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# Partitioning of elements between majorite garnet and melt and implications for petrogenesis of komatiite

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**Abstract.** Partitioning of elements between majorite garnet and ultrabasic melt has been studied at 16 GPa and 1950° C. Ca, Ti, La, Sm, Gd, Zr, Hf, Fe, Ni, Mn, K, and Na are enriched in the melt, whereas Al, Cr, V, Sc and Yb are concentrated in majorite garnet. Thus, majorite garnet fractionation by partial melting could produce chemical heterogeneities in these elements deviating from chondritic abundance. Using the partitioning behaviour of elements between majorite garnet and ultrabasic melt, the petrogenesis of komatiite is discussed. A simple model to explain the chemical varieties of komatiites is as follows. Alumina-depleted komatiite was generated by partial melting of the primitive mantle at 200–650 km depth, and alumina-enriched komatiite is the product of remelting of the residual solid at the same depths, whereas alumina-undepleted komatiite was formed by partial melting of the primitive upper mantle at depths shallower than 200 km. We suggest the possibility of large-scale chemical layering or heterogeneity in the early Archean upper mantle as an alternative model for komatiite genesis; shallower mantle depleted in majorite garnet and the underlying mantle enriched in majorite garnet. Alumina-depleted and alumina-enriched komatiites in the early Archean might be generated by a high degree of partial melting of the layered mantle. Such chemical layering could have been homogenized by the late Archean. This explains the observations that alumina-depleted and alumina-enriched komatiites were generally formed in the early Archean but alumina-undepleted komatiite was erupted in the late Archean.

## Introduction

Recent experiments on melting of ultrabasic compositions such as chondritic and peridotite mantle compositions have clarified the changes in melting relations with increasing pressure up to ca. 25 GPa; olivine is the liquidus phase down to the base of the upper mantle; majorite garnet crystallizes along the liquidus curve at depths of the transition zone of the mantle, whereas Mg-perovskite is the liquidus phase in the lower mantle (e.g. Takahashi 1986; Ohtani et al. 1986; Ohtani and Sawamoto 1987; Ito and Takahashi 1987). These results indicate that majorite garnet fractionation is important during solid-melt segregation in the transition zone (from 15 to 25 GPa). The melting relations in these two ultrabasic compositions are illustrated in Fig. 1.

Recently, Kato et al. (1987, 1988) reported the partition coefficients of major and minor elements between majorite garnet, Mg-perovskite, Ca-perovskite, and melt at high pressure and high temperature. On the basis of the experimental partitioning data, they concluded that fractionation of majorite garnet and Mg-perovskite did not occur in the early stages of terrestrial evolution. However, compositions of komatiites suggest the possibility of global chemical layering or heterogeneity created by garnet fractionation in the early Archean mantle (e.g. Ohtani 1984; Arndt 1986). In order to discuss the petrogenesis of komatiites, partitioning of some key elements between majorite garnet and ultra-

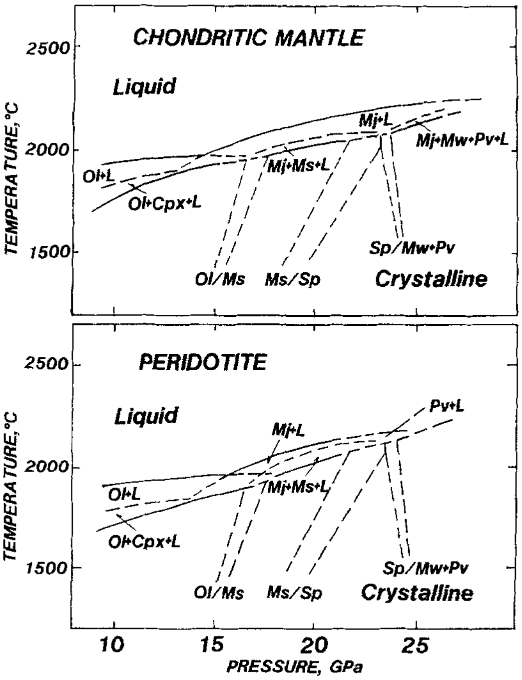

**Fig. 1.** Schematic phase relations of the chondritic mantle and peridotite compositions. The phase relations of chondritic mantle are based on Ohtani et al. (1986) and Ohtani and Sawamoto (1987), while those of peridotite are based on Ohtani (1987) and Nagata et al. (1987). *Ol*, Olivine; *Cpx*, Ca-poor clinopyroxene; *Mj*, majorite garnet; *β*, modified spinel; *Sp*, spinel; *Mw*, magnesiowustite; *Pv*, Mg-perovskite; *L*, liquid

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basic liquid was studied experimentally. The elements studied were 11 refractory lithophile elements (Al, Ca, Ti, Sc, V, La, Sm, Gd, Yb, Zr, and Hf) 4 siderophile elements (Fe, Ni, Mn, and Cr), and 2 volatile elements (Na and K).

