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Magma Generation in the Upper Mantle, Field Evidence from Ophiolite Suites, and Application to the Generation of Oceanic Lithosphere [and Discussion]

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Magma generation in the upper mantle, field evidence from ophiolite suites, and application to the generation of oceanic lithosphere

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The Bay of Islands ophiolite may be divided into a lower series of ultramafic tectonites representing mantle material and a higher series of cumulate and extrusive rocks and sediments which may be correlated with oceanic crust. The tectonite series consists of a lower spinel–lherzolite member overlain by harzburgites. Both are cut by numerous olivine–pyroxene veins which represent early crystallization products from a picritic tholeiite magma derived at 18–22 kbar by approximately 23 % partial melting of the spinel lherzolite. The remainder of this magma crystallized as differentiated cumulate and extrusive rocks under crustal conditions.

The petrogenetic model derived implies a rising diapiric body beneath an accreting centre and allows for the production of tholeiitic, transitional and mildly alkaline basalts from a single parent. The nature of the basalt erupted depends upon the rate of upwelling and crystallization in the diapir. The model suggests that crystallization takes place in the diapir at depths of less than 60 km.

INTRODUCTION

The problem of the genesis of oceanic crust may be investigated in a number of ways. Those that require examination of present-day oceanic material are, however, hampered by logistical difficulties in sampling beneath considerable depths of water and, as yet, no method has been devised to sample the mantle directly below the crust.

In recent years most Earth scientists have agreed upon an oceanic origin for the ophiolite suites observed in many mountain belts. These rocks are believed to have been displaced from their oceanic régime by plate tectonic mechanisms during the early stages of orogenesis resulting from plate collision. They now provide ideal situations in which to investigate the origin of crustal rocks of the ocean floor.

Complete ophiolite sequences form the four massifs of the Bay of Islands Complex in western Newfoundland as the upper members of a number of thrust slices emplaced during the Taconic phase of Appalachian Mountain building (Williams 1973). It is the purpose of this paper to examine the petrogenesis of the ophiolite suite.

Lithologies

Stratigraphic columnar sections of the three northernmost massifs are presented in figure 1. Two of the massifs (Blow-me-down and North Arm Mountains) display a completely developed ophiolite sequence. Each of the massifs has a basal portion composed of ultramafic rocks ranging from approximately 2.5 km thick on North Arm Mountain to nearly 5 km thick on Table Mountain which are immediately underlain by a dynamothermal aureole produced during displacement of the thrust slices (Malpas, Stevens & Strong 1973; Williams & Smyth 1973). These ultramafic rocks are dominantly harzburgitic although a considerable thickness of spinel lherzolite occurs at the base of Table Mountain. Both types clearly display tectonite fabrics (Nicolas, Bouchez, Boudier & Mercier 1971) thought to have developed by solid state deformation in the mantle. The tectonite fabrics both cut and are cut by numerous pyroxenite and dunite veins and dikes which were quite clearly intruded as liquids during the deformation of the host ultramafics. Immediately overlying the harzburgites are dunites and feldspathic dunites displaying cumulate rather than tectonite fabrics. Orthopyroxene is only rarely found in these rocks but clinopyroxene and chromite are often quite common. Gabbroic rocks lie above the ultramafic rocks and an interbanded zone separates these two rock types. This zone has been called the 'critical zone' by Smith (1958) and Irvine & Findlay (1972) and reaches thicknesses of the order of 400 m. On Table Mountain the gabbros are not complete but

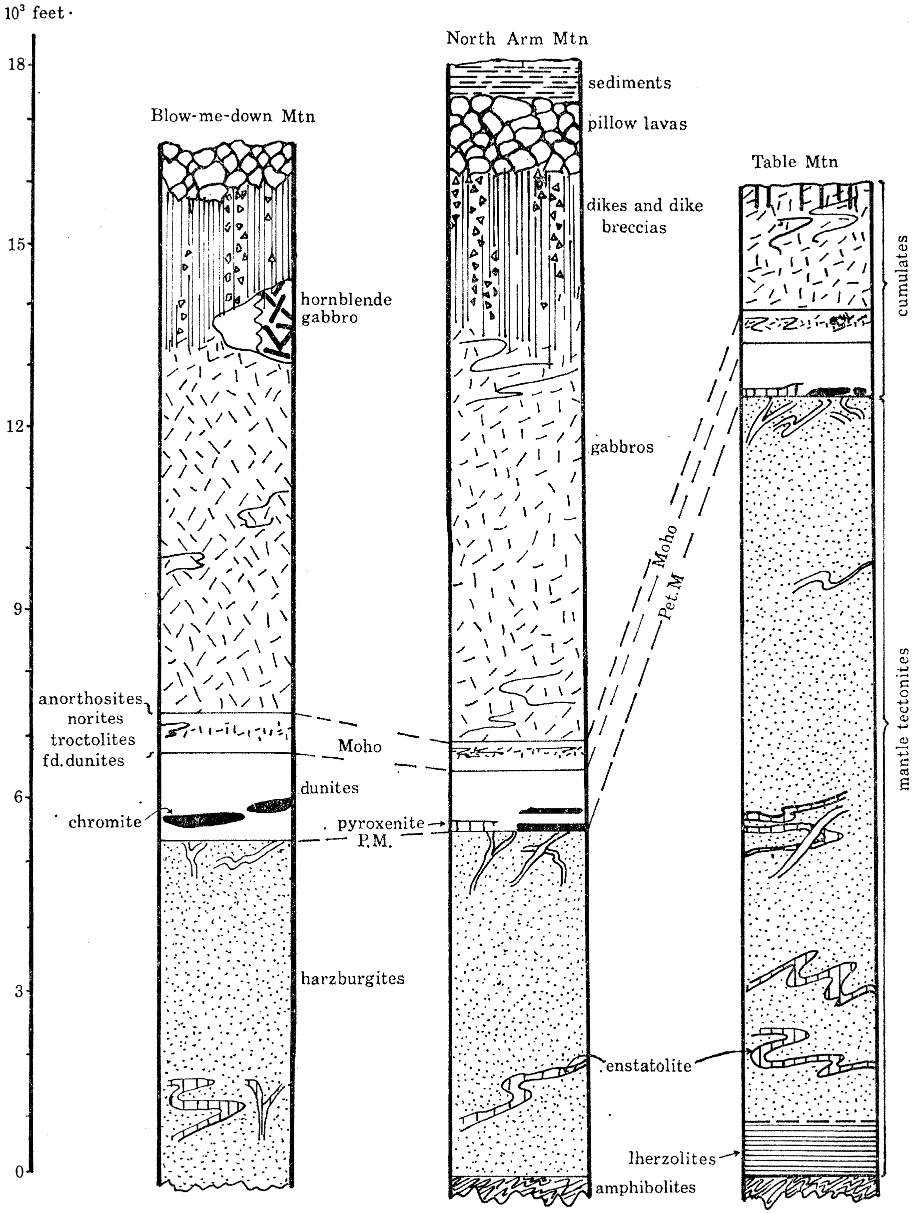

FIGURE 1. Columnar sections of Bay of Islands massifs. P.M. = petrologic Moho.

where the full sequence is developed on North Arm and Blow-me-down Mountains, they reach an approximate thickness of 4 km. On these two massifs, diabase dikes increase in quantity at the top of the gabbro and higher up they become 'sheeted' (100 % diabase dikes) and feed overlying pillow lavas. The diabases and pillow lavas reach a total thickness of about 2.5 km where they are overlain by siltstone and silty-sandstones.

Chemistry and petrogenesis

Chemical analyses have been obtained for a number of minerals and rocks of the Bay of Islands Complex. These analyses are illustrated diagrammatically here. Full analyses and details of analytical methods can be obtained on request.

