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Magnetite lamellae in olivine and clinohumite from Dabie UHP ultramafic rocks, central China

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ABSTRACT

Ultrahigh-pressure (UHP) minerals of the Maowu mafic-ultramafic complex in the Dabieshan, east-central China exhibit many exsolution textures. Magnetite lamellae are common in olivines and Ti clinohumites from harzburgite and garnet pyroxenite. Monazite-(Ce) lamellae occur in apatites from the garnet pyroxenite. Independent P-T estimates suggest that these ultramafic rocks formed at $P > 5$ GPa and $T = 700 \pm 50$ °C. The lamellae-bearing minerals are believed to preserve an earlier, higher P-T record prior to exsolution. Compositions and unit-cell parameters of the magnetite and host olivine and intergrowth relations were determined using a newly developed X-ray diffraction microprobe technique employing synchrotron radiation. The host olivine and magnetite lamellae bear a topotaxial relation with $[220]_{\mathrm{Mag}}\,||\,[200]_{\mathrm{Ol}}$, $[111]_{\mathrm{Mag}}\,||\,[3\,\bar{3}\,\bar{1}]_{\mathrm{Ol}}$, $[11\bar{1}]_{\mathrm{Mag}}\,||\,[331]_{\mathrm{Ol}}$, $[242]_{\mathrm{Mag}}\,||\,[2\,\bar{2}\,0]_{\mathrm{Ol}}$. The recalculated composition of primary olivine may contain up to 1.5 wt% Fe_2O_3. Four hypotheses may explain the observed intergrowths of oriented magnetite lamellae in olivine: (1) oxidation of olivine; (2) decomposition of Fe^{3+}-bearing olivine formed at >6 GPa; (3) exsolution of a spinel (wadsleyite) solid-solution Fe_3O_4-$(Fe,Mg)_2SiO_4$ during decompression; and (4) breakdown of phase A $[Mg_7Si_2(OH)_6]$ + enstatite. The third hypothesis appears to be the most likely inasmuch as no additional silicate phase occurs as associated inclusions in the olivine host. However, the actual mechanism for exsolution of magnetite from olivine remains to be studied experimentally.

INTRODUCTION

The petrogenesis of meta-ultramafic rocks from the Sulu-Dabie terrane of east-central China has emerged recently as a key problem for studies of the upper mantle, because they contain evidence for ultrahigh-pressure (UHP) recrystallization (Yang et al. 1993; Okay 1994; Zhang et al. 1994, 1995; Hiramatsu et al. 1995; Hacker et al. 1997). The Maowu ultramafic body is enclosed in a coesite-bearing gneiss (Sobolev 1994; Xue et al. 1996), and consists of layered harzburgite, garnet ortho- and clinopyroxenite, eclogite, and coesite-bearing omphacitite. These ultramafic rocks contain garnet, enstatite, diopside, forsterite, and subordinate clinohumite, magnesite, chromite, rutile, monazite-(Ce), and apatite. Observed parageneses and compositions of minerals suggest peak metamorphism at about 700 ± 50 °C and 4.0–6.0 GPa for the Maowu harzburgite and garnet orthopyroxenites (Liou and Zhang 1998).

Exsolution lamellae of ilmenite, chromite, kirschsteinite, and symplectic intergrowths of augite and magnetite have been reported in terrestrial and chondritic olivines (Moseley 1981; Ashworth 1979; Drury and van Roermund 1988; Green and Gueguen 1983; Arai 1978; Mikouchi et al. 1995); however, the most common platelets in olivine are ilmenite and Cr-spinel. For example, Dobrzhinetskaya et al. (1996) suggested recently that olivine with $FeTiO_3$ lamellae from the Alpe Arami peridotite, central Alps, were derived from mantle transition depths of 300–400 km. Similar lamellae in olivines have been described in a lherzolite from Rongcheng in the Sulu region, but without the implication of especially deep P-T conditions (Hacker et al. 1997).

Magnetite lamellae in olivine have been described in a few mafic-ultramafic rocks and are intergrown with other silicate phases in symplectites. These include: (1) Shangla ultramafic body in the Indus suture zone of Pakistan (Arif and Jan 1993); (2) intergrowths of augite and magnetite from several mafic-ultramafic complexes (Moseley 1984); (3) intergrowths of orthopyroxene and magnetite in basaltic olivine (Haggerty and Baker 1967), Precambrian cumulates (Goode 1974), and a layered mafic intrusion (Ambler and Ashley 1977); and (4) maghemite intergrowth with olivine in stratospheric interplanetary dust particles (Rietmeijer 1996). These lamellae have been interpreted as the result of oxidation of olivine or as a decomposition product of Fe^{3+}-olivines produced during cooling of mafic-ultramafic rocks.

However, magnetite exsolution lamellae in olivines from UHP metamorphic ultramafic rocks have not been reported previously. Detailed petrographic examination of the Maowu harzburgite and garnet pyroxenite has revealed ubiquitous occurrences of submicrometer-size thin lamellae in coarse-grained

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minerals, including magnetite lamellae in olivine and Ti clinohumite, and monazite lamellae in apatite; which are described below. The crystal structures of the thin lamellae and the intergrowth relationships with their host minerals were determined using a newly developed synchrotron X-ray diffraction (XRD) microprobe technique. These lamellae in UHP olivines are characterized by lack of associated silicate phases, and their genesis apparently resulted from decompression instead of thermal relaxation based on the established *P-T* path of UHP rocks from the Dabie terrane.