## Experimental procedure

### *Preparation of the starting materials*

We have used two types of starting materials, A (chondrite-REE) and B (chondrite-B), for the partitioning experiments. These are shown in Table 1. In both starting materials A and B, the major elements Mg, Si, Ca, and Al are chondritic, whereas the iron content is reduced to Fe/(Mg+Fe) = 0.1 in atomic ratio. These compositions are similar to those used for the previous experiments on melting of chondritic mantle (Ohtani and Sawamoto 1987). Starting material A contains about 1200 ppm of La, Sm, Gd, Yb, Zr, and Hf each. The reagent grade oxides of these elements were dissolved in an acid solution of $HNO_3$ or HCl. The desired values of the solution containing these elements were dripped on a sintered disc of a mixture of olivine, pyroxenes, and anorthite. The disc was dried with an infrared lamp, and then pulverized in an agate mortar. This powder was further heated at 1150° C for 10 h in a furnace with controlled oxygen partial pressure of $pO_2 = 10^{-12}$ atm. The B starting material contains about 2000 ppm of Ti, V, Ni, Mn, Cr, Na, and K. The reagent grade oxides or carbonates of these elements were added to a mixture of olivine, pyroxenes, and anorthite in the $CaO-FeO-MgO-Al_2O_3-SiO_2$ system and heated at 1150° C for 10 h in a furnace with controlled oxygen partial pressure ($pO_2 = 10^{-12}$). The starting materials A and B thus synthesized were mixtures of olivine, pyroxenes, and anorthite.

### *High pressure and temperature experiments*

A multiple-anvil (MA8) high-pressure apparatus was used for the high-pressure and temperature experiments. The MA8 apparatus was driven by the guide-block/1500-ton uniaxial press system of Ehime University (Ohtani et al. 1987). The furnace assembly is similar to that used in previous experiments on melting of chondritic mantle and peridotite (Kato and Kumazawa 1986; Ohtani 1987). The sample was enclosed in a graphite capsule, which was placed in the centre of the pressure medium made of semi-sintered magnesia. A sheet heater made of a hybrid material of tungsten carbide and diamond was used for the experiments. The run temperatures were monitored by a W3%Re-W25%Re thermocouple with a diameter of 0.1 mm. No correction was made for the effect of pressure on the EMF of the thermocouple.

The experiments were made by using the quenching method. The pressure was increased to 16 GPa first, and then the temperature was elevated to the desired value and held constant for about 4 min. The sample was quenched by shutting off the power supply of the furnace, and the pressure was released to recover the sample. Several experiments were made at 16 GPa. The recovered samples for the successful runs with starting materials A and B were selected and used for chemical analyses of the elements. The conditions of the experiments (runs 59 and N33) used for chemical analyses were 16 GPa and 1950 ± 10° C.

### *Chemical analysis of the run products*

Chemical analysis of the run products was made by EPMA. The samples recovered were polished by using diamond powders for EPMA analysis. A JEOL 733 electron microprobe was operated at an accelerating voltage of 15 kV and a beam current of $5 \times 10^{-9}$ A. The counting time was 10 s for the chemical analyses of the elements in the starting material B. X-ray intensities were corrected for deadtime, background, and matrix effects using Bence and Albee's (1968) method. Analyses of the minor elements, La, Gd, Sm, Yb, Zr, and Hf, were made by the same electron microprobe with the operating conditions of accelerating voltage of 15 kV, the beam current of $5 \times 10^{-9}$ A, and a counting time of 800 s. Four glasses containing La, Sm, Gd, Yb, Zr and Hf were prepared as standards to analyse these elements with the electron microprobe analyser. The compositions of the standard glasses were determined by analyses of separate aliquots using an inductively coupled plas-