Mineral chemistry

Olivine

Fo contents (mol %) of olivines, as determined by microprobe analysis, vary from 75.1 to 90.8 in the cumulate rocks, with the lower values representative of the gabbros above the dunites. Olivines from the lherzolites have Fo contents of approximately 89.5. The average Fo content of olivines from the harzburgites is slightly higher (91.1) and the range is very small (90.5–91.6). A number of these analyses have been supplemented by determination of olivine composition by X-ray methods using the determinative curve based on the olivine (130) spacing (Yoder & Sahama 1957), and slight discrepancies are attributed to differences in trace element content, e.g. dunitic olivines appear to contain more manganese and nickel, and in some cases more chromium than those from the harzburgites. The electron microprobe analyses of olivines show them to be chemically homogeneous; there are no significant differences from grain to grain within any one sample and no zoning within individual grains.

A number of workers (e.g. Jackson 1961; Smith 1962) have suggested that olivines of alpine-type intrusions are generally more magnesian than those from layered intrusions. Although Challis (1965) has pointed out that data on olivine compositions do not really allow conclusions to be drawn on the origin of ultramafic rocks, if the rock types from the Bay of Islands Complex are classified on the basis of texture as cumulate or tectonite, then a slight difference in Fo content of the olivines is noted, the tectonite olivines being generally more magnesian, although there is an overlap in compositions. It appears that the Fo values in the tectonites are relatively constant, whereas in the dunites the values range between 90.8 and 83.4. The Cr_2O_3 content of the dunitic olivines is relatively high (up to 0.16 % by mass), possibly indicating relatively low fo_2 during crystallization. Such conditions might be expected to result in exsolution of Ni metal in the olivines, but in sections examined at this stage none has been identified.

Orthopyroxenes

The orthopyroxenes analysed are all enstatites with En values between 88.5 and 92.1, and no marked difference in En value occurs between harzburgitic, lherzolitic and dunitic pyroxenes. This is typical of the compositional range of orthopyroxenes from alpine peridotites (Ross, Foster & Myers 1954; Green 1964; Page 1967; Himmelberg & Coleman 1968). Although Hess (1960) indicated that the Ca^{2+} content of orthopyroxenes was representative of the magma type from which the pyroxene had crystallized, further evidence (Jackson 1961; Rothstein 1958; Challis 1965) seems to dispute this. The work of Atlas (1952), however, suggests that Ca^{2+}

content may be dependent on crystallization or equilibration temperature. Values for many of the pyroxenes analysed here exceed 0.050 (maximum 0.072) possibly suggesting equilibration temperatures in excess of 1000 °C. If this is the case then (i) other pyroxenes in these rocks have not accepted as much calcium into their structure as theoretically possible or (ii) if it was there, it has subsequently been removed. The exsolution lamellae of clinopyroxene in some enstatites support the latter conclusion.

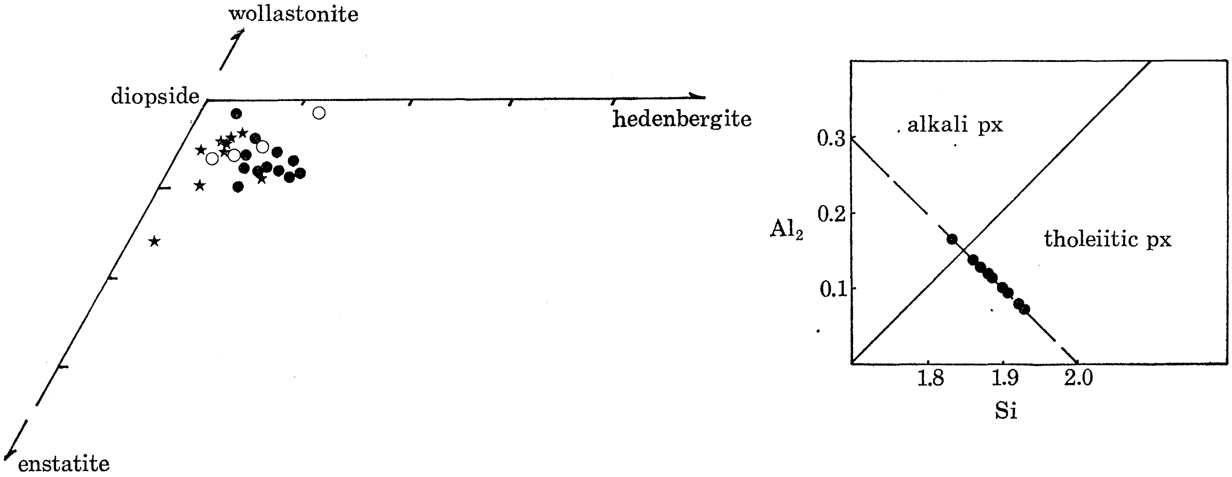

FIGURE 2. Clinopyroxene compositions: ★, harzburgite pyroxenes; ○, lherzolitic pyroxenes; ●, cumulate pyroxenes.

Clinopyroxenes

In the harzburgites the clinopyroxene occurs as small interstitial crystals and is relatively rare. In the lherzolites the clinopyroxenes are texturally identical but are more common. In both cases, the clinopyroxene is clearly not a cumulate phase. On the other hand, in the cumulate olivine-pyroxenites and some gabbros, clinopyroxene occurs as large crystals poikilitically enclosing olivine, chromite and plagioclase, and appears in these cases to be the last mineral to crystallize.

The analyses of the clinopyroxenes are represented diagrammatically in figure 2 where they all lie in the diopside or endiopside fields (Poldevaart & Hess 1951). The clinopyroxenes of the cumulate rocks are generally distinguishable from those of the harzburgites and lherzolites by their more iron-rich compositions and a corresponding titanium enrichment, although there is some overlap in values. This titanium enrichment seems to be less dependent on pressure of crystallization (Akella & Boyd 1973) than on availability of titanium and its substitution for iron. The clinopyroxenes of the lherzolites and harzburgites have higher chromium contents than those from the cumulate rocks, except where the pyroxene is from chromite-rich bands, and this enrichment in the Kosmochlor molecule is comparable in amount with that in the chrome-diopsides produced in the experiments of Dickey & Yoder (1972) at pressures up to 20 kbar.

The lherzolitic clinopyroxenes are richer in both Al_2O_3 and Na_2O compared with the harzburgitic pyroxenes. This may be taken as an effect of equilibration under higher pressures (Kushiro 1960; Rothstein 1962; Kuno 1964; Green & Ringwood 1967; Aoki & Kushiro 1968; Burns 1970; Irving 1974; Munoz & Sagredo 1974). Church (1972) recognized similar differences in Al_2O_3 contents, and divided the rocks on this basis into 'high-pressure' and 'low-pressure' associations. The present writer agrees with Church in this division but suggests that their re-equilibration need not, as he states, have involved crystallization from a magma body (Church 1972, p. 79).

Since diopside appears to have been the last mineral to crystallize in the gabbros and olivine pyroxenites, its composition may be expected to reflect the composition of the residual magma at this stage. Kushiro's (1960) plot has been modified in figure 2 so that only substitutions in the tetrahedral sites are plotted. In the diagram, all but one of the pyroxenes plot clearly in the tholeiitic field. At first, this seems incompatible with the slightly alkaline nature of some bulk rock analyses but may be explained in terms of pressure of crystallization, and an increased solubility of enstatite in diopside in pyroxenes crystallizing at pressures greater than 1 atm.† This effect may push the clinopyroxene composition into the tholeiitic field, although the magma composition is slightly alkaline.

Coexisting orthopyroxene and clinopyroxene; estimation of equilibration temperatures

The clinopyroxenes from the harzburgites generally have a higher $Mg/(Mg+Fe^{2+}+Mn)$ ratio and a slightly lower Al_2O_3 content compared with the coexisting orthopyroxenes. The Al_2O_3 is present chiefly in Tschermak substitution [$(Ca, Mg)\ Al_2SiO_6$] and not as jadeite. These clinopyroxenes also have slightly more Cr_2O_3 than the orthopyroxenes, although the difference is not significant.