SAMPLE DESCRIPTION

The Maowu harzburgite consists of coarse-grained forsterite (60–70 vol%) and enstatite (25–35 vol%), with minor Ti clinohumite, garnet, magnesite and chlorite; its texture is granoblastic with ~120° triple junctions. Enstatite contains olivine inclusions. Garnet occurs as subrounded inclusions 0.07 mm in diameter in both olivine and orthopyroxene, and as separate grains in the matrix (0.3–0.45 mm). Chromite was identified as inclusions in olivine and enstatite, and rarely as euhedral intergranular grains ranging from 0.2 to 0.4 mm in diameter. Electron microprobe analyses of minerals from a harzburgite and orthopyroxenite are listed in Table 1. Olivine (Fo_{93}) grains range from 0.9 to 4.5 mm in size (most grains are 2–3 mm) and are compositionally homogeneous. The intergranular garnet contains considerable amounts of Cr_2O_3 (>2 wt%) and a high pyrope content (>70 mol%); enstatite is characterized by low Al_2O_3 (<0.1 wt%) and high MgO contents. The harzburgite has been subjected to UHP metamorphism at 750 °C and 6 GPa based on Mg-Fe and Al partitioning between garnet and orthopyroxene (for details, see Liou and Zhang 1998).

TABLE 1. Compositions of minerals in harzburgite

	1	2	3	4	5
SiO_2	41.06	0.05	37.99	41.47	58.80
TiO_2	0.01	0.03	0.67	0.00	0.02
Cr_2O_3	0.01	44.88	n.d	2.66	0.05
Al_2O_3	0.01	3.61	n.d	21.46	0.09
FeO	6.93	42.30	7.47	11.46	4.85
MnO	0.04	0.48	n.d	0.60	0.09
MgO	51.68	4.62	52.22	21.02	36.91
CaO	0.04	0.01	n.d	1.20	0.03
Na_2O	0.04	0.25	n.d	0.00	0.00
K_2O	0.03	0.01	n.d	0.00	0.00
NiO	0.41	n.d.	n.d	n.d	n.d
Total	100.06	96.24	98.35	99.87	100.84
			Formula proportions of cations		
Si	0.994	0.002	4.049	2.986	1.993
Ti	0.000	0.001	0.054	0.000	0.001
Cr	0.000	1.284	–	0.151	0.001
Al	0.000	0.154	–	1.822	0.004
Fe^{3+}	–	0.574	–	–	–
Fe	0.140	0.706	0.666	0.690	0.000
Mn	0.001	0.012	–	0.029	0.137
Mg	1.864	0.249	8.298	2.256	0.003
Ca	0.001	0.000	–	0.093	1.865
Na	0.002	0.018	–	0.000	0.001
K	0.001	0.000	–	0.000	0.000
Ni	0.004	–	–	–	0.000
Total	3.007	3.000	13.067	8.027	4.004

1. Olvine, 4 oxygen atoms.
2. Cr Spinel, 4 oxygen atoms.
3. Clinohumite, 17 oxygen atoms (includes 1.01 wt% F).
4. Garnet, 12 oxygen atoms (intergranular).
5. Orthopyroxene, 6 oxygen atoms.
n.d. = not detected.

Light-brown magnetite lamellae are common in the olivine grains except along their margins. The magnetite lamellae are 3–75 μm long, 2–30 μm wide, and 0.5–2 μm thick (Figs. 1 and 2), and range in abundance from nil to ~2 vol% based on backscattered electron images. The volume percent of lamellae inferred from imaging is correct only if the cross section investigated is perpendicular to the lamellae, which is not true in most cases. Therefore, if we assume that the average angle between the lamellae and the cross section is 45°, then the magnetite lamellae will range in abundance from nil to 1.5 vol%. These thin magnetite lamellae contain mainly Fe, with 2–3 wt% MgO. One typical analysis shows 2.50 wt% MgO, 0.73 wt% NiO, and 0.45 wt% SiO_2; minor MgO and SiO_2 may reflect excitation from the host olivine. Compositional profiles for MgO, FeO, and SiO_2 in host olivine and magnetite lamellae (Fig. 3) show sharp contacts between the host and lamellae; intermediate compositions indicate mixed analyses of olivine and magnetite.

Ti clinohumite crystals, 0.1 to 1.5 mm in size occur in harzburgite and orthopyroxenite. These contain 0.2–1.6 wt% TiO_2 (some up to 3.16 wt%, Okay 1994) and also have abundant exsolution lamellae of black to deep-brown magnetite. The lamellae are restricted to cores and mantles of Ti clinohumite crystals, and show distinct crystallographic orientation (Fig. 4). The replacement of Ti clinohumite by ilmenite + olivine was not observed.

