Mantle Xenoliths
Edited by P. H. Nixon

## 24

# Ontong Java Plateau: deep-seated xenoliths from thick oceanic lithosphere

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There are abundant and varied mantle xenolith suites in 34 m.y. alnöite intrusions in the northern part of the island of Malaita, Solomon Islands, in the southwestern Pacific. The xenoliths are deep-seated and include garnet-bearing types similar to those in kimberlites and are a reflection of an unusual tectonic setting, an understanding of which is necessary for the interpretation of the xenolith data.

## Tectonic Setting

Malaita lies on the southwestern margin of the Ontong Java Plateau (OJP) occupying a 1600 × 800 km area which has an unusually thick crust for an ocean region. This is over 40 km thick in the northern part (Furumoto *et al.*, 1976). The thinner (ca. 25 km) southwestern margin of the OJP is now exposed as a result of folding with possible obduction following recent (< 10 m.y. ago) collision of the westerly drifting plate with the proto-Solomon Island chain (Coleman and Kroenke, 1981).

The origin of the OJP is uncertain. It may, together with other rises in the Pacific, have resulted from massive volcanism in the Cretaceous during unusually slow spreading at a ridge or triple junction (Kroenke, 1974; Hilde *et al.*, 1977; Taylor, 1978). On the other hand, it may be much older. Nur and Ben-Avraham (1982) regard the OJP as a submerged fragment detached from a previous 'Pacifica domain', possibly a continent, as long ago as the Permian. Similar fragments are postulated to have been consumed at active plate margins where they represent 'allochthonous' or 'suspect' terranes.

Drilling in the Nauru Basin (DSDP site 462) adjacent to the OJP (Figs. 160 and 161) has yielded the following probable sequence of volcanic events stretching back almost to the inferred time of initial fragmentation of the Pacific (Schlanger *et al.*, 1981):

1. Formation of the Pacific Plate 155 m.y. ago at a ridge crest
2. Mid-plate lava flows (ca. 120 m.y.)
3. Dolerite sill emplacement (ca. 110 m.y.) } Cretaceous
4. Volcanic episode from ca. 95 to 70 m.y.

The Cretaceous volcanism not only embraced the OJP but most of the central Pacific (Fig. 160). There was probably a contemporaneous upward bulging of the mantle (Menard's (1964) 'Darwin Rise') due to lithospheric heating which resulted in the end-Mesozoic elevation in sea level.

On the island of Malaita this volcanism is represented by the 'older' tholeiitic lavas (probably early Cretaceous) and the 'younger' alkali basalts of upper Cretaceous to Eocene age (Hughes and Turner, 1977). Offshore on the OJP, early Cretaceous lavas are at the base of the pelagic sedimentary succession at site 298 (Fig. 161) and volcanogenic sediments at higher stratigraphical levels (ca. 95–85 m.y.) at site 288 (Andrews and Packham, 1975).

The emplacement of the alnöites is a relatively recent event (zircon age of 34 m.y.; Davis, 1977) which we regard as taking place after stabilisation of the plate following the Cretaceous volcanism. As with similar alkaline volcanism elsewhere, the alnöites are a manifestation of intra-plate activity.

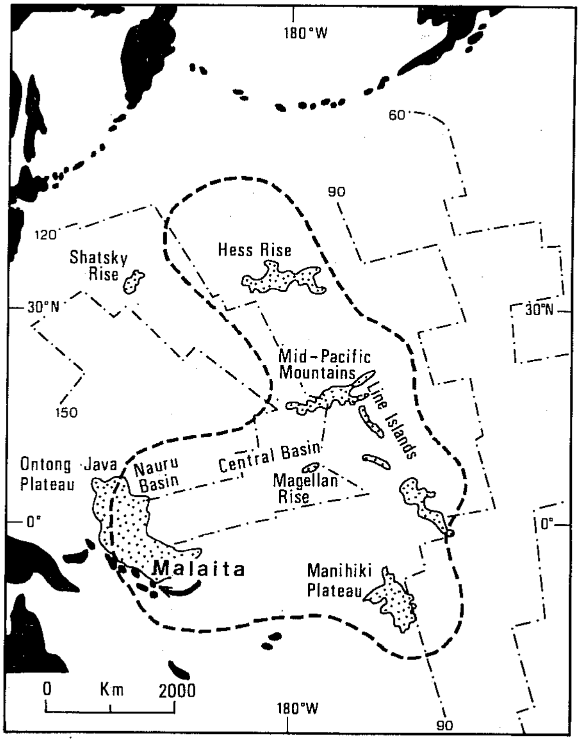

Figure 160. Area of the western Pacific (within the heavy dashed line) characterised by excessive volcanism during the interval 120–90 m.y. according to Watts *et al.* (1980). Crustal isochrons are indicated (Schlanger *et al.*, 1981). Malaita lies in the southern part of the Ontong Java Plateau and contains Cretaceous volcanism and later (34 m.y.) alnöite intrusive rocks with mantle xenoliths. *Reproduced by permission of Elsevier Scientific Publishers*

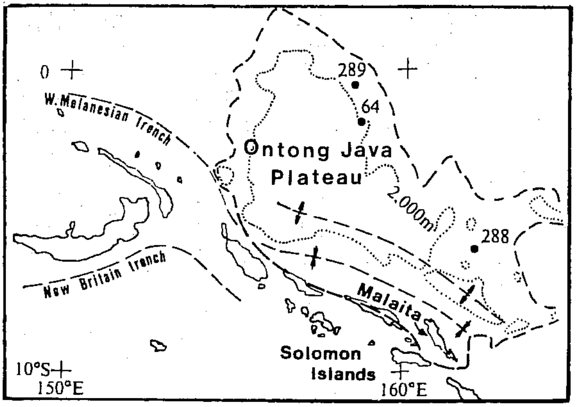

Figure 161. The Ontong Java Plateau delimited by the 2500 m isobath and showing the sites of Deep Sea Drilling (Nixon, 1980). Malaita forms part of the northwest trending folds close to the collision zone with the main part of the Solomon Islands (Coleman and Kroenke, 1981)

If allowance is made for plate movement it is estimated that at the time of alnöite intrusion the OJP was situated over 1000 km to the east in the central Pacific (Fig. 138; Nixon and Coleman, 1978).

### The xenoliths

The alnöites in northern Malaita intrude a sequence of marine lavas and pyroclastic rocks with overlying upper Cretaceous limestones and siltstones (Coleman, 1968) which can be equated with the DSDP drill cores of the OJP (Kroenke, 1972). Several intrusive centres are known and many more suspected from alluvial mineral indications (Allen and Deans, 1965; authors' unpublished data). Most of our samples are from two centres: Babaru'u and Kwaikwai. The host rocks include 'alphanitic' and calcitic fragmental varieties (Plate 15A). The alnöite consists of olivine ($Fo_{85}$), diopside-salite, natro-melilite, Ti-phlogopite, perovskite, spinel, and accessory nepheline, melanite and apatite (Nixon *et al.*, 1980).