One method for determination of temperatures of equilibration of peridotites has been the composition of diopside coexisting with enstatite using the solvus as determined by Davis & Boyd (1966) and Boyd & Schairer (1964). The distribution of Ca^{2+} between coexisting ortho- and clino-pyroxenes from the lherzolites of the Bay of Islands Complex suggests that they last equilibrated on the pyroxene solvus at about 1000 °C. Re-examinations of the solvus at lower temperatures (Nehru & Wyllie 1974; Lindsley & Munoz 1969) and higher temperatures (Howells & O'Hara 1975; Mori & Green 1975) indicate that the solubilities of enstatite in diopside are greater than previously thought at temperatures below 1200 °C. It is probable, therefore, that temperature estimates below 1200 °C might be too low. The applicability of any of these solvi determined in olivine-free assemblages to olivine-saturated natural assemblages is questionable in any case, since the activity of silica has a major influence upon the $Ca/(Ca+Mg)$ ratio of diopside coexisting with enstatite, at least at high temperatures (Howells & O'Hara 1975).

Other estimates of temperature and pressure of equilibration may be obtained using the methods of O'Hara (1967), Wood & Banno (1973), Wood (1974) and MacGregor (1974). Because most of these methods depend on the coexistence of an aluminous phase for the estimation of pressure, only clinopyroxene compositions from the lherzolites are used. In these cases, spinel is the aluminous phase. O'Hara's (1967) α and β parameters for clinopyroxenes from the lherzolites indicate temperatures of the order of 1000–1100 °C and pressures of about 18–20 kbar.‡ If the spinel–lherzolite/garnet–lherzolite phase boundary is slightly modified to increase the pressure range of the spinel–lherzolite field because of the influence of Cr_2O_3 in the system on the pressure required to stabilize garnet–lherzolite (MacGregor 1970), the data points fall clearly within the spinel–lherzolite facies field.

Using partition coefficients of Mg^{2+} and Fe^{2+} between coexisting orthopyroxenes and

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† 1 atm ≈ 10^5 Pa.
‡ 1 kbar = 10^8 Pa.

clinopyroxenes (Wood & Banno 1973), similar temperatures are indicated. Temperature estimates for the harzburgites are also possible by this method, which does not depend on an accompanying aluminous phase. With these temperatures the curves of Wood (1974) and MacGregor (1974), based on Al^{2+} substitution in M_1 sites of orthopyroxenes coexisting with an aluminous phase, give pressures of 17–20 kbar for the lherzolites.

Thus all methods give relatively consistent temperatures and pressures of re-equilibration of about 1000 °C and 18–20 kbar. This corroborates textural evidence for a certain amount of subsolidus recrystallization.

Spinel group

Spinel end members have been calculated according to the method of Irvine (1965) and end member proportions may be plotted in the Johnson spinel prism (Malpas & Strong 1975). All analyses plot close to the spinel, hercynite, chromite, picrochromite plane. Projection of these data points onto the two planes expressing $Cr/(Cr+Al)$ against $Mg/(Mg+Fe^{2+})$ and $Fe^{3+}/(Cr+Al+Fe^{3+})$ show a number of features. Most importantly, the spinels show a large variation in composition, especially with regards to $Cr/(Cr+Al)$. Generally the $Mg/(Mg+Fe^{2+})$ ratio increases as the $Cr/(Cr+Al)$ ratio decreases. This has been shown to be the general case for alpine peridotites (Irvine 1967; Himmelberg & Coleman 1968; Loney, Himmelberg & Coleman 1971). The harzburgite spinels show distinctly higher Cr^{3+} than those of the dunites. They also show correspondingly (although only slightly) lower $Mg/(Mg+Fe^{2+})$. In general the dunite spinels contain more iron in the Fe^{3+} state than those of the harzburgite. Spinel from the lherzolite has a $Mg/(Mg+Fe^{2+})$ ratio within the limits for pleonaste spinels (Deer, Howie & Zussman 1962). These spinels are highly aluminous by comparison with other spinels from the harzburgites. It is notable that the Cr_2O_3 in the clinopyroxenes of the lherzolites probably accounts for much of the bulk rock Cr_2O_3.

Irvine & Findlay (1972) suggested a correlation between chromite cell size and calcium content of the bulk rock, as a result of reaction between liquid and chromite to produce a more aluminous spinel and clinopyroxene. Loney *et al.* (1971) similarly suggest that the different compositions of the chrome spinels in alpine peridotites are not a result of different conditions of crystallization, but reflect differences in original bulk rock composition. Malpas & Strong (1975) have suggested that differences in environment of crystallization may, however, account for much of the variation in the Bay of Islands spinels, in as much as Cr^{3+} is concentrated in residual phases (Dickey & Yoder 1972; Burns 1973).

Coexisting olivine and chromian spinel

Cation distribution between coexisting olivine and chromian spinel has been discussed by Irvine (1965), Jackson (1969) and Loney *et al.* (1971). Estimated temperatures for coexisting chrome spinels and olivines are of the order of 1300 °C in the harzburgites and 1200 °C in the dunites, up to 300 °C higher than the other temperature estimates. The range of temperature estimates may not, however, be so significant and represents small analytical errors and estimates in recalculation of results. For temperature calculations, Fe^{2+} was taken as total iron for the olivines. Jackson (1969) estimates that variations of up to ± 300 °C are possible when errors in spinel free-energies alone are considered.

Bulk rock chemistry

The variation in degree of serpentinization of the ultramafic rocks results in variations of H_2O content which makes it difficult to compare chemical analyses. Interpretation of the analyses is also complicated by the possibility that serpentinization may have removed major constituents. Recent literature tends to suggest, however, that this is not generally the case (Green 1964; Hostetler, Coleman, Mumpton & Evans 1966; Barnes, La Marche & Himmelberg 1967; Himmelberg & Coleman 1968; Loney *et al.* 1971). Another major variation in the chemistry of the peridotite samples is the Fe_2O_3 content. This largely reflects the amount of magnetite present and this in turn is related to the degree of serpentinization. Except in chrome spinels, Fe_2O_3 is not a major component of the primary minerals of the peridotites.

Calculation of C.I.P.W. norms was carried out after adjustments for hydration and oxidation (Fe_2O_3 recalculated as 1.5 mass %).

The analyses have been compared on a number of standard variation diagrams (figure 3)

F.m.a. diagram

In figure 3 the Bay of Islands Complex rocks are compared with standard differentiation trends of Skaergaard and Hawaii, and trends exhibited by the recognized ophiolite suites of Papua and Oman. All the ophiolites plot along a similar trend of iron enrichment typical of tholeiitic suites (see also Norman & Strong 1975), and are comparable with the general trend suggested by Thayer (1967) for alpine peridotites. The spread of the gabbroic rocks apparently indicating their enrichment in alkalis might be explained by their various concentration of cumulate plagioclase. The enrichment of some basalts in alkalis is possibly attributable, to a certain degree, to alteration. The quartz–diorites are apparently crystallized from the final residual liquid of the fractionation series.

The f.m.a. diagram (figure 3*a*) clearly indicates that the pillow lavas are crystallized from a magma that has already undergone fractionation, and are not representative of a primary melt.

NiO *against* Cr_2O_3 (*figure* 3*c*)

Irvine & Findlay (1972) used Cr_2O_3 and NiO relations to separate alpine type peridotites and cumulate rock associations. They suggested a similarity between alpine peridotites and the tectonic harzburgites and lherzolites of the Bay of Islands Complex and attributed to them a mantle origin. They distinguished rocks of the critical zone and gabbro zone from the tectonites with the suggestion that their similarity to layered intrusions indicated that they crystallized under relatively low pressures in the crust. The present study supplements this view in that rocks from the dunite zone, below the critical zone, and not specifically identified by Irvine & Findlay may also be considered of the same origin. These rocks are clearly cumulate and must therefore be considered as part of a low pressure fractionation series which includes the critical zone, gabbros, diabases and basalts, and which is distinct from the ultramafic tectonites which it overlies (Malpas 1973).