**Table 1.** Chemical compositions (wt%) of the starting materials

| | Composition A Chondrite-REE | Composition B Chondrite-B |
|---|---|---|
| $SiO_2$ | 48.35 | 48.18 |
| $Al_2O_3$ | 4.00 | 3.99 |
| FeO | 6.84 | 6.82 |
| MgO | 36.56 | 36.43 |
| CaO | 3.17 | 3.16 |
| $TiO_2$ | – | 0.16 |
| $V_2O_3$ | – | 0.1 |
| $Cr_2O_3$ | – | 0.17 |
| NiO | – | 0.2 |
| MnO | – | 0.2 |
| $Na_2O$ | – | 0.2 |
| $K_2O$ | – | 0.2 |
| $La_2O_3$ | 0.140 | – |
| $Sm_2O_3$ | 0.150 | – |
| $Gd_2O_3$ | 0.146 | – |
| $Yb_2O_3$ | 0.148 | – |
| $Sc_2O_3$ | 0.197 | – |
| $ZrO_2$ | 0.167 | – |
| $HfO_2$ | 0.129 | – |

Determined by ICP-AES

**Table 2.** Chemical compositions of standard glasses for the EPMA analysis

| | Glass 4 | Glass 1 | Glass 2 | Glass 3 |
|---|---|---|---|---|
| La (ppm) | 473 | 964 | 1390 | 1930 |
| Ce | 424 | 917 | 1310 | 1870 |
| Sm | 480 | 948 | 1410 | 1960 |
| Gd | 470 | 985 | 1410 | 1990 |
| Yb | 485 | 970 | 1450 | 2000 |
| Sc | 486 | 984 | 1410 | 1990 |
| Zr | 491 | 1040 | 1460 | 2030 |
| Hf | 480 | 1050 | 1470 | 2070 |
| $Al_2O_3$ | 26.94 | 27.00 | 26.66 | 26.79 |
| MgO | 9.43 | 9.36 | 9.41 | 9.41 |
| CaO | 20.48 | 20.51 | 20.74 | 20.44 |
| FeO | 0.043 | 0.041 | 0.053 | 0.091 |
| $SiO_2$ | 42.65 | 42.14 | 41.76 | 41.34 |

Determined by ICP-AES

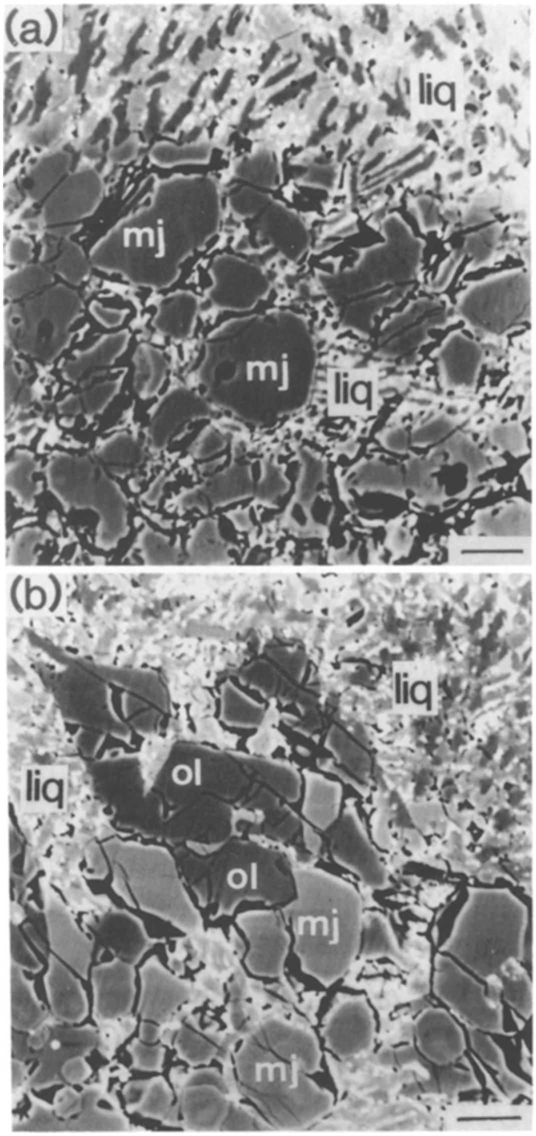

**Fig. 2a, b.** Back-scattered electron images of the run products made at 16 GPa and 1950° C. **a** Run 59 in which the starting material A was used; **b** run N33 in which the starting material B was used. *mj*, Majorite garnet; *liq*, liquid; *ol*, olivine. The *scale bar* represents 10 μm