Crustal xenoliths are similar to exposed crustal rocks and no possible deep-seated types, e.g. garnet granulite, which might suggest a continental structure, have been found.

## Peridotite Suite

These xenoliths, although common, are invariably serpentinised and friable, which gives problems

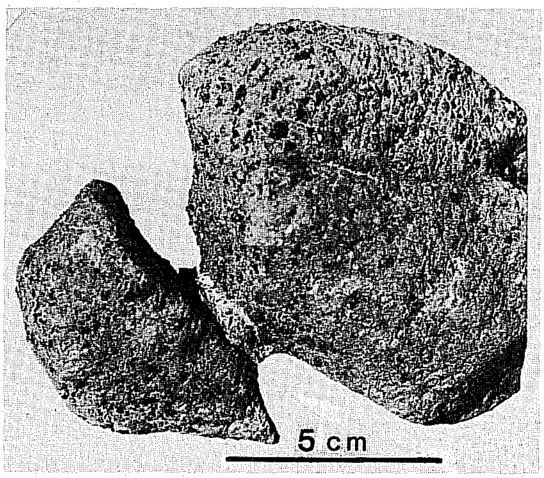

Figure 162. Spinel lherzolite xenolith sculpted along a plane of weakness (calcite vein) during fluidised eruption of the alnöite, Babaru'u, Malaita

when sampling and later when determining modes. The largest samples are about 30 cm across. They range from wholly spinel-bearing (Fig. 162) to garnet-bearing; amphibole is found throughout the range but phlogopite is surprisingly rare. Another feature is that clinopyroxene is of similar, and often greater abundance than orthopyroxene. Most peridotites are therefore lherzolites. The descriptions given below supplement those of Nixon and Boyd (1979). In particular, xenoliths from the previously unreported intrusion at Kwaikwai have a much higher proportion of spinel lherzolites, suggestive of shallower mantle sampling by the intrusion than at Babaru'u.

## Spinel lherzolites

Forty-five small specimens, mostly from Kwaikwai, average about 70% serpentinised olivine; in most cases clinopyroxene exceeds orthopyroxene (assuming no differential serpentinisation). Spinel is an accessory mineral and amphibole is present in seven specimens (Neal, 1985).

Textures are mainly coarse (2–5 mm grains) but brittle fracturing of pyroxenes may be present. Recrystallisation to granuloblastic textures occurs. Interstitial 'holly-leaf' spinels may be mantled by replacing secondary amphibole; clinopyroxene, but not orthopyroxene, is the other reactant.

*Olivines* from three less altered xenoliths gave

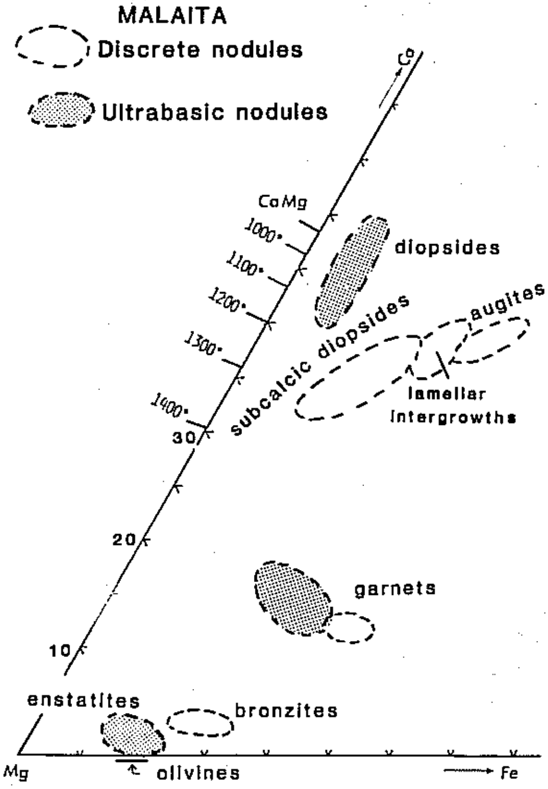

Figure 163. Composition fields of minerals from discrete nodules (megacrysts) and ultrabasic (peridotite) xenoliths, Babaru'u and Kwaikwai intrusions, Malaita. *After Nixon and Boyd (1979). Additional data courtesy of F. R. Boyd (cpx-ilm field) and Neal (1985)*

TABLE 97. Spinel lherzolite mineral compositions, Malaita

| | ol | cpx | cpx, 34 xenoliths | opx | opx, 32 xenoliths | amph | sp | sp, 30 xenoliths |
|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 40.60 | 52.85 | 51.67 –53.22 | 55.36 | 54.42 –55.79 | 44.06 | — | — |
| $TiO_2$ | 0.01 | 0.46 | 0.35 – 0.80 | 0.11 | 0.07 – 0.17 | 0.92 | 0.14 | 0.08 – 0.18 |
| $Al_2O_3$ | 0.57 | 5.80 | 5.12 – 6.86 | 4.69 | 3.96 – 4.91 | 15.33 | 57.16 | 56.18 –59.20 |
| $Cr_2O_3$ | 0.02 | 0.64 | 0.60 – 0.96 | 0.33 | 0.25 – 0.41 | 0.84 | 9.27 | 8.16 –11.29 |
| $Fe_2O_3$ | | | | | | | 2.39 | 0.91 – 3.24 |
| FeO | 9.67 | 2.62 | 2.14 – 2.72 | 6.55 | 6.01 – 6.76 | 3.93 | 10.17 | 8.85 –10.91 |
| MnO | 0.19 | 0.09 | — | 0.21 | 0.12 – 0.20 | 0.06 | 0.12 | 0.00 – 0.17 |
| MgO | 48.48 | 15.02 | 14.20 –15.17 | 32.51 | 32.15 –33.31 | 17.82 | 20.13 | 19.08 –20.74 |
| CaO | | 21.28 | 20.27 –22.00 | 0.45 | 0.41–0.76 | 10.79 | — | — |
| $Na_2O$ | | 1.39 | 1.24 – 2.18 | — | — | 3.76 | — | — |
| NiO | 0.40 | | | | | | 0.42 | 0.24 – 0.48 |
| | 99.94 | 100.15 | | 100.21 | | 97.51 | | |
| $Mg/(Mg+Fe^{2+})$ | 0.897 | 0.911 | 0.924– 0.908 | 0.898 | 0.905– 0.898 | 0.890 | 0.799 | 0.793– 0.772 |
| $Ca/(Ca+Mg)$ | — | 0.505 | 0.500– 0.510 | 0.010 | 0.009– 0.016 | 0.303 | — | — |
| $Cr/(Cr+Al)$ | — | 0.068 | 0.063– 0.084 | 0.045 | 0.035– 0.125 | 0.036 | 0.094 | 0.089– 0.113 |