Total alkalis against SiO_2 (*figure* 3*b*)

The total alkalis versus silica diagram shows that although the majority of the rocks from the ophiolite suite can be considered tholeiitic, a number are transitional or definitely alkaline

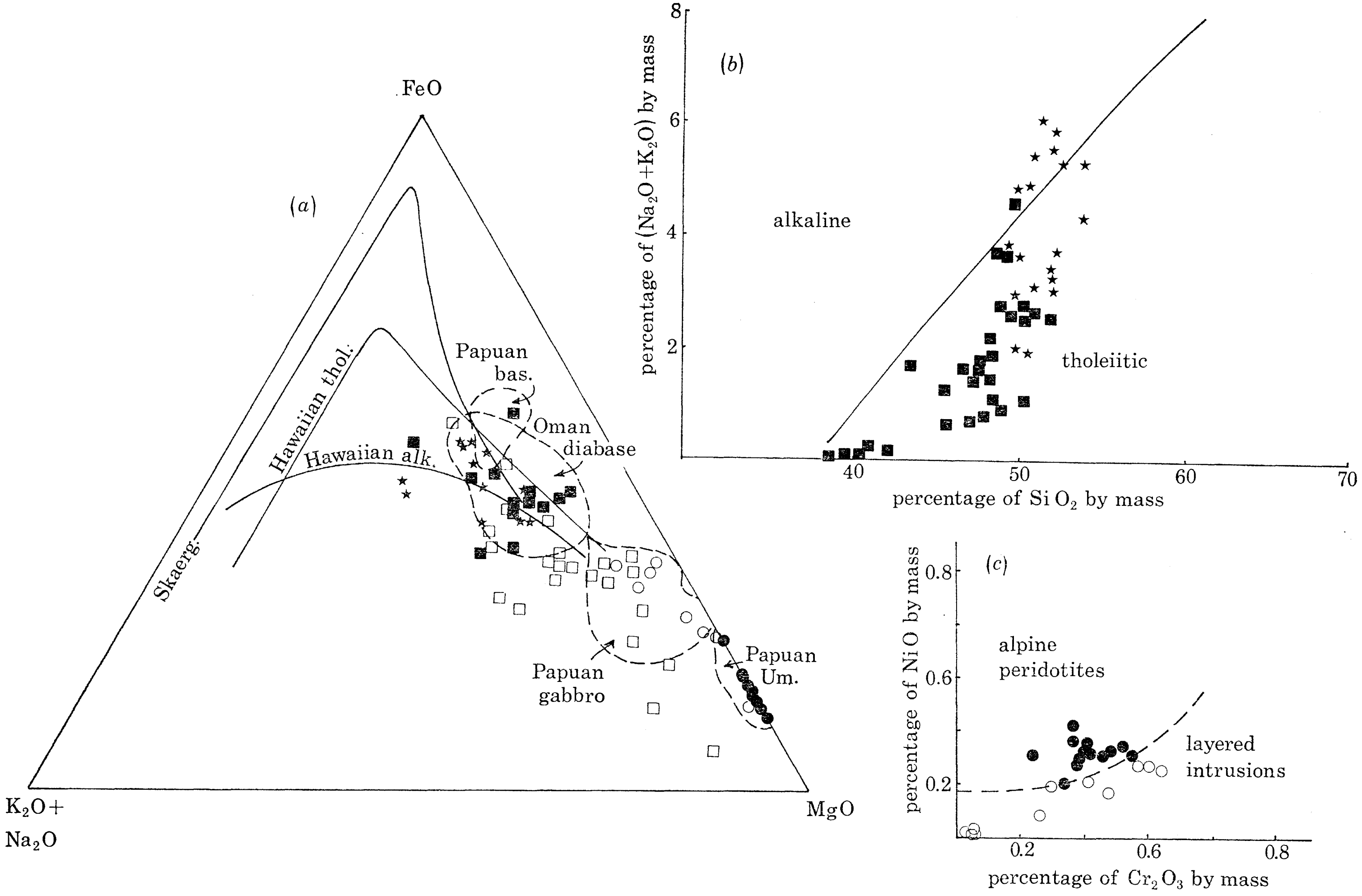

FIGURE 3. Bulk rock compositions: (*a*) f.m.a. plot; (*b*) total alkalis plotted against SiO_2; (*c*) NiO plotted against Cr_2O_3. ●, tectonites; ○, dunite cumulates; □, gabbros and feldspathic dunites; ■, diabases; ★, basalts.

in their present chemistry. There are two possible reasons for this alkaline affinity. Firstly, the rocks are altered and the increased alkalinity is due primarily to soda-metasomatism during low-grade metamorphism. It is notable that in this regard the two samples which plot furthest into the alkaline field are clearly metasomatized pillow-selvages. However, the cores of these pillows are themselves transitional in composition. The second alternative is that the rocks form a series that was primarily transitional in nature and which is comparable to rocks described by Aumento (1968) from Confederation Peak and Muir & Tilley (1964) from Discovery Tablemount, both in the mid-Atlantic. The importance of each of these alternatives must be considered after further evidence cited below.

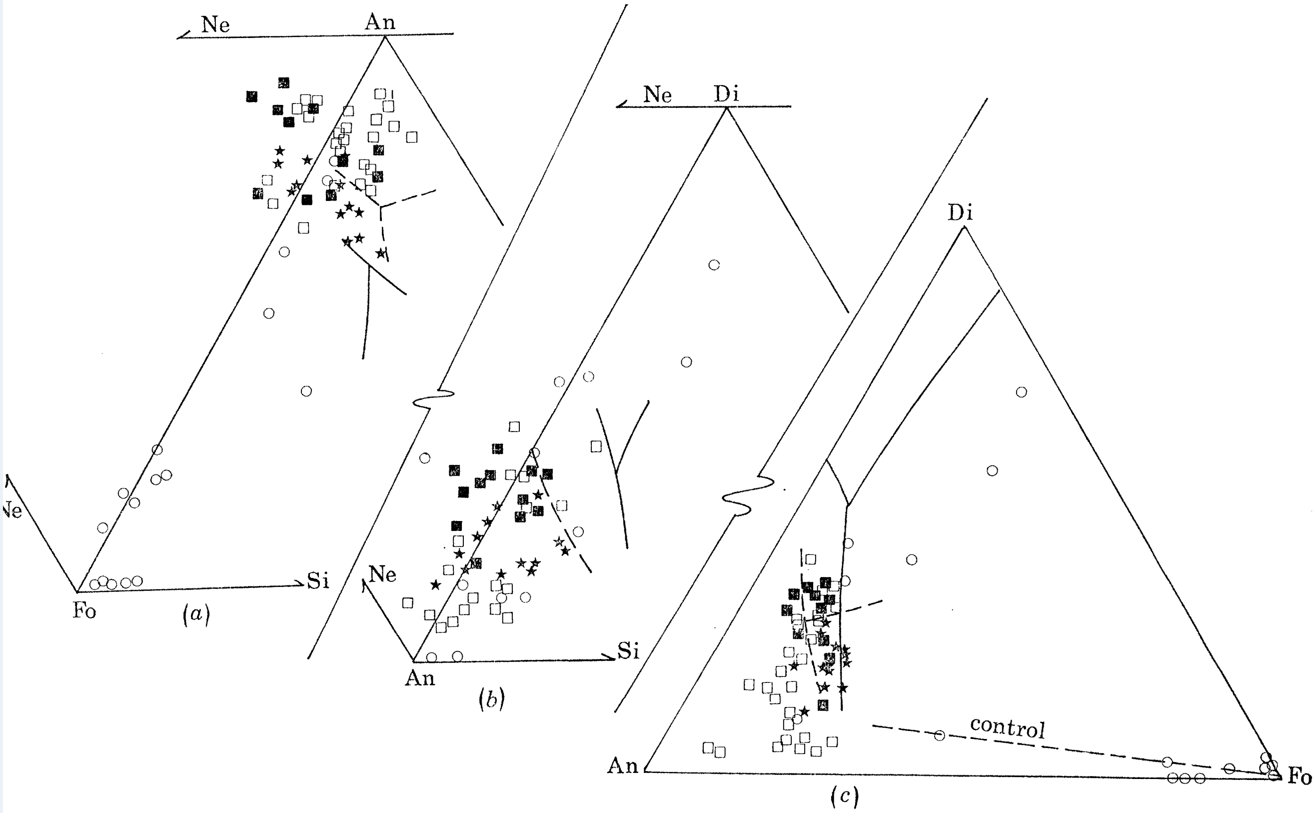

Figure 4. Basalt tetrahedron projections: (*a*) from diopside; (*b*) from forsterite; (*c*) from nepheline/qtz. Broken lines, natural system cotectics; solid lines, synthetic system cotectics. Symbols as for figure 3.