**Table 3.** Chemical compositions (wt%) of minerals and partial melt at 16 GPa and 1900° C

| | Run 59 | | Run N33 | | |
|---|---|---|---|---|---|
| | Majorite | Liquid | Majorite | Olivine | Liquid |
| $SiO_2$ | 52.56 | 49.11 | 51.48 | 42.66 | 48.20 |
| $TiO_2$ | – | – | 1.69 | 0.07 | 2.69 |
| $Al_2O_3$ | 9.99 | 3.42 | 9.06 | 0.14 | 2.86 |
| $Cr_2O_3$ | – | – | 0.79 | 0.09 | 0.34 |
| $V_2O_3$ | – | – | 1.07 | 0.16 | 0.74 |
| FeO | 2.45 | 5.96 | 2.65 | 3.64 | 6.11 |
| NiO | – | – | 0.05 | 0.32 | 0.34 |
| MnO | – | – | 0.48 | 0.34 | 0.80 |
| MgO | 33.07 | 37.44 | 30.71 | 53.14 | 34.18 |
| CaO | 1.27 | 3.22 | 1.86 | 0.07 | 3.40 |
| $Na_2O$ | – | – | 0.02 | 0.05 | 0.38 |
| $K_2O$ | – | – | 0.01 | <0.01 | 0.26 |
| $La_2O_3$ | 0.0078 | 0.0882 | – | – | – |
| $Sm_2O_3$ | 0.0269 | 0.1339 | – | – | – |
| $Gd_2O_3$ | 0.0352 | 0.1348 | – | – | – |
| $Yb_2O_3$ | 0.1655 | 0.1572 | – | – | – |
| $Sc_2O_3$ | 0.3526 | 0.2148 | – | – | – |
| $ZrO_2$ | 0.0853 | 0.1849 | – | – | – |
| $HfO_2$ | 0.0919 | 0.1325 | – | – | – |
| Total | 100.01 | 100.20 | 99.86 | 100.74 | 100.30 |

**Table 4.** Partition coefficients between minerals and melt at 16 GPa

| | D(Mj/Liq) | | D(Ol/Liq) |
|---|---|---|---|
| | Present study | Kato et al. (1988) | Present study |
| $Al_2O_3$ | 3.06 | 2.26 | 0.05 |
| $Cr_2O_3$ | 2.33 | 1.52 | 0.26 |
| $V_2O_3$ | 1.45 | – | 0.22 |
| $Sc_2O_3$ | 1.64 | 1.7 | – |
| $TiO_2$ | 0.63 | 0.4 | 0.03 |
| CaO | 0.47 | 0.6 | 0.02 |
| FeO | 0.42 | 0.6 | 0.59 |
| NiO | 0.15 | – | 0.94 |
| MnO | 0.60 | – | 0.42 |
| $Na_2O$ | 0.05 | 0.1 | 0.13 |
| $K_2O$ | 0.04 | – | <0.04 |
| La | 0.09 | <0.1 | – |
| Sm | 0.20 | 0.2 | – |
| Gd | 0.26 | – | – |
| Yb | 1.05 | 1.4 | – |
| Zr | 0.46 | 0.6 | – |
| Hf | 0.69 | 0.8 | – |

ma-atomic emission spectrometer (Shimazu ICPS-50) and are listed in Table 2.

The liquidus majorite garnet crystals are usually greater than 10 μm in diameter. An electron beam of about 1 μm diameter was used to measure compositions of majorite garnet crystals co-existing with the quenched liquid. A defocussed beam of 50 μm diameter was used to analyze the quenched liquid composed of a fine-grained mixture of quench crystals with a typical grain size from 1 to 5 μm. The quenched liquid was mainly composed of olivine, pyroxenes, and garnet with a fine-grained lamellar textures as shown in Fig. 2a and b.