CRYSTAL STRUCTURE

Cell parameters and crystallographic orientations of magnetite lamellae and host olivine were determined employing energy dispersive X-ray diffraction (EDXD) at the superconducting wiggler beam line X17C using the synchrotron light source of the Brookhaven National Laboratory. A polished thin section of the harzburgite was mounted on a three-circle diffractometer, and the lamellae (1 × 3 × 30 μm in size) structure was analyzed in situ. Thirty diffraction peaks of magnetite lamellae and 110 diffraction peaks of the host olivine were collected. The *d*-spacings and orientation matrix of the host and lamellae were obtained. The host olivine has orthorhombic cell parameters: $a = 6.007(5)$ Å, $b = 10.249(8)$ Å, and $c = 4.757(2)$ Å, $V = 239.42(18)$ $Å^3$. The magnetite is cubic: $a =$

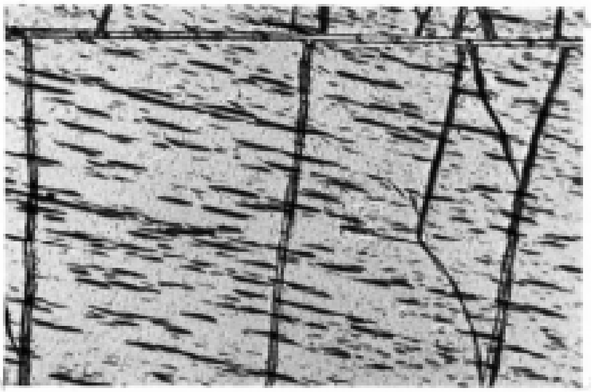

FIGURE 1. Photomicrograph showing magnetite lamellae in olivine from harzburgite. Irregular thin fractures are filled with serpentine + minute grains of magnetite (plane polarized light, width of view = 0.39 mm).

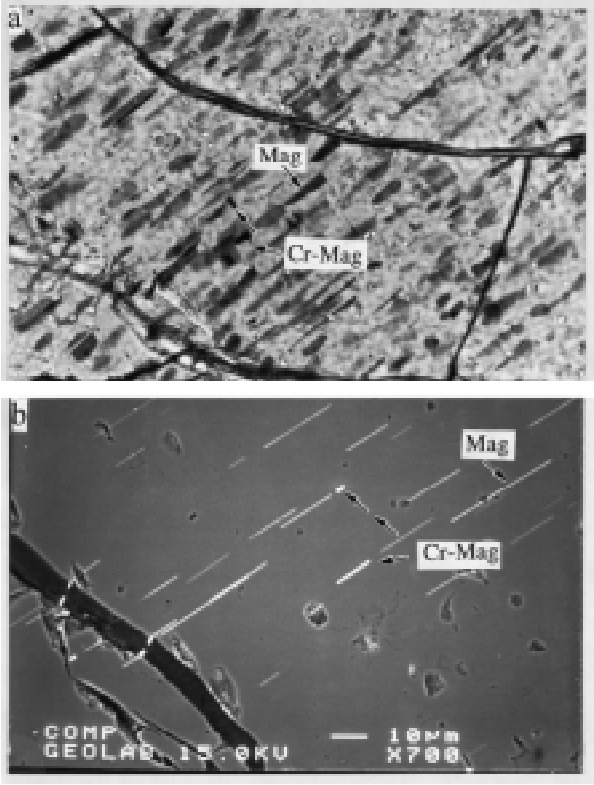

FIGURE 2. (a) Magnetite (Mag) lamellae exsolved from harzburgitic olivine and crosscut by fractures filled with serpentine and minute grains of magnetite. Cr-bearing magnetite (Cr-Mag) lamellae also occur (plane polarized light, width of view = 0.39 mm). (**b**) Back-scattered electron image of a small area from (**a**). Some lamellae beneath the surface do not appear in this image. The rods marked by arrows are the same as those in Figure 2a, (scale bar = 10 µm).

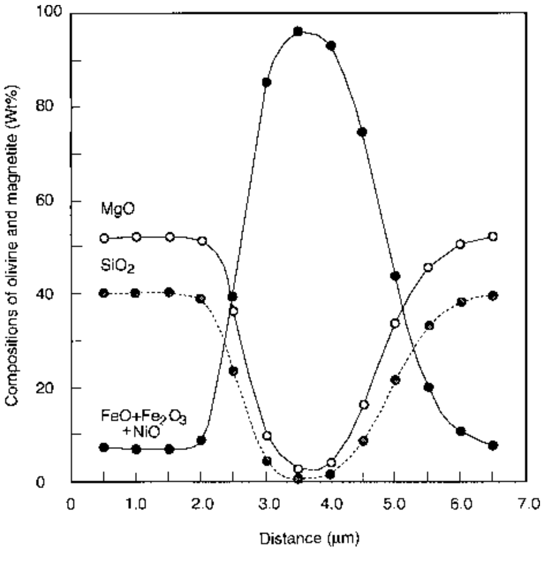

FIGURE 3. Compositional profile from host olivine through magnetite lamellae to host olivine (analysis interval = 50 µm). Two thirds of total iron is expressed as Fe_2O_3 for magnetite and analytical totals are normalized to 100%.