Basalt tetrahedron (*figure* 4)

The cumulate rocks, diabases and volcanics from the Bay of Islands Complex have been plotted as projections within the simple normative basalt tetrahedron of Yoder & Tilley (1962). The cotectics are obtained from Clarke (1970) and based upon experimental work by Tilley, Yoder & Schairer (1963, 1964, 1965, 1967).

In the olivine projection a considerable spread of points from the dunite zone is a result of their close proximity to the projecting phase such that a small variation in their composition is reflected as a large spread after projection. Relative to the natural system cotectic in the olivine projection, plagioclase appears to be the second phase to be followed by clinopyroxene.

An olivine control line is suggested in the diopside projection, with the basalts falling between the natural and synthetic system plagioclase–olivine cotectic. The gabbros are clearly enriched in cumulate plagioclase. Since all points plot at a distance from quartz or nepheline in these projections, a projection from quartz/nepheline is of some use in indicating liquid lines of descent. Such a projection clearly shows the initial olivine control and the crystallization sequence, i.e. (1) olivine, (2) olivine + plagioclase, (3) olivine + plagioclase + clinopyroxene. This sequence is exactly that observed in the field and is upheld by textural evidence in the gabbros where olivine is found included in plagioclase and clinopyroxene is generally intercumulate. The basalts clearly plot around the low pressure natural system eutectic, and the dikes along the plagioclase–olivine cotectic in this projection.

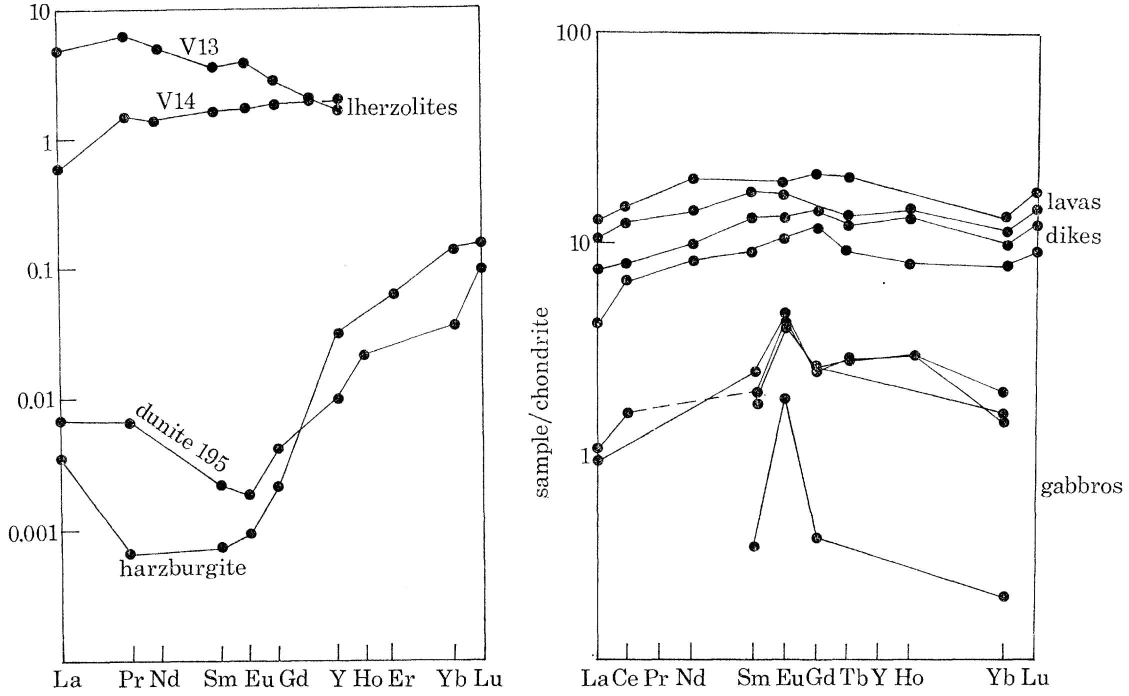

FIGURE 5. Rare earths and yttrium for Bay of Islands rocks.

All projections within the basalt tetrahedron indicate a spread of compositions across the critical plane of silica undersaturation. The spread observed is difficult to relate entirely to alteration, the effects of varying Fe_2O_3/FeO ratio having been reduced for these plots by normalizing the ratio of 0.25 for all rocks. The projections therefore substantiate that the suite is transitional in nature.

Rare earth elements (*figure* 5)

Determinations of rare earth elements (r.e.e.) and yttrium were carried out on an A.E.I. MS7 spark-source mass spectrometer. Results are shown diagrammatically in figure 5 where the rare earth abundances have been normalized against chondritic abundances and plotted against atomic weight. The following points are noteworthy:

(*a*) There is a marked similarity of the r.e.e. abundances of the pillow lavas with mid-Atlantic Ridge basalts (data from Frey *et al.* 1968). They show a slight depletion of the light rare earths (*ca.* 11 times chondritic l.r.e.e.) with respect to chondritic r.e.e. abundances.

Oceanic alkaline basalts usually show a preferential enrichment in l.r.e.e. (Frey 1970; Graham & Nicholls 1969).

(*b*) There is generally a lower r.e.e. abundance and stronger l.r.e.e. depletion observed in the diabases compared with the pillow lavas. This would indicate that the lavas are slightly more fractionated than the diabases, which agrees with the results of major element chemistry.

(*c*) The gabbros are relatively poor in r.e.e. However, definite Eu enrichment anomalies are observed indicating a strong accumulation of plagioclase. This agrees with the conclusion from textural evidence that the gabbros are plagioclase cumulates and suggests that their crystallization took place under conditions of low oxygen fugacity. Comparison of patterns from oceanic gabbros show that these are also plagioclase cumulates. Eu anomalies are small in the lavas and diabases.

(*d*) The dunite 195 shows strong enrichment in h.r.e.e., but the overall pattern is V-shaped and characteristic of olivine (Frey 1970). The enrichment in h.r.e.e. is due to the presence of clinopyroxene in the dunites, this phase preferentially incorporating the heavier atomic weights.

(*e*) The harzburgite r.e.e. pattern is typical of residual peridotites after the removal of a tholeiitic partial melt (Frey, Haskin & Haskin 1971; Frey & Green 1974) showing a strong depletion of l.r.e.e. with respect to chondritic abundances. The low r.e.e. abundances are typical of rocks containing olivine and orthopyroxene and patterns are comparable with the Lizard residual peridotite (Frey 1970). It is notable that, if the dunite (195) represents an early crystallization product of the melt derived by partial melting leaving the harzburgite as residue, then a similarity in their r.e.e. patterns might be expected if the melt was derived under equilibrium conditions. From the data available, this seems to be the case.

(*f*) The variability of r.e.e. patterns for lherzolitic rocks has been pointed out by Phillpotts, Schnetsler & Thomas (1972), and Frey & Green (1974). The two lherzolites analysed here are similarly diverse, with V14 showing a steady although slight enrichment in heavy rare earths, indicating that it is probably partly depleted in tholeiitic melt. V13, on the other hand, shows an enrichment in light rare earths which may not be related to the process of partial melting as reflected in the resultant tholeiitic magma but can be explained as the process described by Frey & Green (1974).

The two divergent patterns at least indicate the inhomogeneity of the lherzolites within even a small volume.

Conclusions

The mineral and bulk rock chemistry support the following conclusions.

Nickel/chromium ratios, Fo contents of olivine, spinel compositions and petrographic evidence all suggest different origins for the tectonite peridotite and cumulate mafic series. Such arguments are supported by the work of Irvine & Findlay (1972). The cumulate mafic series, diabases and pillow-lavas are genetically related. These rocks are the result of fractionation, presumably at low pressure, of olivine, clinopyroxene and plagioclase from a low-Ti and low-K picritic magma derived from a mantle source. The pillow lavas are part of this fractionation series and not crystallization products of a primary magma. There is a systematic pattern of compositional variation from basalts, through diabases and gabbros in present oceanic tholeiitic suites, suggesting that modern abyssal tholeiites likewise cannot be considered unfractionated primary magmas, but rather crystalline products of a fractionated magma at low-pressure eutectics (O'Hara 1968; Norman & Strong 1975). The gabbros are clearly not melts but cumulates formed from the low-pressure crystallization sequence (*a*) olivine, (*b*) olivine and plagioclase, (*c*) olivine and plagioclase and clinopyroxene. This is supported by textural evidence. The dikes and pillow lavas are liquid compositions themselves indicating a differentiation trend. Although the suite is generally tholeiitic in composition there are strong suggestions that it is transitional to mildly alkaline in parts. This is supported by petrographic analyses which indicate the absence of orthopyroxene in most of the fractionation series.