## Experimental results

The compositions of garnet and co-existing liquid at 16 GPa and 1950° C are summarized in Table 3. Majorite garnet was the liquidus phase for starting material of composition A (run 59), whereas the starting material of composition B had olivine together with majorite garnet at around the liquidus temperature (run N33; see Fig. 2a, b). This difference in the melting relations may originate in the minor differences in the bulk compositions of the starting materials, i.e. addition of a small amount of elements such as Na and K may enhance the olivine stability field. This result is consistent with that reported by Kushiro (1975) in the lower pressure range.

The partition coefficient, $D$, is defined as the ratio of the weight fraction, $C_{\mathrm{Mj}}/C_{\mathrm{L}}$, of an element in majorite gar-

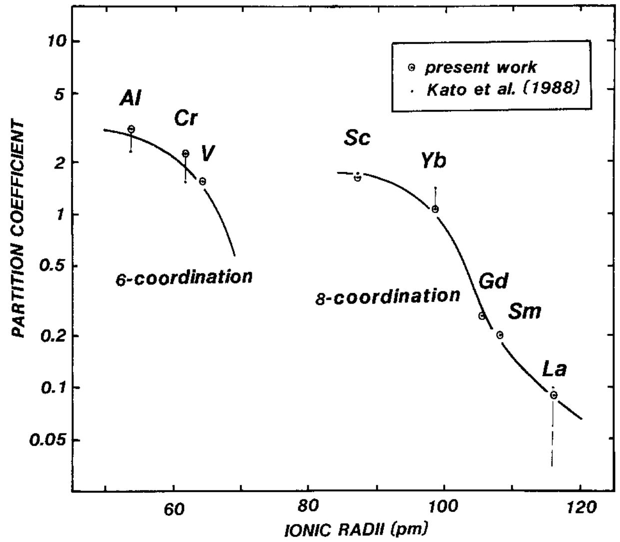

**Fig. 3.** The relation of partition coefficient and ionic radii. Six coordination ionic radii (Shannon 1976) are used for Al, Cr, and V, while eight coordination ionic radii are used for La, Sm, Gd, Yb, Sc. The data of Kato et al. (1988) are also shown

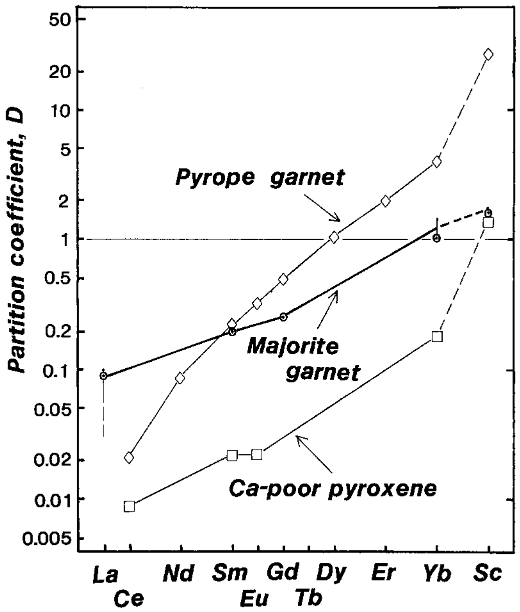

**Fig. 4.** The partition coefficients, *D*, of REE and Sc between minerals and melt. D for majorite garnet and the liquid are from the present work and Kato et al. (1988); that for pyrope garnet and the liquid is based on Shimizu and Kushiro (1975) and Irving (1978); that for Ca-poor pyroxene and the liquid is based on Weill and McKay (1975) and McKay and Weill (1977)

net, $C_{Mj}$, and the partial melt, $C_L$. The partition coefficients of the major and minor elements measured in the present experiments are summarized in Table 4. The results of Kato et al. (1988) are also shown in the same table. The present result is qualitatively consistent with that of Kato et al. (1988).