Single analyses are from xenolith CRN 205, Babaru'u with exception of olivine which was completely serpentinised; the olivine shown here is from xenolith CRN 214. The ranges shown exclude some aberrant values resulting from metasomatism and other factors mentioned in the text.

compositions close to $Fo_{90}$ (Table 97; Fig. 163). The *clinopyroxenes* are calcic Cr-diopsides with low equilibration temperatures. They are aluminous and magnesian and have a restricted range in *mg*, apart from occasional deviations probably due to metasomatism; see below. *Orthopyroxenes* are close to $En_{90}$ with significant aluminium and low calcium compatible with the low-temperature clinopyroxenes. *Amphibole* is of pargasite composition but has variable *mg* (83 to 89%) and $K_2O$ ($< 0.05$ to 0.76). The appearance of this mineral is equated with the metasomatic introduction of potassium, titanium, and sodium (Neal, 1985). The spinels (pleonastes) are rich in magnesium and aluminium and unlike those found in most kimberlitic xenoliths.

## Spinel garnet lherzolites

The development of garnet coronas around spinels is the most striking feature of these xenoliths (Fig. 164). Neal and Nixon (1985) have traced the evolution of these textures from an original Al-spinel lherzolite to a Cr-spinel garnet lherzolite resulting from decreased temperature/increased pressure followed by a late partial reversal due to reheating/decompression during alnöite eruption. Figure 165 illustrates this sequence and formation of secondary spinel, and slivers of clinopyroxene and amphibole during this process

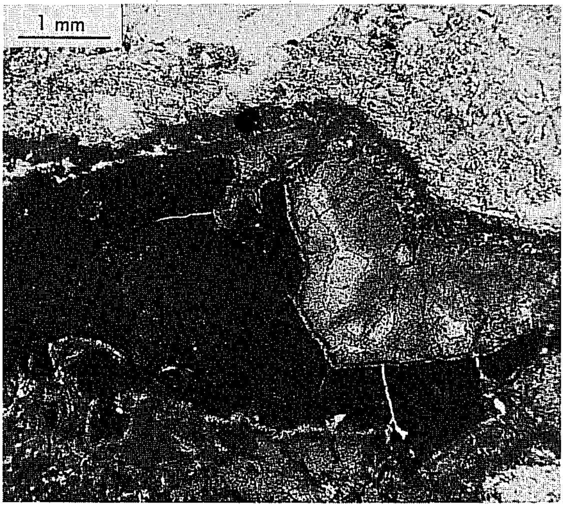

Figure 164. Thin section showing coarse spinel (dark grey) with kelyphitised corona (medium grey) in which relict patches of garnet are still visible (pale areas). Garnet spinel lherzolite PHN 3567, Babaru'u, Malaita. Ordinary light

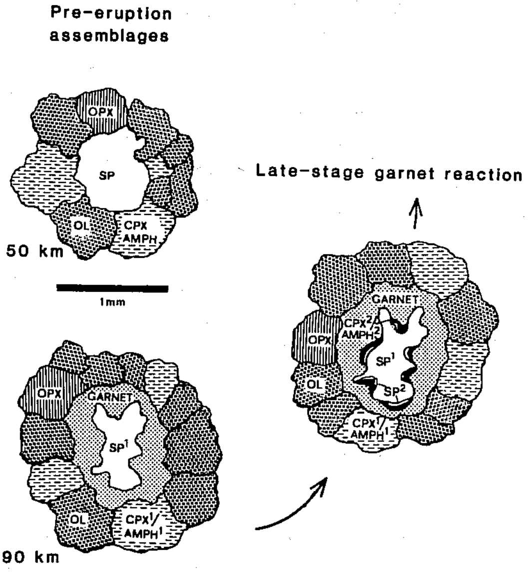

Figure 165. Sketch of aluminous spinel assemblages in shallow lherzolite xenoliths and spinel (more Cr-rich) garnet assemblages at greater depths. A late heating and/or decompression (eruptive) event produces slivers of secondary aluminous spinel, and subcalcic clinopyroxene and amphibole (modified from Neal and Nixon, 1985)

Where the first reaction has proceeded to completion there is no spinel remaining, but more often increased residual chromium enrichment stabilises the spinel so that it co-exists with garnet (MacGregor, 1970). Specimens in this group include those where the former presence of garnet is represented by kelyphite alteration. Symplectitic intergrowths of spinel and calcic clinopyroxene may also be indicative of garnet breakdown. Lamellar two-pyroxene exsolution features are common in this group

Compared with the spinel lherzolites, olivine may be modally less abundant; clinopyroxene is again conspicuous and no harzburgites were recorded. Amphibole occurs in more than half the specimens; in one specimen it poikilitically encloses garnet grains in apparent equilibrium with this mineral. There is no spinel in this specimen.

The *olivines* have a forsterite range of 89–92 (mean 90) (see Nixon and Boyd, 1979). The *clinopyroxene* range overlaps that of the spinel lherzolites but extends to lower $Ca/(Ca + Mg)$, i.e. higher equilibration temperatures, rather lower $Al_2O_3$ wt%, and includes more depleted varieties in terms of $Cr/(Cr + Al)$ and Mg/

TABLE 98. Compositions of primary and secondary minerals in spinel garnet lherzolites, Malaita

| | PHN 4069 | | PHN 4009 | | | | | |
|---|---|---|---|---|---|---|---|---|
| | cpx[1] | cpx[2] | opx | sp[1] | sp[2] | gt | amph[1] | amph[2] |
| $SiO_2$ | 53.00 | 51.47 | 55.15 | — | — | 41.88 | 45.78 | 45.03 |
| $TiO_2$ | 0.26 | 0.30 | 0.13 | 0.32 | 0.16 | 0.07 | 0.95 | 0.25 |
| $Al_2O_3$ | 2.87 | 6.00 | 3.94 | 44.46 | 57.73 | 22.98 | 12.09 | 14.80 |
| $Cr_2O_3$ | 0.64 | 0.81 | 0.28 | 21.11 | 8.91 | 0.61 | 0.75 | 0.59 |
| $Fe_2O_3$ | — | — | — | 3.30 | 1.95 | — | — | — |
| FeO | 2.68 | 4.16 | 6.63 | 11.56 | 9.87 | 9.12 | 4.44 | 5.73 |
| MnO | 0.05 | 0.30 | 0.16 | 0.05 | 0.07 | 0.53 | 0.75 | 0.60 |
| MgO | 16.36 | 16.11 | 32.95 | 17.66 | 20.67 | 19.63 | 18.88 | 18.58 |
| CaO | 22.63 | 19.70 | 0.41 | — | — | 4.84 | 10.14 | 8.73 |
| $Na_2O$ | 0.77 | 1.14 | 0.28 | — | — | — | 4.21 | 3.94 |
| $K_2O$ | — | — | — | (NiO, 0.33 | 0.47) | — | 0.67 | 0.25 |
| Total | 99.26 | 99.99 | 99.93 | 98.79 | 99.83 | 99.66 | 98.66 | 98.50 |