The high proportion of harzburgite to overlying cumulates does not support any hypothesis that these peridotites formed by cumulate processes from a basaltic magma. The low pressure anhydrous (< 5 kbar) reaction relation between olivine and orthopyroxene, which persists to higher pressures even under low water contents (Kushiro, Syono & Akimoto 1968) prevents the simultaneous crystallization of olivine and orthopyroxene in constant proportions as observed in the harzburgite. The harzburgites exhibit very little cryptic mineralogical or bulk chemical variation as might be expected in cumulate systems and what is observed can be attributed to interstitial liquid re-equilibrating with the harzburgite (Irvine & Findlay 1972). The harzburgites consist of an assemblage of highly refractory chemistry with bulk Mg/(Mg + Fe) ratios of 0.94, alkali content and CaO and Al_2O_3 generally less than 1 % and low Ti and P. Finally field evidence reveals differences in textures and structures, the harzburgites being distinctly tectonites whilst the overlying cumulates are relatively free of deformation structures. Temperatures and pressures of equilibration of the present mineralogy of the tectonites appear to be of the order of 1000 °C and 18–20 kbar.

The harzburgites are therefore thought to represent depleted upper mantle produced by the partial melting of aluminous upper mantle peridotite and subsequent removal of the basaltic liquid (Hamilton & Mountjoy 1965; Menzies & Allen 1974). The further extraction of much basaltic liquid would seem impossible from the low abundance of elements that are liquid accumulative, and from the highly magnesian nature of the harzburgite mineral phases. Spinel lherzolite rocks may be modified mantle material which retain evidence of incomplete extraction processes in the form of clinopyroxene and spinel, or primary mantle compositions showing little or no partial melting.

The bulk of the dunitic rocks in the Bay of Islands Complex lie above the harzburgites. Dunite veins and dikes within the harzburgites generally have sharp contacts with the host rock, and are believed to be the result of crystallization of olivine and chromite from liquids trapped in the harzburgite during its convective ascent, probably below a spreading ridge axis. There is no evidence for residual 100 % or 90 % olivine dunites which would be expected to have a much higher Cr, Ni and lower Ti, Fe and Mn content than refractory harzburgites (Burns 1973). A cumulate origin can be similarly argued for the main dunite zone, suggesting that it forms the basal portion of the fractionated stratiform series, and this is especially evident where it contains euhedral chromite accumulations strung out parallel to the contact with the harzburgites.

The significance of the contact between the harzburgite and overlying dunite is that it must represent the junction between rocks of mantle origin and rocks which formed under crustal conditions in relatively high-level stratiform intrusions. The horizon has been called the 'petrological Moho', or the genetic discontinuity between crust and mantle (Malpas 1973). Such a discontinuity was suggested by the models of Greenbaum (1972) for the Troodos Complex, Cyprus. The Critical Zone, higher in the succession, represents the Moho as defined geophysically, since it is here that major density changes take place in the lithologies.

Petrogenesis

C.m.a.s. projections (*figure* 6)

One atmosphere phase relation can generally be shown adequately in the normative basalt tetrahedron. However, since its usefulness is restricted to portrayal of processes acting near the surface, phase relations at mantle pressures, which determine the compositions of partial melts produced at depth, are more usefully shown in the system CaO–MgO–Al_2O_3–SiO_2 (c.m.a.s.).

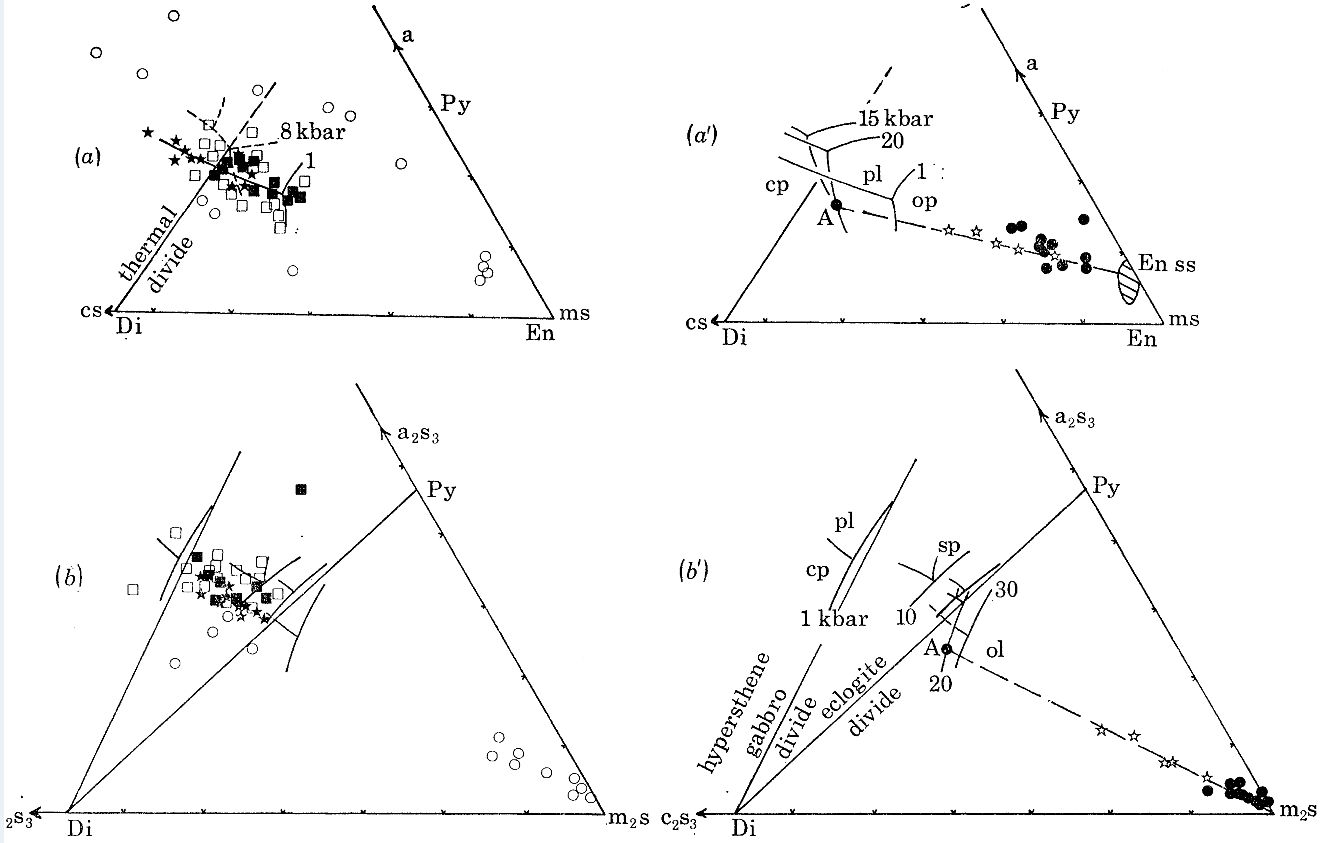

FIGURE 6. C.m.a.s. system projections: (*a*), (*a*′) from olivine on to cs–ms–a; (*b*) (*b*′) from enstatite on to c_2s_3–m_2s–a_2s_3. Di, diopside; En, enstatite; Py, pyrope; pl, plagioclase; op, orthopyroxene; cp, clinopyroxene; ol, olivine; sp, spinel; other symbols as for figure 3.