As shown in Table 4, Al, Cr, V, and Sc are concentrated in majorite garnet, whereas the other elements, refractory lithophile elements such as Ti, Ca, and light and intermediate REE such as La, Sm, Gd, siderophile elements, Fe, Ni, and Mn, and volatiles, Na and K are concentrated in the melts. The partition coefficients, *D*, of trivalent cations are plotted against the ionic radii (Shannon 1976) in Fig. 3; the 8-coordination radii are used for Sc, and REE in garnet structure, whereas the 6-coordination radii are used for Al, Cr, and V. The systematic change in partition coefficient with ionic radii of elements is clearly observed in this figure. Figure 4 summarizes the partition coefficients, D, of REE and Sc between pyrope-rich garnet and the melt (Shimizu and Kushiro 1975; Irving 1978), and those between Ca-poor pyroxene and the melt (Weill and McKay 1975; McKay and Weill 1977) together with those of majorite garnet and the melt determined in this work. The partition coefficients, D, for majorite garnet and liquid are qualitatively similar to those of pyrope-rich garnet and liquid, i.e. low D values for light REE, while relatively high D values for heavy REE. However, the fractionation of REE and Sc by separation of majorite garnet is weaker than that generated by separation of pyrope-rich garnet.

## Discussion

### *Majorite garnet fractionation and petrogenesis of komatiites*

The present results clearly demonstrated the signature of majorite garnet fractionation. If partial melting during the early stages of terrestrial evolution proceeded into the deep upper mantle and the transition zone, majorite garnet is the last phase to be consumed in the melt, i.e. it is the liquidus phase of ultrabasic melts at high pressure. The melt co-existing with majorite garnet in Table 3 has a high-MgO komatiite composition. Partial melting of a wide range of ultramafic materials such as chondrite and lherzolite can produce komatiite liquids co-existing with a residual solid of majorite garnet at above 16 GPa (e.g. Takahashi 1986; Ohtani et al. 1986). Thus, majorite garnet fractionation during partial melting at the base of the upper mantle or in the transition zone can produce komatiitic melt enriched in Ca, Ti, and LREE and depleted in Al, Sc, V, and HREE relative to chondritic abundances.

Komatiites, which are usually observed in the Archean, may be divided into two or three types on the basis of the major and minor element chemistry (e.g. Nesbitt et al. 1979): alumina-undepleted type, alumina-depleted type, and alumina-enriched type komatiites. Alumina-undepleted komatiite has roughly chondritic ratios of $CaO/Al_2O_3$, $Al_2O_3/TiO_2$, and Gd/Yb ratios, whereas alumina-depleted komatiite, which has been reported from Barberton and some other localities, has high $CaO/Al_2O_3$ ratios (about 1.5), low $Al_2O_3/TiO_2$ ratios (about 11) and depletion of heavy REE, V, and Sc compared to the chondritic values (Viljoen and Viljoen 1969).

A model involving the fractionation of pyrope-rich garnet has been suggested to explain the chemical features of komatiites with non-chondritic $CaO/Al_2O_3$, $Al_2O_3/TiO_2$, and Gd/Yb ratios (e.g. Nesbitt et al. 1979; Jahn et al. 1982). However, with the previous experimental data, it appears to be impossible to generate komatiites and retain garnet in the residue by partial melting of lherzolite at about 3 GPa, because pyrope-rich garnet is the first phase to disappear during melting of lherzolite as was discussed by Smith and Erlank (1978).

Ohtani (1984) proposed that non-chondritic komatiites such as alumina-depleted komatiite could be formed by partial melting and selective gravitational separation of garnet

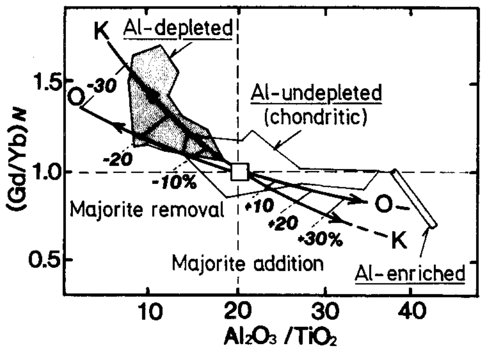

**Fig. 5.** Change in $Al_2O_3/TiO_2$ and the chondrite normalized Gd/Yb ratios by removal and addition of majorite garnet. *O*, Partition coefficients of the present work are used; *K*, those by Kato et al. (1988) are used. Removal and addition of majorite garnet up 30wt% are shown. The compositional range of alumina-depleted, alumina-enriched, and alumina-undepleted komatiites after Jahn et al. (1982) and Arndt (1986) are also shown