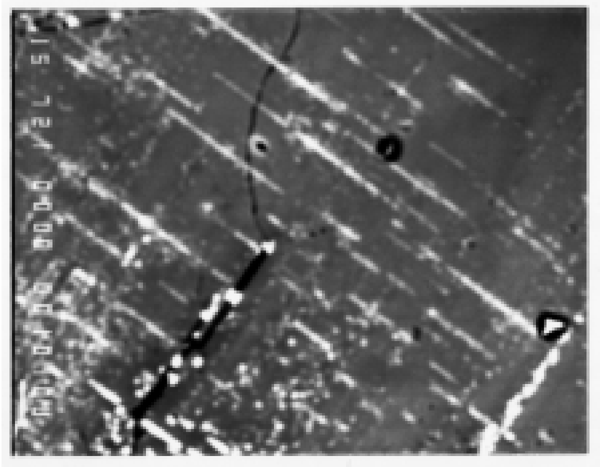

FIGURE 4. Backscattered electron image showing oriented magnetite lamellae (white) in clinohumite from a garnet-bearing orthopyroxenite, (scale bar = 10 µm).

8.411(2) Å, and V= 594.95(38) Å^3. The topotaxial intergrowth between olivine and magnetite was identified as: $[220]_{Mag} \,||\, [200]_{Ol}$, $[111]_{Mag} \,||\, [3\bar{3}\bar{1}]_{Ol}$, $[11\bar{1}]_{Mag} \,||\, [331]_{Ol}$, $[242]_{Mag} \,||\, [2\bar{2}0]_{Ol}$. Such a topotaxial intergrowth and even distribution of magnetite lamellae in the host olivine suggest an exsolution origin as described below.

DISCUSSION

Several interpretations have been proposed for the origin of oriented lamellae (or rods) of $FeTiO_3$ and chromite (including Cr spinel) in olivine. The ilmenite rods in olivine from the Alpe Arami peridotite massif, with three previously unknown crystal structures, were interpreted to possess the high-pressure perovskite structures of ilmenite; the originally unexsolved olivine with high TiO_2 contents (7000–20 000 ppm) may have been the β-phase (Dobrzhinetskaya et al. 1996). However, Hacker et al. (1997) have shown that the volume of ilmenite rods is much lower than originally reported. Reinvestigation of the $FeTiO_3$ rods in olivine by TEM (Risold et al. 1997) and experimental solubility data of TiO_2 in olivine under UHP conditions (Ulmer and Trommsdorff 1997), together with early microfolded amphibole inclusions in olivine (Risold and Trommosdorff 1996), all suggest that these rods formed during decompression and cooling from 3 GPa, 800 °C. Similar olivine with oriented ilmenite rods from the Rongcheng peridotite contains only 269 ± 188 ppm TiO_2 and may have formed at ~120 km depth, which corresponds to 4 GPa (or higher) rather than >300 km as suggested from Al partitioning between enstatite and garnet (Zhang et al. 1994; Hacker et al. 1997). Chromian spinel lamellae in olivine from the Iwanai-dake peridotite, Japan, have been ascribed to exsolution during annealing (Arai 1978). Some chromite inclusions and exsolved lamellae in forsterite from upper mantle samples may have formed as a result of pressure-released oxidation of Cr^{2+} originally present in the olivine or in the UHP spinel analogue (Burns 1975).

Formation of magnetite or symplectic intergrowths of magnetite with silica (or other silicate minerals) in olivine from igneous rocks has been attributed to (1) serpentinization (Arif and Jan 1993); (2) decomposition of Fe^{3+}-bearing, high-T olivine during cooling (Moseley 1984); or (3) oxidation (Haggerty and Baker 1967; Champness 1970; Putnis 1979; Rietmeijer 1996). Among these hypotheses, oxidation has been used most frequently to explain such intergrowths in olivine from basalt and mafic-ultramafic cumulate complexes. Experimental studies of olivine oxidation have shown that an oriented hematite- and/or magnetite-like phase and amorphous silica precipitated in Mg-Fe olivine at 500–800 °C, whereas coarser Fe-oxide and a more ordered silica phase were produced at 1000 °C (Champness 1970). In both natural and experimental samples, the magnetite lamellae invariably are associated with a silica or silicate mineral, such as orthopyroxene and rare andradite-rich garnet (Champness 1970; Blanchard and Cunningham 1974; Putnis 1979; Rietmeijer 1996). For such oxidation processes, a topotaxial relationship between magnetite lamellae and the host olivine is not required. In fact, in well-documented natural occurrences resulting from oxidation, magnetite occurs as vermicular or cellular intergrowths with orthopyroxene or olivine (Goode 1974; Rietmeijer 1996), and shows random dispersal in olivine (Rietmeijer 1996; Haggerty 1981). The symplectic exsolution of magnetite and augite in olivine from the Rhum pluton, the Cuillin and Bushveld complexes, and the Skaergaard intrusion has been interpreted as resulting from decomposition of a high-*T*, Fe^{3+}-bearing olivine during cooling (Moseley 1984). These olivines have a wide compositional range ($Fo_{86–0}$) and contain Fe^{3+}; the reaction of $3Fe^{3+}_{4/3}SiO_4 + Fe_2SiO_4 + 4X_2SiO_4 = 2Fe_3O_4 + 4X_2Si_2O_6$ (where X = Ca, Mg, Fe) was attributed to such intergrowths. The silicate product phase constitutes twice the amount of magnetite produced in the reaction, which is roughly consistent with estimates from TEM micrographs.