The olivine–orthopyroxene control plane during partial melting of spinel lherzolite compositions is especially evident on olivine projections. It is drawn through the lherzolite average composition and residual harzburgite average composition and is controlled by the solid solution compositions of olivine and enstatite. (The scatter of some harzburgite compositions is a result of the projection method.) The composition of the final liquid produced during partial melting of the spinel lherzolite must also lie on this control line, and since some small amount of clinopyroxene remains in the harzburgites, must also be on the olivine orthopyroxene, clinopyroxene cotectic, i.e. point A, at 18–20 kbar. This pressure is chosen since it is the minimum pressure at which the harzburgite mineralogy equilibrated. If continued melting had taken place at any lower pressure, then presumably the harzburgite would register lower equilibrium pressures. The primary melt, i.e. composition A, has komatiitic affinities (Cawthorn & Strong 1975).

Basalt compositions project close to low-pressure plagioclase–clinopyroxene cotectics. The spread of points is mainly resultant from the projection method. Some dikes, all gabbros and the cumulate rocks plot away from the cotectics indicating cumulate enrichment in plagioclase and/or olivine, but dominantly along low-pressure control planes defined by these minerals. No rocks appear to have crystallized at the low-pressure eutectic in agreement with the absence of orthopyroxene in the basalts. The basalt compositions straddle the low-pressure thermal divide between alkaline and olivine–tholeiite fields indicating the transitional nature of the series.

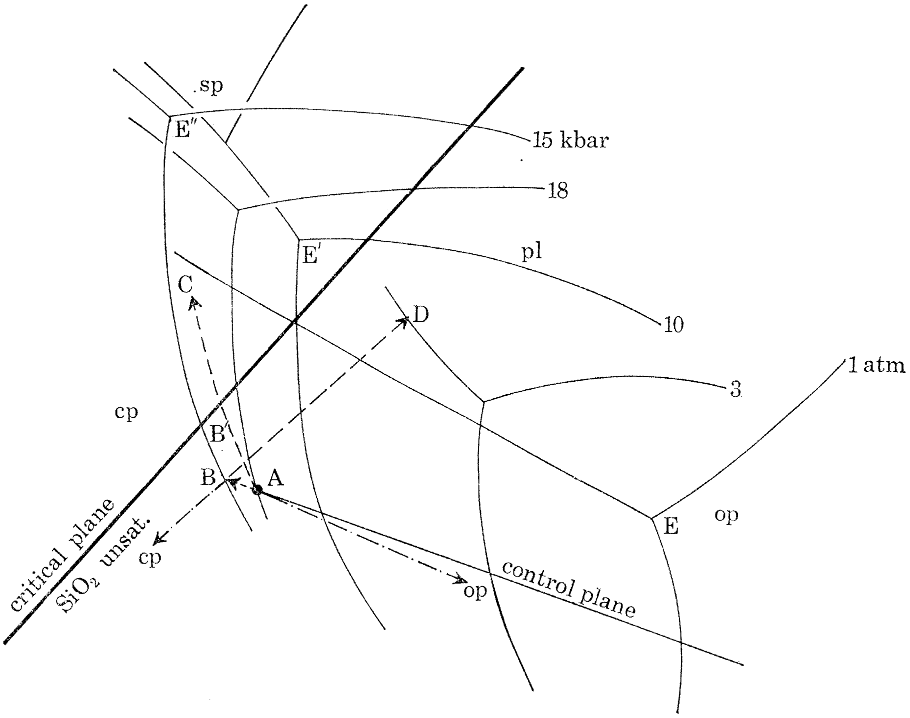

FIGURE 7. Enlargement of figure 6*a*′ showing derivation of partial-melt liquids. Symbols as for figure 6.

The convective rise of mantle material into pressure régimes of less than 20 kbar caused crystallization of the melt that had been produced at greater pressures. This convective rise can be viewed as a diapiric upwelling of a solid/liquid mixture. The nature of the crystallization depended upon the rate of uprise of the diapir, thence the rate of pressure reduction, and the rate of crystallization of the liquid. With reduction in pressure the cotectics move as indicated in the projections. If reduction in pressure took place slowly, allowing the rate of crystallization to maintain a liquid path along the olivine–clinopyroxene–orthopyroxene cotectic, then between pressures of 18–20 kb and 15 kb, the liquid path would be along a line such as AB′C (figure 7) such that an initially tholeiitic liquid (primary melt A) would become eventually alkaline as it crossed the low pressure thermal divide. Such would be the case if continued fractionation took place along this cotectic at any pressures between 20 and 10 kbar (figure 7). It is estimated from the projections that 25 % crystallization of melt A at such pressures would cause the transition into the alkali field. Below 10 kbar continued fractionation would give a residual liquid once more in the tholeiitic field.

If pressure reduction took place relatively quickly, such that cotectic crystallization was not maintained, then olivine alone would crystallize out initially. The olivine would be joined by orthopyroxene if the olivine–orthopyroxene cotectic was intersected (figure 7). Co-crystallization of olivine and orthopyroxene would be possible down to pressures of 15 kbar (i.e. path A⟶B) when movement of the cotectic to lower pressure positions would leave liquid B in the olivine–clinopyroxene field. Orthopyroxene would then be replaced by clinopyroxene in the crystallization sequence causing liquid composition to move from B towards D. At any stage during such a crystallization sequence, the residual liquid is tholeiitic in composition.

At lower pressures, clinopyroxene and olivine would be joined by plagioclase as fractionating phases (e.g. position D, figure 7) and the sequence as explained in the one atmosphere basalt tetrahedron derived.

Clearly, if cotectic crystallization had taken place during slow ascent of the diapir, then orthopyroxene would disappear as a crystallizing phase when the path of the eutectic E″–E′–E passed across the path of liquid descent and intersection of the clinopyroxene–plagioclase cotectic was possible.

Any combination of these crystallization paths can occur according to the changing rate of pressure reduction, or the crystallization rate; e.g. a period of cotectic crystallization might be followed by an increase in rate of upward movement of the diapir causing the crystallization of only one phase, or vice versa.

The various proportions of the three phases, olivine, orthopyroxene, clinopyroxene, that are obtained by crystallization under polybaric high-pressure conditions explains the various modal proportions of these minerals comprising the veins in the tectonites. These veins represent early fractionation products of the primary melt as it crystallizes at pressures below 20 kbar. Dunite veins represent non-cotectic crystallization of olivine; enstatolite, crystallization of orthopyroxene and olivine along a line such as AB in figure 6 and pyroxenite veins, fractionation of dominantly clinopyroxene, e.g. along line BD in figure 7.

Starting with a primary melt of composition A, it is possible to derive both nepheline normative alkaline liquids (e.g. composition C) and tholeiitic liquids. Release of these liquids and their subsequent eruption may explain the transitional nature of the mafic series of the Bay of Islands Complex. Release of the primary melt, or a very slightly fractionated derivative of it would give rise to picritic lavas such as described by Upadhyay (1973) and Norman (1974) found associated with other Newfoundland ophiolite suites. Control of the time and depth of release of these liquids would be dominantly controlled by the amount of liquid present and the rate of ascent of the diapir.

More significantly perhaps, the petrogenetic scheme presented may explain the increase in alkalinity of volcanics away from ridge axes in present oceans (McBirney & Gass 1967; Strong 1974) and support O'Hara's (1973) view that this is due to increased opportunity for crystal fractionation. Liquids erupted in off-ridge environments have progressively thicker, colder and perhaps more solid lithosphere to pass through before eruption and therefore greater opportunity for high-pressure cotectic crystallization as their upward progress is physically impeded by the lithosphere plate. Material ascending directly beneath the ridge-axis encounters virtually no lithosphere and will be erupted after little high-pressure cotectic crystallization. Material ascending just off-axis might travel laterally and undergo further fractionation before eruption.

The petrogenetic model also indicates that at pressures of less than 20 kbar, no melting has taken place in the upwelling diapir. If applicable on a wider scale, this is contrary to the views of many workers which state that release of pressure and accompanied adiabatic temperature increase lead to melting at any depth.