at depths from 200 km to at least 450 km, because garnet is the only phase which is denser than the ultrabasic liquid. Recent data on melting relations of chondritic mantle and peridotite compositions to 25 GPa (e.g. Takahashi 1986; Ohtani et al. 1986; Ohtani and Sawamoto 1987; Ito and Takahashi 1987) showed that majorite garnet is the liquidus phase at depths of the deep upper mantle and the transition zone. Alumina-depleted komatiite may be produced by partial melting of the primitive mantle, retaining majorite garnet as the residual solid at depths from 450 km to 650 km. Therefore, alumina-depleted komatiite can be generated by partial melting of the ascending diapir at depths from 200 to 650 km, where garnet occurs as the residual solid and separates from the melt. Alumina-enriched komatiite may be generated by further ascending and melting of the lower garnet-rich portion of the diapir. On the other hand, alumina-undepleted komatiite could be derived from a high degree of partial melting of primitive mantle at shallower depths ($<450$ km), where olivine is the liquidus phase. Komatiites generally have various MgO values, from basaltic to peridotite composition (e.g. Jahn et al. 1982). Such variation in MgO and $SiO_2$ in komatiites can be explained by difference in depth of melting, degree of partial melting, or degree of olivine fractionation in shallow levels.

Figure 5 shows changes in Gd/Yb and $Al_2O_3/TiO_2$ ratios caused by removal and addition of majorite garnet. The compositions for alumina-depleted, alumina-enriched, and alumina-undepleted komatiites from Jahn et al. (1982) and Arndt (1986) are also shown in the same figure. It clearly implies that removal and addition of majorite garnet can explain minor element features of non-chondritic komatiites.

## *Possibility of chemical layering in the Archean mantle and convective mixing of the layered mantle*

A model of global melting in the Archean mantle has been proposed by Nisbet and Walker (1982) to explain the petrogenesis of komatiites. On the other hand, Ohtani et al. (1986) and Herzberg et al. (1988) suggested the possibility of global melting and subsequent fractionation of majorite garnet in the early stages of the earth's history. Majorite garnet fractionation by melting could produce a shallower mantle depleted in majorite garnet, and the underlying mantle enriched in majorite garnet. The geochemical nature of non-chondritic komatiites may be explained by chemical layering (or heterogeneity) involving majorite garnet in the Archean mantle; alumina-depleted komatiite was formed by a high degree of partial melting of the shallower mantle which was depleted in majorite garnet.

Hf isotopic data for early and late Archean komatiites (Gruau, personal communication) suggest no evidence of fractionation between Lu and Hf in their source region, i.e. the source region was near chondritic in terms of Lu and Hf, although the earlier preliminary data (Gruau et al. 1986) suggested large negative $\varepsilon_{Hf}$ values for the early Archean komatiites. Present data on partritioning of Yb and Hf between majorite garnet and ultrabasic liquid indicate that almost no fractionation of Lu and Hf is expected to occur by majorite garnet separation, making a reasonable assumption that partitioning behaviour of Lu is similar to that of Yb; 30% separation of majorite garnet will produce a residual mantle with a Lu/Hf ratio only about 0.9 times of the chondritic ratio. Thus, a model of chemical layering in the early Archean mantle involving depletion and enrichment of majorite garnet, which was formed in the earlier stages of the earth's history, seems to be consistent with the existing data on Hf isotope of komatiites. However, it is undoubtedly important to determine directly the partition coefficient of Lu between majorite garnet and melt in the future, because Lu and Yb might have different partitioning behaviour.

