for the magma types appearing early in the rifting process, it is difficult to envisage a 'steady state' process as the oceanic rift widens in which the source material reaching the rift area always undergoes a preliminary low degree of partial melting as it approaches the crest. This difficulty has influenced the choice of a model in which the lower part of the low-velocity zone has the low-potassium characteristics as a general feature. Although the movement towards the spreading centre is shown as localized within the lower part of the low-velocity zone, this movement could be distributed through a much greater depth interval and thus be substantially vertical at the ridge areas. The model does however require the establishment of a source region enriched in incompatible elements, probably at the top of the low-velocity zone, very early in the evolution of the oceanic crust–lithosphere system. This is attributed to the 'degassing' mechanism within the low-velocity zone previously discussed. The figure 4, stage III diagram, illustrates an example of a peridotite diapir reaching crustal levels (< 10 km) and differentiating in place to a layered complex of gabbroic accumulates and harzburgite residue. Also, diapiric ascent of a small, detached, high-temperature peridotite along a line of structural weakness is illustrated. These aspects of emplacement of ultramafic complexes at oceanic ridges are discussed in more detail elsewhere (Green 1970).

Figures 5 and 6 illustrate a model for the development of an oceanic chain of volcanic islands due to migration of a tensional fracture, approximately parallel to the plate movement vector but probably offset by transform faults, through a moving oceanic lithosphere. The fracture of the lithosphere is considered to produce a very rapid upwelling from the low-velocity zone, resulting in high degree ($\sim 30\%$) of melting at 40 to 60 km depth and very rapid building of the main cone phase. As the creation of new crust and lithosphere is not a continuous process, the lower part of the low-velocity zone does not normally attain depths suitable for partial melting, although it may be noted that the low-potassium tholeiites of the Pololu Series, Kohala, may be derived from this deeper source material. As the volcanoes enter the waning phase with magmas of the alkali olivine basalt suite, the pyrolite source lies within the lithosphere and moves with it. From geochronological data the time between stages I and III for Kauai Island (McDougall 1964) is $\geqslant 4$ Ma. If the Pacific plate is moving at 5 cm a^{-1} this implies a displacement of the $\geqslant 200$ km during the complete life cycle of a Hawaiian volcano.

Conclusions

Experimental studies of the high-pressure behaviour of basalts and potential mantle source material have been used to derive a detailed model of magma evolution and upper mantle composition. It is considered that in this way petrological data on characteristics of different types of oceanic volcanism can be used to place important constraints on the nature of processes occurring in the upper mantle. An important concept developed in this paper is that of mantle source inhomogeneity with a specific suggestion that the upper part of the low-velocity zone is enriched in incompatible elements relative to deeper levels. The petrochemical character of basalt magmas reflects these source characteristics as well as the conditions of partial melting and later magma fractionation.

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DISCUSSION

DR M. J. O'HARA (*University of Edinburgh*). The term pyrolite is ill defined and frequently changing, both in terms of chemistry and mineralogy (O'Hara 1968 *a*, pp. 105–113). Believers should specify which version they credit at any moment, and remember that the 'pyrolite models' *assume* the composition of the residual mantle, the composition of the primary partial melt magma, and the proportion of that liquid produced, ensuring a circular debate on the three most pressing questions of upper mantle petrology. The assumptions are, moreover, un-satisfactory in detail and lead to a postulated mantle composition which is *critically* different from that based on observed ultrabasic inclusions in kimberlite.

The pyrolite model cannot explain the chemistry of some important magmas, nor the origin of garnet–harzburgite nodules in kimberlite, whereas they are explained by a model based on real garnet–lherzolite nodules in kimberlite which can also explain all other observations accounted for by the pyrolite model (O'Hara 1970). The pyrolite model may not be satisfactory for the suboceanic mantle except in so far as *any* ultrabasic material would be.

Dr D. H. Green's latest hypothesis of magma origins by variable partial melting of 'damp' peridotites is based upon the observation that the orthopyroxene primary phase volume extends far into critically undersaturated compositions at high-water vapour pressures, although restricted to hypersthene saturated compositions in the absence of water. I have doubted the adequacy of composition control in the original experiments on which this observation is based (O'Hara 1968 *a*, p. 103) and look forward to seeing the new data on which the hypothesis is now based. Meantime I am impressed by the results of wet melting experiments carried out in the systems forsterite–silica, forsterite–diopside–silica, forsterite–Ca Tschermak's molecule–silica, forsterite–nepheline–silica, forsterite–diopside–anorthite–silica, natural tholeiite and natural high alumina basalt, by such workers as Chinner, Kushiro, Tilley and Yoder at the Geophysical Laboratory; and in the synthetic basalt–andesite system by T. H. Green and Ringwood at Canberra, which show the effect that the presence of water greatly reduces the extension of the orthopyroxene primary phase volume into critically undersaturated compositions; in fact it causes an expansion of the olivine primary phase volume into silica oversaturated compositions (O'Hara 1968 *a*). This is the precise opposite of the situation required by Dr Green's hypothesis and it may, therefore, be premature to accept this hypothesis as a model to explain magma generation under the oceanic crust.