Range of compositions of primary minerals in spinel garnet lherzolites, Malaita

| | cpx (11) | opx (12) | sp (12) | gt (10) | amph (7) |
|---|---|---|---|---|---|
| $TiO_2$wt% | 0.3 –0.7 | 0.1 –0.2 | 0.1 – 1.5 | 0.0 – 0.15 | 0.3 – 3.0 |
| $Al_2O_3$wt% | 4.7 –6.4 | 1.9 –4.3 | 25.0 –57.0 | 22.0 –24.0 | 12.0 –15.5 |
| $Mg/(Mg + Fe^{2+})$ | 0.90–0.93 | 0.89 –0.91 | 0.68– 0.81 | 0.77– 0.86 | 0.80– 0.88 |
| $Ca/(Ca + Mg)$ | 0.37–0.51 | 0.009–0.038 | — | 0.15– 0.19 | 0.26– 0.47 |
| $Cr/(Cr + Al)$ | 0.06–0.13 | 0.03 –0.15 | 0.09– 0.54 | 0.01– 0.04 | 0.04– 0.07 |

Superscripts 1 and 2 refer to primary and secondary minerals respectively.

Nos. of xenoliths anlysed shown in brackets; aberrant secondary minerals excluded; also those from a few garnet lherzolites with no spinel. Analyses from Neal (1985).

Note that 'primary' means the earliest generation seen in the rock; thus primary spinel coexisting with garnet differs from that in the original garnet-free rock. In xenolith PHN 4009 an Al-spinel enclosed by orthopyroxene is deemed to be a relict of the original spinel lherzolite assemblage by Neal and Nixon (1985; their sp[1]).

(Mg + Fe) ratios (Table 98). This table includes a xenolith CRN 209 with a 'hot' clinopyroxene having Ca/(Ca + Mg) = 0.367 (and opx = 0.038). The orthopyroxenes are of similar composition to those in the spinel peridotites with the exception of CRN 209. The *amphiboles* are pargasites as in the spinel lherzolites and, although of possible early metasomatic origin, they are texturally in equilibrium with the other 'primary' minerals. Secondary slivers of amphibole and clinopyroxene are described below. The *spinel* core compositions are significantly enriched in titanium compared with those in the spinel lherzolites. Compositions overlap somewhat but there is a marked trend to chromite compositions (max. $Cr_2O_3 = 46$ wt%). The *garnets* are pyropes (mostly within the range of 67–71%) with low titanium and lower chromium (0.77–1.32 wt%) than in most comparable xenoliths from elsewhere.

In three *garnet only* lherzolites (i.e. spinel is absent, having presumably been consumed by the garnet-forming reaction) the garnet contains significantly higher $Cr_2O_3$ wt%, e.g. 1.78 (PHN 4034) and 4.61 (PHN 3538; Nixon and Boyd, 1979). In addition, the high-temperature (and therefore probably deeper-seated) spinel-bearing xenolith CRN 209 also has a high-chromium pyrope (4.90 wt% $Cr_2O_3$). The minerals in all these rocks tend to be depleted in iron, aluminium, sodium, and titanium, suggesting depletion of the basal lithosphere of the OJP (Nixon and Boyd, 1979). It is worth drawing an analogy with the occurrence of ultradepleted peridotites postulated at the base of the continental cratonic lithosphere (Boyd and Gurney, 1982).

### Late thermal event

Slivers of secondary clinopyroxene and amphibole have formed along the contact of spinel and its garnet envelope in several xenoliths. Neal and Nixon (1985) suggest that they have formed as a result of a partial reversal of postulated reactions which led to the formation of garnet in the first instance:

$$\mathrm{Sp + En + Di^{*} \rightleftharpoons Fo + Gross + Py + (H_2O)}$$

*or pargasite

The garnet-forming prograde reaction was at the expense of aluminous spinel, which progressively moved towards chromite composition and hence became stable under higher PT conditions. An equilibrium was usually reached but garnet lherzolites in which there is no spinel (see above) could indicate that the reaction has gone to completion. Where spinels survive, the reversal produces thin ( ~ 20 $\mu$m) Al- and Mg-rich rims (Table 98). No secondary orthopyroxene is seen. This component may be dissolved in the secondary clinopyroxene or amphibole, judging from their slightly subcalcic natures (Table 98), i.e. lower Ca/(Ca + Mg). It is inferred that the reversal represents a heating and/or decompression effect due to contact with the erupting alnöite (Fig. 165).

## Discrete Nodule (Megacryst) Suite

The alnöite intrusions of northern Malaita contain unrivalled suites of garnet, clinopyroxene, orthopyroxene, clinopyroxene-ilmenite intergrowths, ilmenite, and zircon. In rocks not subjected to mining excavation it is remarkable that specimens of over 1 kg can be picked up on the surface (Plate 15B, Figs. 166 and 167).

The main features have been described by Nixon and Boyd (1979). The garnets are rounded and fractured with sometimes sinusoidal trails of altered alnöitic (?) material as if caught up during kneading of a garnet having rheologically weak consistency. It is uncertain whether similar deformation has contribution to some granular ilmenite-clinopyroxene intergrowths (Fig. 168) or irregular distribution of sulphides in pyroxene (Fig. 169). In the latter case the concentrations of *sulphides* occur as oriented blebs within their particular domain and would appear to be crystal lattice controlled, subsolidus exsolution features. Alternatively, they may have been included during the growth of the megacryst (cf. Irving, 1974c).