Early Archean komatiites observed (about 3.5 Ga old) are generally of the alumina-depleted type, and have high $CaO/Al_2O_3$, low $Al_2O_3/TiO_2$ ratios and depletion of HREE, V, and Sc compared to the chondritic values. On the other hand, the later Archean (about 2.7 Ga old) komatiites generally show near-chondritic abundances of these elements (e.g. Nesbit et al. 1979; Jahn et al. 1982). These geochemical features of the later Archean komatiites may be explained by convective mixing of the old layered mantle, as has been suggested by Arndt (1986) and Sun (1987). The chemical layering involving depletion and enrichment in majorite garnet existed in the early Archean mantle, whereas this layering was homogenized by convection by the late Archean. This model can be supported by recent numerical calculations of convective mixing in the layered mantle with a density contrast. The upper boundary of the density contrast, which is destroyed by whole mantle convection, is 1% according to Richter and McKenzie (1981), 2%–3% according to Christensen and Yuen (1984), and 4% according to Olsen and Yuen (1982). The density contrast of 2% corresponds to about 20% difference in volume fraction of majorite garnet in the olivine zone of the upper mantle. The amount of majorite garnet needed to produce this magnitude of density contrast is much larger in the transition zone, because density of (modified) spinel is very close to that of majorite garnet. Thus chemical layering produced by majorite garnet fractionation in the early global melting stages tend to be homogenized very easily, and convective mixing is a physically plausible mechanism for removing heterogeneities in the mantle. The important point of the present model is that the chondritic abundances of the sources region are not necessarily diagnostic of a primitive nature, but this may be the result of convective mixing. It is well known that the upper mantle of the MORB source generally has a composition close to chondritic in terms of moderately incompatible lithophile elements such as Ti, Sc, HREE, Hf, and Zr, although it is depleted in incompatible elements. Thus Kato et al. (1988) denied the possibilities of majorite garnet and Mg-perovskite fractionations by global melting in the accretional stages. However, positive evidence of the layered mantle is observed in the geochemical feature of alumina-depleted komatiites in the early Archean as discussed in the above. Such chemical layering observed in the early Archean mantle could be a product of majorite garnet fractionation in the global melting stages succeeding the accretion of the earth and it might have been much more widespread and distinct in the earlier stages of terrestrial evolution. The present model of convective mixing of an old layered mantle indicates that the modern MORB may have originated from the homogenized portion of the upper mantle.

It should be noted that, strictly speaking, the alumina undepleted-komatiite does not have a chondritic $CaO/Al_2O_3$ ratio, but generally is about 20% higher in this ratio (Sun 1987). This observation is consistent with the high $CaO/Al_2O_3$ ratios for peridotite nodules and Alpine type peridotites (Palme and Nickel 1985), although there is a possibility that slightly high $CaO/Al_2O_3$ ratios in peridotites may be due to artificially biased correction of the samples with high clinopyroxene content (e.g. Hart and Zindler 1986). The non-chondritic feature of Ca and Al, together with depletion of Si in the "homogenized" near-chondritic upper mantle in the later Archean and present upper mantle might be a remnant of the primordial fractionation in the early stage of the earth's history, which was not erased by the convective mixing (Sun 1987). Mg-perovskite and/or Ca-perovskite together with majorite garnet might have been involved in this primordial fractionation.

## Conclusions

1. Partitioning experiments of some major and minor elements between majorite garnet and the ultrabasic liquid have been conducted at 16 GPa. Enrichment of Al, Cr, V, and Sc (and very weak enrichment of Yb) in majorite garnet was observed, while refractory lithophiles, Ca, Ti, La, Sm, Gd, Zr, Hf, siderophiles, Fe, Ni, Mn, and volatiles, Na, and K are enriched in the ultrabasic melt. The partitioning behaviour of majorite garnet is qualitatively similar to that of pyrope garnet. Deviation of Ca/Al, Ti/Al, and LREE/HREE ratios from the chondritic ratios is expected to occur by majorite garnet fractionation.

2. The geochemical features of komatiites can be explained by differences in depth of melting in the primitive mantle; alumina-depleted komatiite is the product of partial melting of the ascending diapir at 200–650 km depths, and alumina-enriched komatiite is generated by melting of the majorite-rich residual portion of the diapir. Alumina-undepleted komatiite is formed by a high degree of partial melting of the shallower primitive upper mantle of $<450$ km depths.

3. As an alternative model for komatiite genesis, we suggest the possibility of large-scale chemical layering in the early Archean mantle; shallower mantle depleted in majorite garnet and the underlying mantle enriched in majorite garnet. Non-chondritic komatiites such as alumina-depleted and alumina-enriched komatiites might be generated from the layered mantle which existed in the early Archean. Such chemical layering could have been homogenized by late Archean time. Recent numerical calculations on mantle convection indicate that convective mixing can homogenize the layered mantle that existed in the early stage of terrestrial evolution. The younger alumina-undepleted komatiite typically in the late Archean might have been formed by partial melting of the homogenized mantle. There is a possibility that near-chondritic abundance of the moderately incompatible refractory lithophile elements of the modern MORB source may not be diagnostic of the primitive nature of the upper mantle, but may be a result of convective mixing during the geological time.

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