I turn now to the difficulties identified by several speakers concerning magma genesis under the mid-ocean ridges. The assumption that the ocean-floor tholeiites are primary magmas is the cause of these difficulties, which melt away when a more sophisticated model is adopted. The compositions of most ocean-floor tholeiites lie on the boundary of the primary phase volume of olivine at 2 to 5 kbar (0.2 to 0.5 GN m^{-2}), but outside it at higher pressures. They cannot

therefore be formed as primary partial melts of olivine-rich rocks at higher pressures. Worse, many of their compositions do not lie on the boundary of, or within, the primary phase volume of orthopyroxene at any pressure, hence they cannot under any circumstances be primary magmas derived from enstatite-bearing peridotites remotely similar to alleged upper mantle compositions. These difficulties are removed if it is accepted that olivine crystals have formed and fractionated in substantial amounts during the ascent of the magma. This argument has been developed in detail elsewhere (O'Hara 1968 b).

The calculation which purports to show that eclogite fractionation, followed by olivine fractionation, results in erupted liquids of too high an Fe/Fe + Mg ratio is invalid because it *assumes* the pyrolite model for the upper mantle, which has previously *assumed* that the basalt used in its calculation has *not* suffered eclogite fractionation. I suggest that it has, and that there are, therefore, two eclogite fractionation events concealed within Dr Green's calculation. This underlines my earlier point that the pyrolite model is unsatisfactory because it assumes the answers to the very questions it is used to answer, with predictable results.

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Written Reply to Discussion by M. J. O'Hara (D. H. GREEN)

The criticisms by Dr M. J. O'Hara are all-embracing, generalized and unsubstantiated by argument or comparable experiments. His alternative hypothesis (O'Hara & Yoder, 1967; O'Hara 1968) is given in general terms and has not been tested by simple arithmetical calculation, experimental investigation or even by examination for self-consistency. The misrepresentation in O'Hara's discussion of the term 'pyrolite' should be apparent from the 'Introduction' in my paper in this Symposium (see also Ringwood 1970, p. 135; Green 1970). Pyrolite is a clearly defined but general term for a hypothetical *chemical* composition; precisely two pyrolite compositions have been calculated (Green & Ringwood 1963; Ringwood 1966) and the later calculation reflects advance in factual knowledge of the high-pressure crystallization of olivine tholeiite and picrite magmas (Green & Ringwood, 1964, 1967 b). Experimental data on the subsolidus mineralogy of these calculated pyrolite compositions have been obtained over some years and the present knowledge is presented in Green & Ringwood (1969).

The pyrolite model *assumes only* that observed primary basalts are derived from a *peridotitic source rock* by partial melting. The only competing assumption to this is that basalts are derived from an eclogitic source rock—this is apparently not supported by O'Hara. Basalt magmas themselves place strong constraints on their source composition. Examples of alkali basalts, basanites, olivine nephelinites, etc., sometimes contain dense ($\rho = 3.2$–3.4) high pressure, xenoliths or xenocrysts which can be shown to be derived from $P > 10$ kbar (1 GN m^{-2}) and in some cases (Green & Hibberson 1969) from $P = 14$ to 16 kbar (1.4 to 1.6 GN m^{-2}). Such magmas are unmodified by crystal fractionation processes at low pressure. Examination of 94 analyses of such basalts of world-wide occurrence shows an extremely strong concentration of the atomic ratio $100\,Mg/(Mg + Fe^{2+})$ in the range 63 to 73. For a basalt of a given $100\,Mg/(Mg + Fe^{2+})$ ratio, the liquidus olivine, or residual olivine of a potential source peridotite, is

uniquely fixed—the Mg/Fe partition relation can be established from natural phenocryst/basalt data and from experimental studies of basalt crystallization. Basalts with $100\,Mg/(Mg+Fe^{2+})$ = 63 to 73 could be in equilibrium with olivine of $100\,Mg/(Mg+Fe^{2+})$ = 87 to 92. The pyrolite of Ringwood (1966a) has $100\,Mg/(Mg+Fe^{2+})$ = 89. Similarly, spinel lherzolite xenoliths in undersaturated basalts, high temperature peridotite intrusions, garnet lherzolite xenoliths in kimberlite have $100\,Mg/(Mg+Fe^{2+})\approx 88$ to 92. The pyrolite model interprets the latter rock types as ranging from possible 'pyrolite' composition to peridotites residual after extraction of a basaltic magma or volatile-rich component from 'pyrolite'. This aspect of the relationship of pyrolite composition to natural peridotites is treated in more detail in Green (1970) and Kleeman, Green & Lovering (1969).