Degrees of partial melting

An 'extraction programme' has been used to calculate the degree of partial melting of the spinel lherzolites that had to take place to produce a melt of composition A. In this case an average Bay of Islands spinel lherzolite composition was used as starting material and the most residual harzburgite composition subtracted. The resultant liquid must plot at point A on all c.m.a.s. projections and should have an $Mg/(Mg + Fe^{2+})$ ratio consistent with it being in equilibrium with the harzburgite, i.e. 0.77–0.78. It has been found that partial melting of 23 % produced a liquid composition fulfilling these requirements. This liquid composition is noticeably similar to the picritic basalts of Clarke (1970) who suggests their formation as a primary magma by partial melting of garnet peridotite at 30 kbar.

Degrees of fractional crystallization

The original magma, of composition A, produced in equilibrium with the harzburgites, must have had a $Mg/(Mg + Fe^{2+})$ ratio in equilibrium with olivine Fo_{92} (olivine composition of harzburgites). Experimentally determined partition coefficients between olivine and liquid suggest that this ratio should be 0.77–0.78 (Roeder & Emslie 1970). Clearly, therefore, the basalts do not represent this original composition but are crystallized from a later liquid after fractionation of a certain proportion of cumulates. These proportions can be estimated approximately as follows:

(i) Assuming that only olivine is removed as cumulative material and that all olivine removed has an average content Fo_{87} (Fo_{90} is removed during early stages of fractionation and much more fayalitic compositions later), then, working in percentages by mass,

$$MgO_{\text{initial magma}} = MgO_{\text{basalt}} + x\, MgO_{Fo_{87}};$$
$$FeO_{\text{initial magma}} = FeO_{\text{basalt}} + x\, FeO_{Fo_{87}};$$

where x = percentage of fractionated Fo_{87} and $\{Mg/(Mg + Fe^{2+})\}_{\text{initial magma}} = 0.78$.

MgO in Fo_{87} = 47.5 %;
FeO in Fo_{87} = 12.5 %;
MgO in basalt = 8 %;
FeO in basalt = 7 %.

Therefore for basalt + 20 % Fo_{87}:

$$MgO = 17.5\,\%; \quad FeO = 9.5\,\%; \quad Mg/(Mg + Fe^{2+}) = 0.768.$$

Therefore for basalt + 25 % Fo_{87}:

$$MgO = 19.8\,\%; \quad FeO = 10.1\,\%; \quad Mg/(Mg + Fe^{2+}) = 0.779.$$

This calculation suggests that an initial magma produced in equilibrium with Fo_{92} would require fractionation of 20–25 % olivine (Fo_{87}) to produce the basalt composition. Although no accurate estimate of the amount of vein material cutting the harzburgites is available, the total amount of fractionated olivine from field evidence does not appear to exceed 10 % of the total mafic sequence. It is therefore necessary to take into account fractionation of gabbroic composition in addition to olivines. The presence of cumulate gabbros in the field support this reasoning.

(ii) Assuming 10 % fractionation of olivine Fo_{87}:

$$\mathrm{MgO} = 12.71\,\%;\quad \mathrm{FeO} = 8.25\,\%;\quad \mathrm{Mg/(Mg+Fe^{2+})} = 0.735.$$

Now with further fractionation of gabbroic material (*including critical zone rocks*),

$$\mathrm{MgO_{gabbro}} \approx 12.0\,\%$$
$$\mathrm{FeO_{gabbro}} \approx 4.1\,\%.$$

Then basalt + 10 % Fo_{87} + 60 % gabbro gives

$$\mathrm{MgO} = 19.91\,\%;\quad \mathrm{FeO} = 10.65\,\%;\quad \mathrm{Mg/(Mg+Fe^{2+})} = 0.771.$$

This calculation assumes 10 % olivine fractionation (orthopyroxene and clinopyroxene have also been fractionated, especially in the veins, but since their Mg/Fe ratios are similar to olivine and since they crystallize in relatively small amounts, they are neglected for this approximation), followed by 60 % gabbro fractionation, leaving 30 % liquid. That is, the primary magma, A, should give rise to

olivine (pyroxene) cumulates including veins:	10 %;
gabbro cumulates + some dikes + c.z.:	60 %;
basalts + dikes:	30 %.

Such proportions do not seem inappropriate for the Bay of Islands Complex, from the field evidence. The proportion of ultramafic cumulates might be increased and gabbro proportion decreased when an accurate estimate of amounts of vein material are made.

At this point it is worth noting that similar arguments applied to contemporary oceanic tholeiites indicate that they do not have an $\mathrm{Mg/(Mg+Fe^{2+})}$ ratio in equilibrium with the mantle composition from which many authors suggest they are directly derived. Calculations made, for example, on the analyses presented by Dimitriev (1974), who explicitly states that abyssal tholeiites are primary and undifferentiated melts of the upper mantle, indicate that a 20 % fractionation of olivine (Fo_{87}) is required to produce his basalt composition from a primary melt in equilibrium with upper mantle lithologies. Such arguments apply to almost all oceanic tholeiites described in the literature, and substantiate the view that oceanic tholeiites are a result of crystallization from a differentiated magma.

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Discussion

Y. Bottinga (*Harvard University, Hoffman Laboratory,* 20 *Oxford Street, Cambridge, MA* 02138, *U.S.A.*). In the petrogenetic model described, to explain basalt eruption at the ridge axis and chemical variations of this basalt, 23 % partial melting occurs at 18 kbar pressure and a temperature of 1100 °C. In the light of the evidence that the water activity and the water concentration in the upper mantle at a spreading centre are very small, how is it possible that the tholeiite erupting at the ridge axis has a liquidus temperature of about 1200 °C or more, while it originates in your model as picritic tholeiite at 1100 °C and 18 kbar? All thermal calculations agree on the fact that in comparison with energy transfer by convection, the energy transport by conduction is insignificant in the upper mantle as long as the vertical convection velocities are equal to or greater than 1 cm/a. Hence it would be expected that in the absence of crystallization in your partial melt, the liquid would erupt at about 1080 °C in the author's model, or more than 100 °C below the tholeiite liquidus. In case crystallization occurs, latent heat of fusion will be liberated, but the temperature cannot really be much higher than 1100 °C, because crystallization occurs only when the temperature decreases. Moreover the temperature tends to be buffered by the surrounding solid which in the author's model should be also about 1100 °C at 18 kbar.

J. Malpas. I agree with Professor Bottinga's remarks and, as pointed out in my paper, I consider that the peridotitic rocks from which the temperatures and pressures are obtained have undergone some subsolidus re-equilibration as corroborated by textural evidence. Nevertheless, this can only suggest that the picritic liquid A was produced at slightly higher temperatures and possibly pressures (or depths), and makes little difference to the overall petrogenetic model in which, in fact, temperature need not be directly considered. However, the 60 km restriction on melting now becomes a *minimum* depth at which partial melting to produce liquid A is replaced by its crystallization.

T. J. G. Francis (*Institute of Oceanographic Sciences, Blacknest, Brimpton, Reading RG*7 4*RS, U.K.*). I am interested in the extent to which seawater penetrates the oceanic crust. I should like to ask Professor Malpas what he can say from the nature of the rocks in the Bay of Islands ophiolite, or in other ophiolite suites, about the extent to which seawater penetrated these rocks soon after the time of their formation.

J. Malpas. There are a number of lines of evidence that seawater may have penetrated to some depths in ophiolite suites. Evidence from oxygen isotope studies is beginning to indicate in some cases (e.g. Liguria) that seawater has been effective in the early sub-seafloor metamorphism of pillow-lavas, diabase dikes and possibly gabbroic portions of ophiolite suites. Additionally, the dike–breccia textures of the Newfoundland ophiolites, now recorded from a number of other localities, has been interpreted as a result of seawater interaction with cooling diabase (Williams & Malpas 1972), suggesting that seawater might have penetrated the oceanic crust to depths of 2–3 km. If seawater had also penetrated to depths of crystallizing gabbroic melts, then it is possible that its presence caused the crystallization of the pegmatitic hornblende–gabbros found towards the top of the gabbroic sequence. This might be checked by isotope studies.

All evidence suggests possibly seawater penetration to depths of 2–3 km soon after formation of the ophiolites and compares favourably with models of present hydrothermal systems on, for example, the mid-Atlantic ridge (see Scott *et al.* 1974).

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