The host olivine with magnetite exsolution lamellae in the Maowu harzburgite described in the previous sections is characterized by: (1) a forsterite-rich (Fo_{93}) composition; (2) a topotaxial relationship with coarser-grained oriented magnetite lamellae; (3) an even distribution of magnetite lamellae; and (4) no accompanying silicate product. Such magnetite lamellae are confined to the core and mantle of host olivines with homogeneous compositions. In addition, these lamellae are crosscut by very thin fractures that are filled with fine-grained serpentine and randomly oriented magnetite. Such a relation suggests that these lamellae formed earlier than the brittle deformation and incipient serpentinization of olivine. These early exsolution features cannot be explained by serpentinization because serpentinization was initiated along grain boundaries and fractures, and the resulting magnetite dust is distributed randomly. The following four hypotheses could explain the observed features:

(1) Oxidation of olivine according to a reaction such as:

$$(Mg_{.925}Fe_{.075})_2SiO_4 + 0.025O_2 =$$

$$0.05Fe_3O_4 + 0.85Mg_2SiO_4 + 0.15MgSiO_3.$$

The reaction was written based on the composition of analyzed Maowu olivine. If so, the product magnetite should be accompanied by three times more enstatite or by 1.5 times more quartz based on another possible reaction:

$$3Fe_2SiO_4 \text{ (olivine)} + 0.5O_2 =$$

$$Fe_3O_4 + 1.5\ Fe_2SiO_4 + 1.5\ SiO_2.$$

Such an oxidation would be expected to have occurred along the margins. This mechanism was suggested to explain the formation of observed magnetite lamellae in orthopyroxene for which an analogous oxidation reaction can be written (e.g., Morse 1975). In the present case, the lack of a silica phase could be due to non-stoichiometry for the primary olivine as suggested for orthropyroxene (Dymek and Gromet 1984) or to Si leaching by hydrothermal fluid at high temperatures (P. Ulmer personal communication 1998).

(2) Decomposition of high-*P*, Fe^{3+}-bearing olivine through reactions such as:

$0.1(Fe^{2+}Fe_2{}^{3+}Si_2O_8) + 1.8M_2SiO_4 \rightarrow 0.1Fe_3O_4 + 1.7M_2SiO_4 + 0.2MSiO_3 + 0.1SiO_2$ where M = Mg/(Mg+Fe) = 0.93. As described above, the Maowu harzburgite and associated mafic-ultramafic rocks experienced UHP metamorphism at 6 GPa, and the original olivine may have been a homogeneous single phase with a substantial amount of Fe^{3+}. Exsolution of magnetite from olivine may have occurred during decompression. Laihunite (or ferrifaylite) from iron-rich rocks has been reported to have a nearly stoichiometric formula $Fe^{2+}Fe_2{}^{3+}(SiO_4)_2$ defined by the $[Fe_4(SiO_4)_2\text{-}(Fe^{2+}Fe^{3+})_3(SiO_4)_2]$ system (Ferrifayalite Research Group 1976; Zhang et al 1981). Laihunite from Hebei, China, which was heated at 400, 600, 900, and 1100 °C at 1 atm for 4 h, has the following run products: no change at 400 °C; amorphous SiO_2 + fayalite + laihunite appeared at 600 °C; and amorphous SiO_2 + magnetite + Fe_2O_3 at 900 °C (Zhang et al. 1981). Such reconnaissance experiments on laihunite indicate that magnetite could be a decomposition product of Fe^{3+}-bearing olivine but one or more additional phases (quartz and/or orthopyroxene) should accompany the magnetite.

(3) Exsolution of magnetite from a very high-P olivine phase such as a spinel solid solution $[Fe_3O_4\text{-}(Fe,Mg)_2SiO_4]$. Although Fe^{3+} and Cr^{3+} prefer octahedral sites, Fe^{3+} may have been accommodated in tetrahedral sites by the substitution of $2Fe^{3+}$ for ${}^{VI}(Mg, Fe){}^{IV}Si$ at high temperature and high pressure. Olivine grains contain up to 1.5 vol% of the magnetite lamellae, which implies that the original olivine contained 1.0–1.5 wt% Fe_2O_3. Such a high Fe_2O_3 content is difficult to accommodate in the forsterite structure at low pressure. If the original "olivine" was wadsleyite (β phase) with a distorted-spinel structure, it would readily accommodate Fe_3O_4 as a spinel solid solution along a binary join $Fe_3O_4\text{-}(Fe,Mg)_2SiO_4$. The "β phase" is stable at pressure exceeding 13 GPa, according to experimental studies of the olivine-spinel transformation in the system $Mg_2SiO_4\text{-}Fe_2SiO_4$ for forsterite (Fo > 90%) (Ito and Takahashi 1989; Akaogi et al. 1989; Katsura and Ito 1989). It is possible that, during decompression of the Maowu ultramafic body, transformation of the wadsleyite (β phase) to olivine occurred, and the excess Fe_3O_4 component exsolved to form the magnetite lamellae. The observed topotaxial relations between the magnetite lamellae and olivine are consistent with such an exsolution process. Ross et al. (1992) synthesized a spinelloid phase ($Fe_{5.2}Si_{0.8}O_8\text{-}Fa_{40}\text{-}Mt_{60}$) at 1200 °C, 7 GPa. More recent experiments in the system $Fe_2SiO_4\text{-}Fe_3O_4$ by Woodland and Angel (1998) yielded a new spinelloid phase ($Fe_{2.45}Si_{0.55}O_4$) with the wadsleyite structure, which suggests that significant amounts of Fe^{3+} can be incorporated into a wadsleyite-type structure through the substitution $2Fe^{3+} = Fe^{2+} + Si^{4+}$. The dif-