The preceding discussion illustrates the search for an *internally consistent model* which can relate specific, observed natural basalt magmas to one another, to a potential source peridotite composition and to observed, natural, mantle-derived peridotites. It is indeed an assumption that such rock types are genetically and not accidentally related, yet this is an assumption that O'Hara (1968) also makes, though in an arbitrary manner, by selecting a particular garnet lherzolite xenolith (in kimberlite) composition as a source peridotite. O'Hara has not examined his models (O'Hara & Yoder 1967; O'Hara 1968) for internal consistency and, even more frustrating for the reader, does not express the models in terms of actual magma compositions, degrees of melting, composition of accumulates, etc. It is difficult not to emphasize the inadequacy and futility of attempting to evaluate complex magma compositions (> 10 components), with their highly significant variations in such diagnostic ratios as $Mg/(Mg+Fe^{2+})$, $Na/(Ca+Na)$, $Na/(Na+K)$, etc. by considering projections into pseudo-quaternary systems ('Al_2O_3' = $Al_2O_3+Fe_2O_3+Cr_2O_3+Na_2O+K_2O+TiO_2$; '$CaO$' = $CaO-3\frac{1}{3}P_2O_5+2Na_2O+2K_2O$; '$MgO$' = $MgO+FeO+NiO+MnO-TiO_2$; 'SiO_2' = $SiO_2-2Na_2O-2K_2O$, O'Hara 1968, p. 87). One can attempt to evaluate the O'Hara & Yoder (1967) and O'Hara (1968, p. 118.) models for internal consistency in the following way.[†] Accepting O'Hara's 1968, garnet lherzolite source, then the source has $100\,Mg/(Mg+Fe^{2+})\leqslant 92$ as deduced from the natural rocks. The partial melt (20%) at 30 kbar (3 GN m^{-2}) is postulated (O'Hara 1968, p. 118; O'Hara & Yoder, 1967, p. 115) to be a tholeiitic picrite—this magma must have $100\,Mg/(Mg+Fe^{2+})\leqslant 72$ if it is less than 20 to 30% melt of the source rock. The picrite is postulated to precipitate some olivine and orthopyroxene (decreasing the liquid $100\,Mg/(Mg+Fe^{2+})$ value), before passing through a reaction of these phases with liquid to precipitate garnet and clinopyroxene ('eclogite'). O'Hara (1968) postulates 50% eclogite fractionation of such a magma followed by a further 40% of olivine crystallization to yield derivative nepheline-rich magmas observed at the earth's surface. Applying the Mg/Fe partition factors for the various phases and their compositions as observed in experiments one can show that the derivative magmas would have $100\,Mg/(Mg+Fe^{2+}) < 55$[‡] and probably < 50—the actual observed magmas of demonstrable direct mantle derivation have $100\,Mg/(Mg+Fe^{2+})$ = 63 to 73.

O'Hara's discussion of this type of calculation illustrates his failure to evaluate critically the implications of *his own model* in terms of specific compositions. For a quantitative evaluation of the role of eclogite fractionation based on experimental high-pressure data, analysed minerals etc., the reader is referred to Bultitude & Green (1971, in press).

[†] I would strongly urge that the authors themselves attempt such evaluation. In this way one could hopefully see whether the model is a viable alternative to the pyrolite basalt genesis models.

[‡] Such magmas would have liquidus olivine of $100\,Mg/(Mg+Fe^{2+})$ = 80 to 83.

Turning to the role of water in basalt fractionation and partial melting, a central point is again the danger of attempting to extrapolate from simple 2 or 3 component systems to > 10 component systems. The relevant experimental data on (alkali basalt + water) and (basanite + water) compositions are referred to in Green & Hibberson (1969) and Green (1969 b, p. 229), and on (quartz tholeiite + water) and (tholeiite + water) compositions in Green (1970 c).

It should also be noted that Kushiro (1970) found that the low melting composition in the system $Fo–Ne–SiO_2–H_2O$ lay in the nepheline normative volume and commented that this brought the simple system studies closer to the observations on natural rock compositions. Actually, the enhanced role of orthopyroxene crystallization in nepheline-normative, olivine-rich compositions in the presence of water, is due not so much to contraction of the olivine crystallization field in favour of orthopyroxene but rather is due to the broadening of the pyroxene miscibility gap at lower temperature, introducing orthopyroxene additional to or in place of clinopyroxene. While I do not accept the validity of O'Hara's comments, I would emphasize that the calculation of model pyrolite composition(s) is bound to require future refinement and that some of the details of the petrogenetic model will probably be proved wrong or inadequate—the paper is presented as the current status in a rapidly progressing branch of petrology.

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