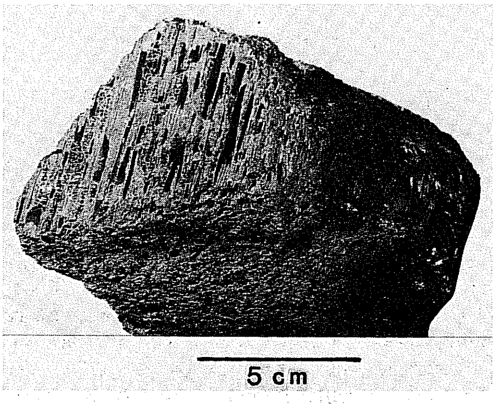

Figure 166. Clinopyroxene-ilmenite lamellar intergrowth weighing about 1 kg, PHN 3865, Babaru'u, Malaita

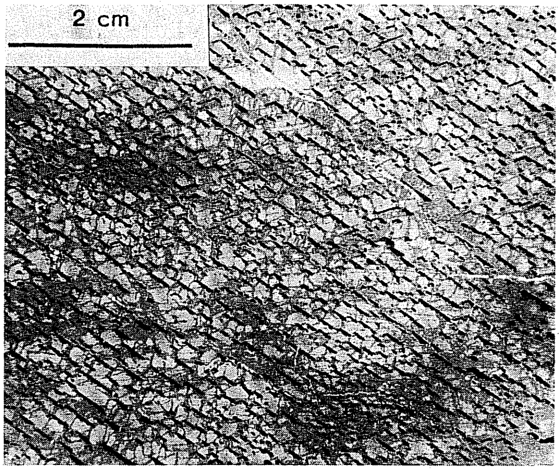

Figure 167. Large thin section of part of clinopyroxene-ilmenite lamellar intergrowth (see Fig. 166). Ordinary light

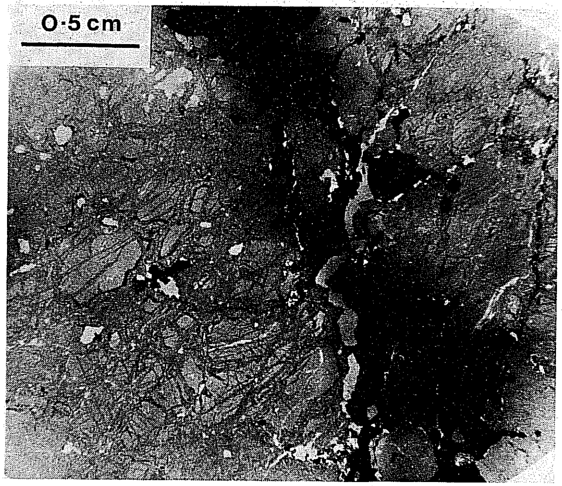

Figure 168. Thin section of clinopyroxene-ilmenite granular intergrowth PHN 3978, Babaru'u, Malaita. Ordinary light

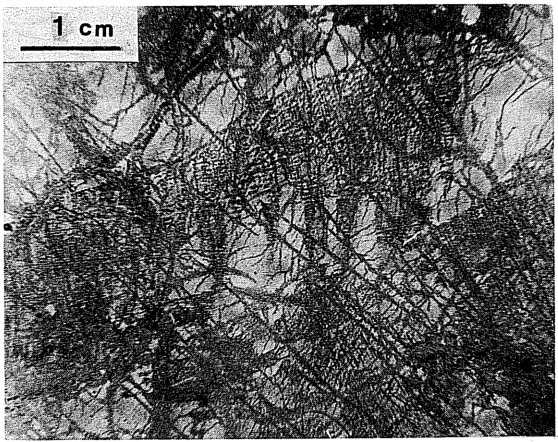

Figure 169. Thin section showing zones of sulphide blebs in clinopyroxene discrete nodule. Within each zone the blebs are well oriented but not with respect to neighbouring zones. PHN 3883, Babaru'u, Malaita. Ordinary light

In subcalcic diopside 3897B there are cloudy zones, possibly due to partial melting and a rounded sulphide segregation consisting predominantly of a pyrrhotite-pentlandite intergrowth with an inclusion of K-Fe-Ni-(Cu)-sulphide and associated glass with up to 10 wt% $K_2O$ (J. D. Pasteris, personal communication). This is interpreted by Pasteris to represent the introduction of a K-rich fluid with consequent incipient melting especially around the sulphide blebs.

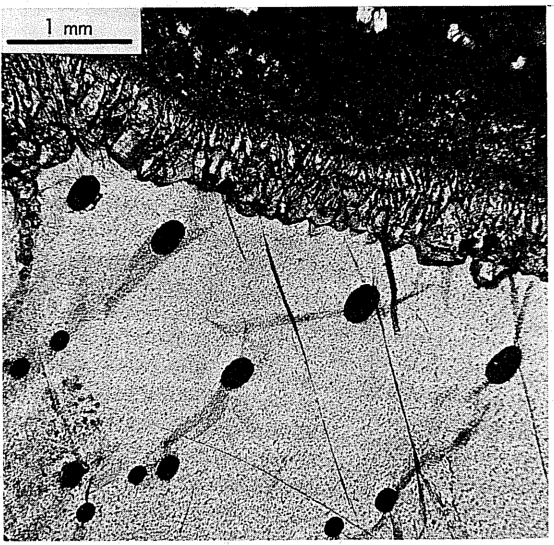

Figure 170. Thin section of bronzite megacryst with sulphide blebs and complex reaction rim against alnöite. PHN 3542, Babaru'u, Malaita (Nixon and Boyd, 1979).

The *garnets* (Plate 15B) are pyrope rich ($Py_{64-70}$) with a low $Cr_2O3$ (0.05–0.26 wt%) compared with those from the lherzolites. They are similar in composition to megacrysts from kimberlites (Schulze, Chapter 30). *Pyroxenes* in particular, show a reaction relationship with the enclosing alnöite (Fig. 170). Clinopyroxenes range from pale-green subcalcic diopsides to darker augites, and illustrate remarkably regular chemical trends (Figs. 171 and 172; see also Nixon and Boyd, 1979). With decreasing temperature, as

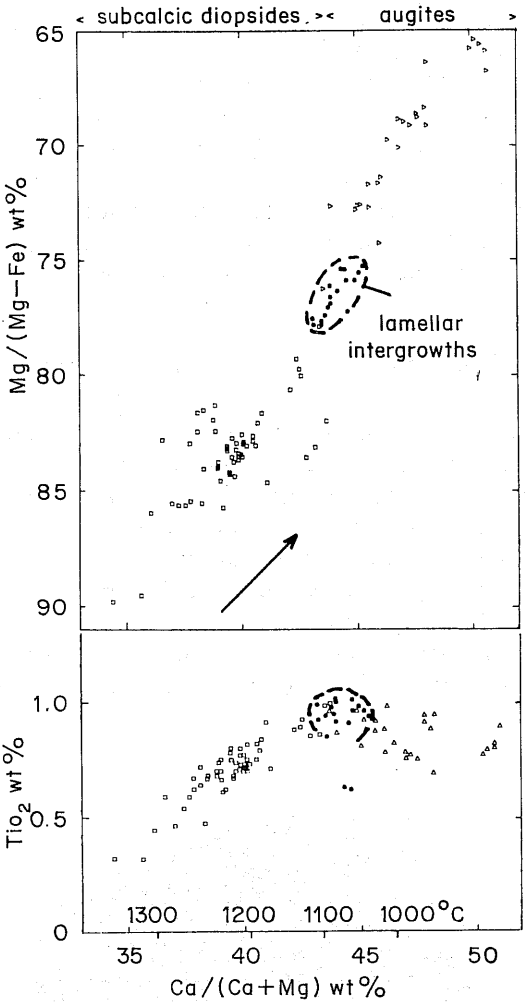

Figure 171. The wide compositional range of clinopyroxene discrete nodules (megacrysts). With progressive crystallisation under decreasing temperature (direction shown by arrow) there is an increase in iron and titanium; the latter drops as soon as the clinopyroxene-ilmenite lamellar intergrowths (encircled areas) start to crystallise. (*Cpx-ilm field courtesy of F. R. Boyd*) cpx symbols: □ subcalcic, • intergrowth, ▷ augite