ference in the stability of Fe^{3+}-substituted wadsleyite (5.0 to 6.0 GPa at 1100–1200 °C) and $(Mg,Fe)_2SiO_4$ wadsleyite (i.e., 12 GPa at 1200 °C) suggests that the addition of Fe^{3+} could act to stabilize $(Mg,Fe)_2SiO_4$ wadsleyites to lower pressures compared with the Fe^{3+} free system; such difference could affect the depth of the "410 km" seismic discontinuity in the mantle (Woodland and Angel 1998). These experimental studies provide theoretical evidence for the third hypothesis, and the incorporation of Fe^{3+} could lower the stability of Maowu olivine $(Mg,Fe)_2SiO_4$ to pressures <12 GPa. However, the exact pressure drop is unknown.

(4) The Maowu harzburgite was metamorphosed at UHP conditions at about 750 °C and 6.0 GPa based on the Mg-Fe and Al partitioning between garnet and pyroxene (Liou and Zhang 1998). These *P-T* estimates are very close to the experimentally calibrated equilibrium points at 720–740 °C at 7 GPa and 800–820 °C at 8 GPa for the reaction:

$$Mg_7Si_2(OH)_6 + 3MgSiO_3 = 5Mg_2SiO_4 + 3H_2O$$

phase A + enstatite = forsterite + vapor

(Luth 1995; Pawley and Wood 1996). In the pyrope-bearing portion of the $MgO\text{-}Al_2O_3\text{-}SiO_2\text{-}H_2O$ system, the assemblage containing phase A reacts to form clinohumite between 740–760 °C at 8 GPa (Luth 1995). If the peak metamorphism of the Maowu harzburgite took place at a slightly higher pressure than the above estimates indicate, the paragenesis of phase A and enstatite would be expected. During decompression, the breakdown of phase A and enstatite to olivine and water would occur, and the phase A would expel Fe, Cr, and other impurities without formation of additional accompanying silicates (Ulmer, personal communication, 1998). Although phase A is a potential precursor to the Maowu olivine suggesting a high-pressure origin at about 200 km, this hypothesis is difficult to prove because it does not explain the process of magnetite exsolution from olivine.

Ti clinohumites have been described in kimberlite, garnet peridotite, serpentinite (Evans and Trommsdorff 1978; Dymek et al. 1988; Deer et al. 1992; Gaspar 1992), and recently in UHP ultramafic rocks from Sulu-Dabie terrane (Yang et al. 1993; Okay 1994; Zhang et al. 1994, 1995). This phase, as a dense hydrous magnesium silicate (DHMS), has been documented to be stable to depths of 150–300 km (5–10 GPa) and temperatures of 700–1100 °C (Ahrens 1989; Thompson 1992). Ti-bearing, F-free clinohumite in Bixiling ultramafic layers in the Dabie UHP terrane was replaced by a symplectite of Mg-rich ilmenite and olivine along its margins during decompression (Zhang et al. 1995). However, Ti clinohumites in Maowu harzburgite and orthopyroxenite contain abundant magnetite lamellae (see Fig. 4) without replacement by ilmenite and olivine. The mechanism of magnetite exsolution in Ti clinohumite is not known.

In summary, oxidation or decomposition during cooling have been proposed as the explanation for the occurrence of some symplectitic intergrowths of magnetite and orthopyroxene or augite in olivine from basalt and mafic-ultramafic complexes. The evolution and *P-T* path of these igneous rocks are different from that of UHP metamorphic ultramafic rocks. For the former, a decrease in temperature is the most important factor during exsolution; for the later, the pressure drop is more important during exhumation of the UHP slab. Lack of an accompanying silicate phase and the topotaxial relationship between magnetite and host olivine cannot be accounted for by the oxidation hypothesis. Decomposition of an original Fe^{3+}-bearing olivine during decompression is possible, however, inclusion of an excess silicate phase in olivine remains to be identified.

The third hypothesis, that the precursor material was a high-pressure phase having a composition lying along the Fe_3O_4-$(Fe,Mg)_2SiO_4$ solid-solution series, may explain the formation of magnetite exsolution lamellae in olivine. This suggestion is consistent with the initial result of experimental synthesis of a spinelloid phase in the Fe_2SiO_4-Fe_3O_4 system by Woodland and Angel (1998). On the other hand, at present, no experimental study on Fe_3O_4-Mg_2SiO_4 system confirms the spinel solid solution $[Fe_3O_4\text{-}(Fe,Mg)_2SiO_4]$ model at high pressures. Nevertheless, the harzburgitic olivine from the Maowu UHP body was formed at pressures greater than 6 GPa (Liou and Zhang 1998), consistent with the occurrence of Cr_2O_3-rich pyrope and Al_2O_3-poor enstatite. In addition, phase A is also a potential high-pressure precursor for the Maowu olivine and magnetite lamellae.

As stated above, mechanisms for exsolution of magnetite lamellae in both olivine and clinohumite are poorly known; each lamellae-bearing host mineral preserves information on the composition and *P-T* condition of the precursor unexsolved phase and the inferred *P-T* path during decompression. Compositional and structural characterization of all lamellae-host mineral pairs will provide a comprehensive petrogenetic record. Unfortunately, studies on exsolution in UHP metamorphic minerals are just beginning, hence only limited experimental data are available. It is possible that the Maowu ultramafic body reached depths of at least ~180 km, where it was entrained in deeply subducted supracrustal rocks and was subsequently exhumed to the surface. The Maowu ultramafic body is characterized by the assemblage of Ti clinohumite, magnesite ($MgCO_3$), monazite, and REE-rich apatite $[Ca_5(PO_4)_3(F,OH,Cl]$. Ubiquitous occurrences of exsolved lamellae in some UHP minerals require further experimental studies of solubilities of trace elements and the effect of pressure on exsolution in olivine, Ti clinohumite, and apatite at extreme high pressures. Such studies are essential for a better understanding of geochemical cycling, mantle dynamics, and metasomatism at mantle depths.

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REFERENCES CITED

Ahrens, T.J. (1989) Water storage in the mantle. Nature, 342, 122–123.
Akaogi, M., Ito, M., and Navrotsky, A. (1989) olivine-modified spinel-spinel transi-

tions in the system Mg_2SiO_4-Fe_2SiO_4 calorimetric measurements, thermochemical calculation, and geophysical application. Journal of Geophysical Research, 94, 15671–1568.

Ambler, E.P. and Ashley, P.M. (1977) Vermicular orthopyroxene-magnetite symplectites from the Wateranga Layered mafic intrusion, Queensland, Australia. Lithos, 10, 163–172.

Arai, S. (1978) Chromian spinel lamellae in olivine from the Iwanai-dake peridotite mass, Hokkaido, Japan. Earth and Planetary Science Letters, 39, 267–273.

Arif, M. and Jan, M.Q. (1993) Chemistry of chromite and associated phases from the Shangla ultramafic body in the Indus suture zone of Pakistan. In P.J. Treloar and M.P. Searle, Eds., Himalayan Tectonics. Geological Society Special Publications, 74, 101–112.

Ashworth, J.R. (1979) Two kinds of exsolution in chondritic olivine. Mineralogical Magazine, 43, 535–538.

Blanchard, M.B. and Cunningham, G.C. (1974) Artificial meteor ablation studies: Olivine. Journal Geophysical Research, 79, 3973–90.

Burns, R.G. (1975) Crystal field effects in chromium and its partitioning in the mantle. Geochimica et Cosmochimica Acta, 39, 857–864.

Champness, P.E. (1970) Nucleation and growth of iron oxides in olivine $(Mg,Fe)_2SiO_4$. Mineralogical Magazine, 37, 790–800.

Deer, W.A., Howie, R.A., and Zussman, J. (1992) An introduction to the rock-forming minerals. 2nd edition, 696 p. Longman Scientific & Technical, Harlow, Essex, England.

Dobrzhinetskaya, L., Green, H.W., and Wang, S. (1996) Alpe Arami: a peridotite massif from depths of more than 300 kilometers. Science, 271, 1841–1846.

Drury, M.R. and van Roermund, H.L.M. (1988) Metasomatic origin for Fe-Tirich multiphase inclusions in olivine from kimberlite xenoliths. Geology, 16, 1035–1038.

Dymek, R.F. and Gromet, L.P. (1984) Nature and origin of orthopyroxene megacrysts for the St-Urbain anorthosite massif, Quebec. Canadian Mineralogist, 22, 297–326.

Dymek, R.F., Boak, J.I., and Brothers, S.C. (1988) Titanian chondrodite- and titanian clinohumite-bearing metadunite from the 3800 Ma Isua supracrustal belt, West Greenland: Chemistry, petrology, and origin. American Mineralogist, 73, 547–558.

Evans, B.W. and Trommsdorff, V. (1978) Petrogenesis of garnet lherzolite, Cima di Gagnone, Lepontine Alps. Earth and Planetary Science Letters, 40, 415–424.

Ferrifayalite Research Group, Department of Geology of Beijing University, Institute of Geology and Mineral Resources of the Chinese Academy of Geological Sciences (1976) Ferrifayalite and its crystal structure. Acta Geologica Sinica, 2, 160–175 (in Chinese).

Gaspar, J.C. (1992) Titanian clinohumite in the carbonatites of the Jacupiranga complex, Brazil: mineral chemistry and comparison with titanian clinohumite from other environments. American Mineralogist, 77, 168–178.

Goode, A.D.T. (1974) Oxidation of natural olivines. Nature, 248, 500–501.

Green, H.W. II and Gueguen, Y. (1983) Deformation of peridotite in the mantle and extraction by kimberlite: A case history documented by fluid and solid precipitates in olivine. Tectonopysics, 92, 71–92.

Hacker, B.R., Sharp, T., Zhang, R.Y., Liou, J.G., and Hervig, R.L. (1997) Low-Ti olivines contradict a transition-zone origin for ultrahigh-pressure lherzolites. Science, 278, 702–704.

Haggerty, S.E. (1981) Opaque mineral oxides in terrestrial igneous rocks. In Mineralogical Society of American Reviews in Mineralogy, 3, 101–300.

Haggerty, S.E. and Baker, I. (1967) The alteration of olivine in basaltic and associated lavas. Part I. High-temperature alteration. Contributions to Mineralogy and Petrology, 16, 233–257.