indicated by Ca/(Ca + Mg), there is a decrease of Cr and an increase in Fe/Mg, Na, and Ti. Mutual inclusion and chemical relationships indicate that some subcalcic diopside crystallised with pyrope garnet and bronzite, and on this general assumption single crystal PT's for bronzite can be estimated (Nixon and Boyd, 1979). The points form a high-temperature inflection of the ambient geotherm (Fig. 174). The subcalcic diopsides grade into augites at about Ca/(Ca + Mg) = 43. The transition is marked by a cluster of clinopyroxenes from ilmenite lamellar intergrowths. There is no evidence that pyrope forms part of this assemblage, nor bronzite (as for example at Frank Smith, South Africa). The subcalcic and lamellar clinopyroxene suites are similar to those found in kimberlites. The augites, in which jadeite and acmite form significant components (Fig. 172), are much more typical of alkali basaltic rocks. In this respect the Malaita intrusions are intermediate in character between the two groups.

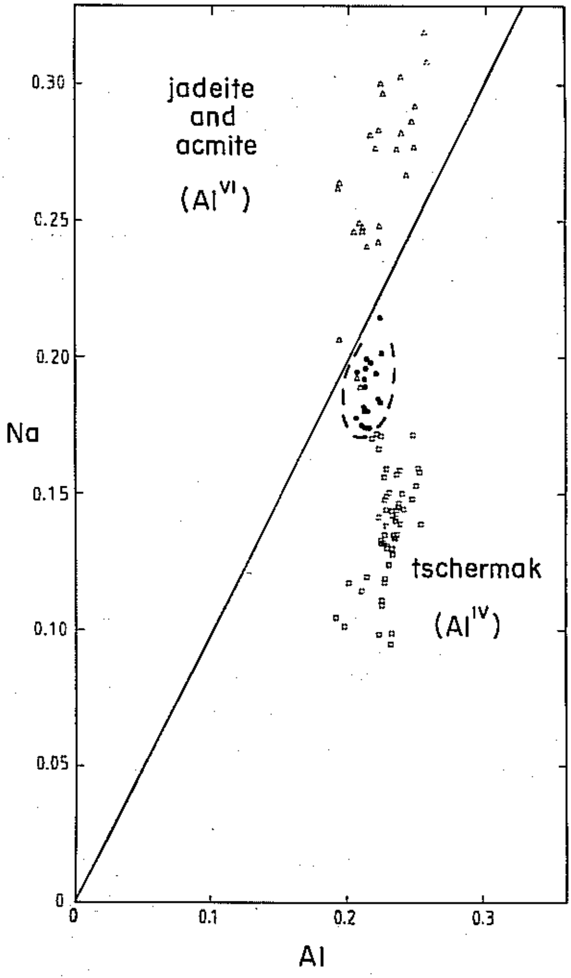

Figure 172. Trend of increasing sodium in subcalcic diopsides (bottom) through cpx from ilm lamellar intergrowths (encircled) to augites (top) without concomitant increase in Al. The coordination state of aluminium changes progressively from 4 to 6. All pyroxenes contain negligible Cr; the excess of Na over Al in the augites must therefore be balanced by ferric iron. (*Cpx-ilm field courtesy of F. R. Boyd*)

*Ilmenite* is conspicuous as lamellar intergrowths with clinopyroxene at Babaru'u and Kwaikwai, but remarkable discrete nodules with a natural polish (Plate 15D), possibly imparted during fluidised eruption, were obtained from an unknown intrusion in the same area. The former compare with ilmenites from similar intergrowths from kimberlites (Fig. 173; they contain between 7 and 8.5 wt% MgO. The discrete ilmenites, however, range to below 4 wt% MgO. $Cr_2O_3$ is negligible in both cases. The trend towards increasing $FeTiO_3$ and $Fe_2O_3$ (Fig. 173) is compatible with the suggestion that the intergrowths crystallised first within a fractionating magma followed by the discrete ilmenite, assuming that the unknown host intrusion has similar characteristics to those of Babaru'u and Kwaikwai (Neal, 1985). Clinopyroxene-ilmenite lamellar intergrowths have been produced experimentally from melts at 38 kb (Wyatt *et al.*, 1975; and Plate 15E). Associated with the polished ilmenites are zircons (Plate 15F) which are also found in gravels derived from alnöites.

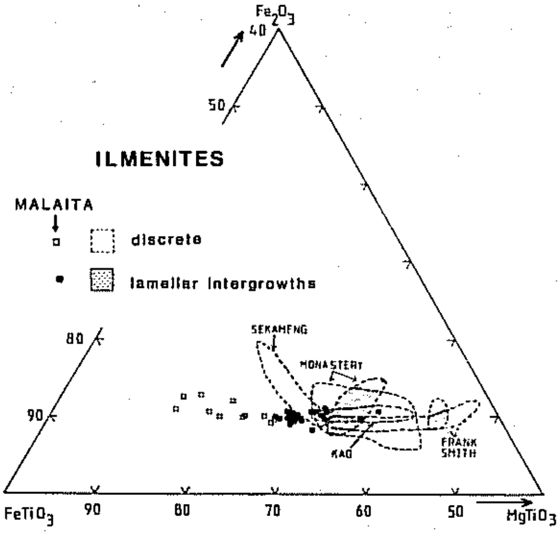

Figure 173. Compositions of some ilmenites from Malaita alnöites (*cpx-ilm field courtesy of F. R. Boyd*)