Hiramatsu, N., Banno, S., Hirajima, T., and Cong, B. (1995) Ultrahigh-pressure garnet lherzolite from Chijiadian, Rongcheng County, in the Su-Lu region of eastern China. The Island Arcs, 4, 324–333.

Katsura, T. and Ito, E. (1989) The system Mg_2SiO_4-Fe_2SiO_4 at high pressure and temperatures: precise determination of stabilities of olivine, modified spinel, and spinel. Journal of Geophysical Research, 94, 15663–15670.

Ito, E. and Takahashi, E. (1989) Postspinel transformations in the system Mg_2SiO_4-Fe_2SiO_4 and some geophysical implications. Journal of Geophysical Research, 94, 10637–10646.

Liou, J.G. and Zhang, R.Y. (1998) Petrogenesis of an ultrahigh-P garnet-bearing ultramafic body from Maowu, Dabie Mountains, east-central China. The Island Arcs, 7, 115–134.

Luth, R.W. (1995) Is phase A relevant to the Earth's mantle? Geochimica et Cosmochimica Acta. 59, 679–682.

Mikouchi, T., Takeda, H., Miyamoto, M., Ohsumi, K., and McKay, G.A. (1995) Exsolution lamellae of kirschsteinite in magnesium-iron olivine from an angrite meteorite. American Mineralogist, 80, 585–592.

Morse, S.A. (1975) Plagioclase lamellae in the hypersthene, Tikkoatokhahk Bay, Labrador. Earth and Planetary Science Letters, 26, 331–336.

Moseley, D. (1981) Ilmenite exsolution in olivine. American Mineralogist, 66, 976–979.

——— (1984) Symplectic exsolution in olivine. American Mineralogist, 69, 139–153.

Okay, A.I. (1994) Sapphirine and Ti clinohumite in ultra-high-pressure garnet-pyroxenite and eclogite from Dabie Shan, China. Contributions to Mineralogy and Petrology, 116, 145–155.

Pawley, A.R. and Wood, B.J. (1996) The low-pressure stability of phase A, $Mg_7Si_2O_8(OH)_6$. Contributions to Mineralogy and Petrology, 124, 90–97.

Putnis, A. (1979) Electron petrography of high-temperature oxidation in olivine from the Rhum layered Intrusion. Mineralogical Magazine, 43, 293–296.

Rietmeijer, F.J.M. (1996) Cellular precipitates of iron oxide in olivine in a stratospheric interplanetary dust particle. Mineralogical Magazine, 60, 877–885.

Risold, A.C. and Trommosdorff, V. (1996) Alpe Arami and Cima di Gagnone garnet peridotites: observations contradicting the Hypothesis of ultra deep origin. Eos, 77, F761.

Risold, A.C., Trommosdorff, V. Reusser, E., and Ulmer, P. (1997) Genesis of $FeTiO_3$ inclusions in garnet peridotites from the central Alps. Abstract supplement No. 1 to TERRA nova, 9 (Fifth IEC, Ascona, 29, p. 28–29).

Ross II, C.R., Armbruster, T., and Canil, D. (1992) Crystal structure refinement of a spinelloid in the system Fe_3O_4-Fe_2SiO_4. American Mineralogist, 77, 507–511.

Sobolev, N.V., Shatsky, V.S., Vavilov, M.A., and Goryainov, S.V. (1994) Zircon from ultrahigh-pressure metamorphic rocks of folded regions as an unique container of inclusions of diamond, coesite and coexisting minerals. Doklady Akademii Nauk, 334, 488–492.

Thompson, A.B. (1992) Water in the Earth's upper mantle. Nature, 358, 295–301.

Ulmer, P. and Trommsdorff, V. (1997) Titanium content of mantle olivine: an experimental study to 10 GPa. Abstract supplement No. 1 to TERRA nova, 9 (Fifth IEC, Ascona, 29, p.39).

Woodland, A.B. and Angel, R.J. (1998) Crystal structure of a new spinelloid with the wadsleyite structure in the system Fe_3O_4-Fe_2SiO_4 and implications for the Earth's mantle. American Mineralogist, 83, 404–408.

Xue, F., Rowley, D.B., and Baker, J. (1996) Refolded syn-ultrahigh-pressure thrust sheets in the south Dabie complex, China: Field evidence and tectonic implications. Geology, 24, 455–458.

Yang, J., Godard, G., Kienast, J.R., Lu, Y., and Sun, J. (1993) Ultrahigh-pressure (60 kbar) magnesite-bearing garnet peridotites from northeastern Jiangsu, China. Journal of Geology, 101, 541–554.

Zhang, R.Y., Liou, J.G., and Cong, B.L. (1994) Petrogenesis of garnet-bearing ultramafic rocks and associated eclogites in the Su-Lu ultrahigh-pressure metamorphic terrane, China. Journal of Metamorphic Geology, 12, 169–186.

——— (1995). Talc-, magnesite- and Ti clinohumite-bearing ultrahigh-pressure mafic and ultramafic complex in the Dabie Mountains, China. Journal of Petrology, 36, 1011–1037.

Zhang, R.Y., Cong, B.L., Ying, Y.P., and Li, J.L. (1981) Ferrifayalite-bearing eulysite from Archaean granulites in Qianan county, Hebei, North China. TMPM Tschermaks Mineralogische und Petrographische Mitteilungen, 28, 167–187.

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