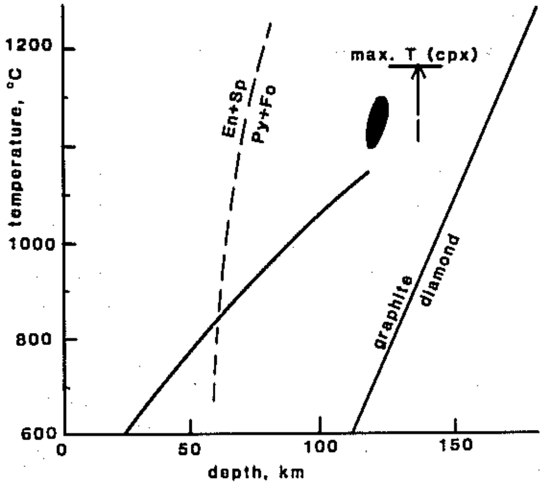

Figure 174. The garnet-bearing peridotites have equilibrated at PTs which can be interpreted to lie on a geotherm prevailing at the time of alnöite eruption and well within the graphite stability field (Finnerty and Boyd, Chapter 27). The black area encloses points derived from analyses of six bronzite discrete nodules (for derivation see Nixon and Boyd, 1979). Temperatures similarly derived from subcalcic clinopyroxenes extend to over 1300 °C. The dashed line represents the experimentally determined spinel-pyrope transition (Wood and Holloway, 1984)

A xenolith consisting of an aggregate of megacrysts (3 mm, max. 9 mm across) mainly of clinopyroxene, orthopyroxene, and pyrope but with some lherzolitic minerals, e.g. Cr-diopside, appears to be a *cumulate*. Most grains are subrounded and show overgrowths or reaction rims with inclusions (Plate 15C) indicative of disequilibrium with an interstitial fluid. The groundmass is fine-grained (rapidly cooled?) phlogopite, ilmenite, garnet, orthopyroxene, amphibole, and sulphides. The origin is uncertain; in some respects the rock resembles the rare polymict aggregates in kimberlites (Lawless *et al.*, 1979 – see Nixon, Chapter 17). It may be a segregation of an early magmatic phase in the evolution of the alnöite intrusion; the Fe-Ti enriched matrix may be related to rare phlogopite-ilmenite inclusions that have been found in northern Malaita.

## Rare Earth Elements and Isotope Data

A summary of some results and conclusions of Neal (1985), which will be reported in detail elsewhere, is given below. Isotopic and REE analyses have been carried out on mineral separates (cpx, amph, and gt) from the peridotites, and on fresh, unaltered whole-rock alnöite.

Clinopyroxenes from *spinel lherzolites* have isotopic ratios which delineate a vertical array on a diagram of $^{143}Nd/^{144}Nd$ vs. $^{87}Sr/^{86}Sr$. The variation is in the Nd isotopic ratios with the Sr ratios relatively constant ($\varepsilon_{Nd} = +2.5$ to $+14.2$; $\varepsilon_{Sr} = -11.7$ to $-29.4$). It is concluded that the vertical array has been produced as a result of contact with a Sr-rich subduction-related fluid. ($^{87}Sr/^{86}Sr = 0.7025-07030$) which had the effect of 'homogenising' Sr isotopic ratios. This fluid also imparted an enrichment of La and Ce on originally LREE depleted profiles (without affecting the Sm/Nd ratios). Nd model ages indicate that this enrichment took place < 200 m.y. ago. Subduction-related volcanics with an $^{87}Sr/^{86}Sr$ ratio of 0.7025–0.7030 have been described by Woodhead and Fraser (1985) from the northern Mariana Islands. Our isotope results are thus compatible with the idea that the mantle beneath Malaita experienced an influx of fluids derived from subducted ocean crust. This interpretation is in contrast to that advocated by Bielski-Zyskind *et al.*, (1984), which predicted a *continental* crust component within the source region of the OJP basalts and alnöites.

Nd isotope systematics of the spinel lherzolites imply that they have undergone partial melting ± enrichment (Neal, 1985). Amphiboles in three spinel lherzolites are in isotopic disequilibrium with the clinopyroxenes. Combined isotope systematics of the amphibole indicate that it cannot have been derived from the alnöite, and calculations suggest it formed < 200 m.y. ago.

The *garnet-bearing* lherzolite data refer almost exclusively to those xenoliths that equilibrated below pressures equivalent to about 95 km; two xenoliths from greater depths show different results which have not yet been evaluated. Most data, although variable, can be integrated with those of the spinel lherzolites. There is Sr isotopic disequilibrium between garnet and clinopyroxene in two xenoliths, which is attributed to inherited values of amphibole assimilated during garnet formation.

The clinopyroxenes form an approximately isochronous relationship for Sm/Nd, interpreted as a mixing between a MORB source (represented by CRN 213 in Fig. 175) and a metasomatic fluid (4013 and 4069); Nd model ages show that mixing occurred less than 240 m.y. ago. Three out of

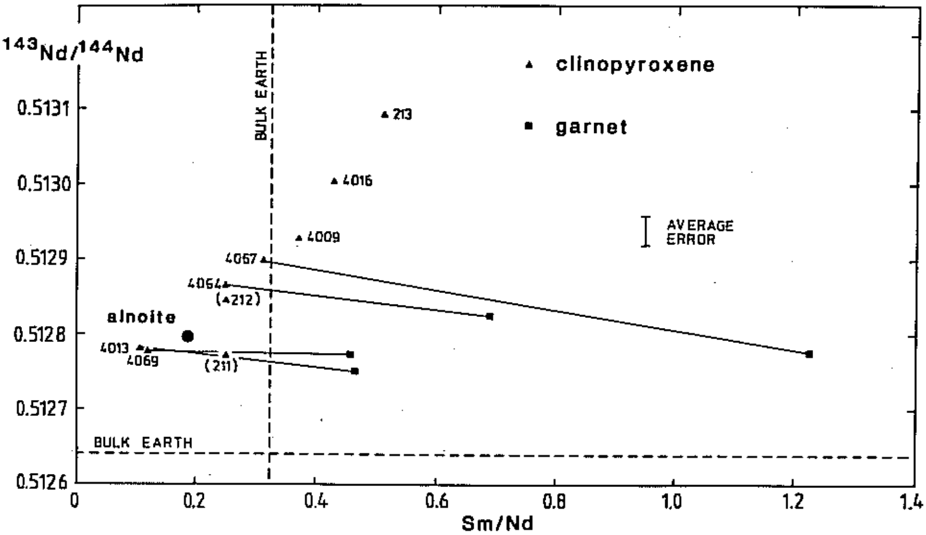

Figure 175. Nd isotopic and Sm/Nd values for seven clinopyroxene and garnet pairs in spinel garnet lherzolites of moderate depth, Malaita. Two clinopyroxenes, 211 and 212, are petrologically and chemical aberrant (see Neal, 1985, for details)

seven clinopyroxenes analysed for Sr isotopes do not conform to such a mixing line and this is thought to be due to an Sr metasomatism similar to that which affected the spinel lherzolites. The alnöite is not the primary cause of the metasomatism since it does not show an 'end member' isochronous relationship (Fig. 175).

Of the *discrete nodule* (*megacryst*) suite, only subcalcic diopsides and garnets have been examined in detail. The REE patterns of the diopsides are of similar humped shape (Fig. 176) with five samples showing La/Nd, 0.261–0.324, and Eu/Yb 1.77–2.36. The sixth sample, 198, has lower total REE and ratios of La/Nd (0.147) and Eu/Yb (2.5). The garnets show little variation in total REE and REE profiles: they are extremely depleted (La/Eu, 0.03–0.10).

Modelling of the REE results after the manner of Jones (Chapter 48) shows that the clinopyroxenes probably crystallised from an alkali basaltic liquid (Neal, 1985). The same cannot be said for garnets, using published distribution coefficients, notwithstanding the fact that mutual inclusion relationships have been observed. It seems likely that the distribution coefficients require reviewing.

Four subcalcic diopside megacrysts have a small range of $\varepsilon_{Sr}$ ($-11.9$ to $-8.4$); sample 198 is higher ($-6.4$). All values are less than the lowest obtained for alnöite ($-4.2$) from which they cannot have crystallised (Neal, 1985). It has not been determined whether the more evolved megacrysts (the most ferriferous augites) were cognate with the alnöite but this seems unlikely since groundmass alnöite clinopyroxenes are comparatively

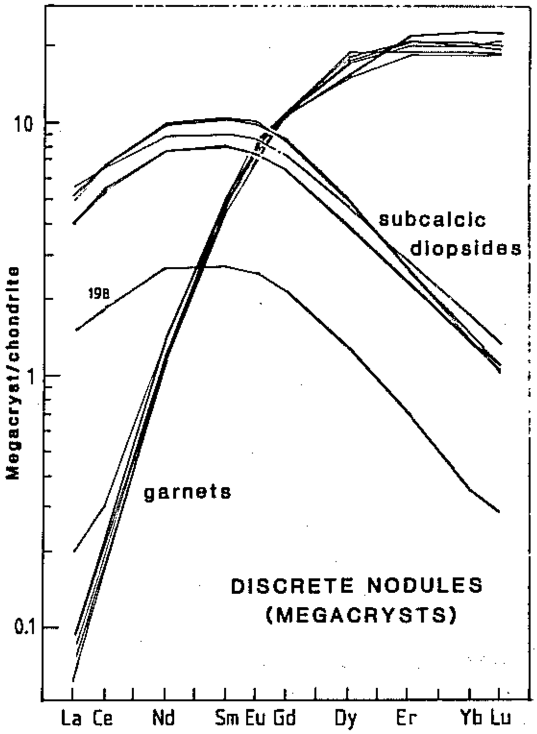

Figure 176. REE patterns for garnet and subcalcic diopside discrete nodules (megacrysts), Malaita. Specimen 198 is a clinopyroxene of 'standard' composition which appears to have evolved differently from the others

poor in iron and sodium (Delaney *et al.*, 1979a; Nixon *et al.*, 1980).

## Discussion

It is essential to understand the mantle stratigraphy of the OJP since chemical and isotopic variations with depth are considerable. The main depth zones (modified from Nixon and Boyd, 1979) are as follows:

thickened basaltic crust up to 40+ km in northern OJP (? lower garnet granulite layer)

depleted harzburgite (by analogy with ophiolites)

spinel lherzolite (xenoliths as described in this chapter)

spinel garnet lherzolite (xenoliths as described in this chapter)

garnet (± chromite) lherzolite (few of these deep-seated lherzolites available)

asthenospheric zone (in which parental magma and deepest megacrysts crystallise)

*(maximum depth represented by xenoliths is 135 km)*

Tectonic, chemical, and isotopic data point to several basaltic depletion events in the Mesozoic. The first of these is the formation of the basaltic crust of the OJP probably contemporaneously with the initial break-up of Pacific Plate. Two later basaltic episodes have been described on Malaita (see above). The depths from which these basalts were generated are uncertain but their collective 'depletion' imprints appear to be in the xenoliths. The enrichment of LREE, Sr, and the development of amphibole arose from metasomatic processes which we argue could have resulted from incorporation of components of oceanic crust during subduction. Subduction of a cool lithospheric slab is also the most likely explanation for the formation of garnet rims which envelope primary spinels within our mantle sample. Such a concept might explain the thickening of the OJP lithosphere, i.e. repetitive slices produced by low-angle thrusts. Alternatively, or in addition, garnet formation might be linked with cessation of the Cretaceous mid-plate volcanism and dolerite sill emplacement, referred to above, i.e. falling temperatures within the lithosphere combined with increasing pressure resulting from deflation of the Pacific floor (Darwin Rise).

The events leading to *alnöite eruption* at a time (34 m.y. B.P.) when the OJP was much further east than at present, are conjectured to have resulted from an asthenospheric hot spot (cf. Sen, Chapter 26) or rising diapir. During its ascent, partial melting occurred (Green and Gueguen, 1974) to produce a proto-alnöite of alkali basalitc character. Crystallisation possibly near the base of the lithosphere gave rise initially to the most magnesian of the discrete nodule (megacryst) suite. With further cooling and differentiation of the magma, successive clinopyroxene megacrysts developed considerably modified chemical parameters, as described above.

The *vertical* range over which crystallisation took place is somewhat controversial (q.v. Schulze, Chapter 30). The large temperature range of 300–400 °C over which the clinopyroxenes crystallised rivals those of the discrete nodules of the Letseng la terae kimberlite, Lesotho (Boyd and Nixon, 1973). These authors proposed that the nodules coexisted with small amounts of interstitial liquids over a depth interval of several tens of kilometres. Delaney *et al.* (1979a) suggest that sequential crystallisation took place beneath Malaita as the magma rose intermittently. At Monastery Mine, South Africa, Gurney *et al.* (1979) interpreted the chemical data of a comprehensive suite to indicate isobaric (single depth) crystallisation. The lack of coarse silicate exsolution features within pyroxenes of such suites, including those studied here, does not support the idea of cooling through several hundred degrees at a single depth. Furthermore, the lack of garnet which can be equated with the augites strongly suggests that the latter have crystallised in the spinel stability field, i.e. at relatively shallow depths.

A late-stage effect of the rising magma was the formation of slivers of secondary clinopyroxene and amphibole described earlier. A consideration of the equilibration temperatures of the spinel garnet lherzolites in which these occur and the large temperature range of the megacrysts makes it likely that the megacryst magma was the responsible magma. If this was the case then the transformation of this magma to alnöite, probably by a process of fractional crystallisation of megacrysts and assimilation of metazsomatized wall rock, took place at shallow depths within the upper mantle.