YAMATO-8002: PARTIAL MELTING RESIDUE ON THE "UNIQUE" CHONDRITE PARENT BODY

Hiroko Nagahara*

Geological Institute, University of Tokyo, Hongo 7-chome, Bunkyo-ku, Tokyo, 113

Abstract: Yamato-8002 is classified as a "unique" meteorite having a granoblastic texture with olivine and orthopyroxene and interstitial diopside, plagioclase, spinel, kamacite schreibersite, troilite, and rare graphite. Olivine, orthopyroxene, and clinopyroxene are homogeneous with respect to Mg and Fe, but Ca in pyroxenes are zoned. Plagioclase and spinel are heterogeneous within and among grains. Mineral compositions, particularly plagioclase, spinel and trace elements in all minerals along with oxygen isotopic compositions suggest a genetic link to Y-74063, Y-75274, ALH77081, Y-791493, Acapulco, and Lodran. They were derived from a common precursor through different degrees of partial melting; Y-8002 and Y-75274 have the highest $Mg/(Mg+Fe)$ ratios of olivine, pyroxenes, and spinel, the highest An mol% of plagioclase, and the higest $Cr/(Cr+Al)$ ratio of spinel, indicating formation through the highest degree of partial melting. They were reduced during melting, resulting in crystallization of extremely magnesian minerals. Y-74063, ALH77081, Y-791493, Lodran, and Acapulco possess nearly primary composition of the ferromagnesian minerals excepting that some have lost and some have added partial melts. The partial melting model is supported by the REE patterns; Y-74063 and ALH77081 are flat, Y-8002 and Lodran are heavy REE enriched, and Acapulco is light REE enriched. Positive or negative Eu anomalies are general in these meteorites, suggesting addition or extraction of interstitial melt rich in plagioclase component. The precursor is chondritic with higher $Mg/(Mg+Fe)$ ratio and higher Cr and Mn contents than ordinary chondrites, and is tentatively called a "U(nique)" chondrite. Y-74357, Y-78230, and MAC88177 are probably related to the "U" chondrite. Oxygen isotopic heterogeneity of those meteorites may be due to weathering on the Antarctic ice, and the effect of weathering was excluded by assuming that all oxygen in Fe_2O_3 was from Antarctic ice. The calculated compositions are similar to each other, suggesting their derivation from a common precursor.

1. Introduction

As part of a consortium study on Antarctic "unique" meteorites, Y-8002 was petrologically studied. The "unique" meteorites are characterized by essentially chondritic composition (Fukuoka and Kimura, 1990) and the Fe/Mg ratio of ferromagnesian minerals intermediate between H and E chondrites. Among them, Y-74063 and ALH-77081 have relict chondrules suggesting an origin as a chondrite (Schultz *et al.*, 1982; Kimura, 1987; Yanai and Kojima, 1991), but the others do not have evidence for chondrite origin. The latter have granoblastic texture

---

* Present address: Geophysical Laboratory, Carnegie Institution of Washington, 5251 Broad Branch Road, N.W., Washington, D.C. 20015–1305, U.S.A.

consisting of coarse-grained olivine, orthopyroxene, and Fe-Ni metal and lesser amounts of clinopyroxene, plagioclase, troilite, spinel, and phosphates or phosphide. Although ALH-77081 has relict chondrules, KRACHER (1988) suggested that metals melted at the eutectic with sulfur on the basis of low Ni and Ge concentrations in kamacite.

Genetic relationship among the "unique" meteorites and non-Antarctic unique meteorites, such as Lodran (BILD and WASSON, 1976; FUKUOKA, et al., 1978; PRINZ et al., 1978), Acapulco (PALME et al., 1981), and "primitive achondrites" (PRINZ et al., 1983) has been proposed (SCHULTZ et al., 1982; KALLEMEYN and WASSON, 1985; NAGAHARA and OZAWA, 1986; NAGAHARA et al., 1990; HIROI and TAKEDA, 1991; KIMURA et al., 1991). Although genetic relationship between Lodran and ureilites was proposed (BILD and WASSON, 1976), mineralogy and oxygen isotopic compositions (CLAYTON and MAYEDA, 1978, 1988; CLAYTON et al., 1984; MAYEDA and CLAYTON, 1980, 1989; MAYEDA et al., 1987) distinguish the "unique" meteorites from ureilites.

Y-8002 has previously been briefly described with a possibly paired meteorite, Y-75274, by YANAI et al. (1984). Y-75274 was studied by MATSUMOTO et al. (1983), MORI et al. (1984), and YANAI et al. (1984). The Rb-Sr model age for Y-8002 is 3.3±0.3 Ga (YAMAMOTO et al., 1991), and is different from 4.5±0.1 Ga for the possibly related meteorite Y-74063, suggesting thermal disturbance for Y-8002 after crystallization. On the other hand, Y-74063 does not show thermal plateau with respect to the ^{40}Ar-^{39}Ar age (KANEOKA et al., 1991). These measurements show that the "unique" meteorites have suffered complex thermal history (histories).

In order to investigate the origin of the "unique" meteorites and their genetic relationship to other meteorite groups, petrological study on Y-8002 was carried out, and the results were compared with other "unique" meteorites. On the basis of the results, condition, formation process, and primary material of the "unique" meteorites will be discussed. Those meteorites may be related to "winonaites" (Winona, Pontlyfni, Mount Morris, Y-74025, Y-75300, and Y-75305); however, the relationship was not investigated in this paper mainly because of insufficient published mineralogical data on "winonaites".

2. Analytical Methods

Polished thin section of Y-8002,51.2 (4×7 mm^2) were observed with a JEOL JSM-840 scanning electron microscope equipped with an energy dispersive spectrometer LINK-SYSTEM AN10000 and chemical composition of minerals were obtained with a JEOL JCXA-733 Mk-II electron microprobe at the Geological Institute of University of Tokyo. Operating conditions for the SEM-EDS analysis were an accelerating voltage of 15 kV, sample current of 1 nA, and counting time of 100 s, and those for EPMA analysis were 15 kV, 12 nA and 30 s, respectively. Opaque mineral analysis was at the accelerating voltage of 25 kV, a current of 20 nA and the counting time of 30 s. Standard ZAF corrections were applied for both silicates and opaque minerals. Mapping analysis for elemental distribution on the whole area of the thin section was carried out with an accelerating voltage

of 15 kV, a sample current of 120 nA, counting time of 60 μs, and pixel size of 5 × 5 μm^2.

3. Petrology of Y-8002

3.1. Texture

Y-8002 has granoblastic texture consisting of coarse-grained olivine, orthopyroxene, and kamacite with smaller amounts of plagioclase, clinopyroxene, spinel, troilite, graphite, and schreibersite (Fig. 1a). Olivine and orthopyroxene exhibit almost equant euhedral to subhedral shapes, and are 0.5 to 2 mm in size. Olivine and orthopyroxene have array of tiny kamacite and/or troilite inclusions (Figs. 1b and c), which is characteristic of ureilites and "unique" meteorites (BERKLEY et al., 1980; NAGAHARA and OZAWA, 1986). The blebs may be metal-sulfide melt trapped during crystallization. Orthopyroxene sometimes poikilitically encloses olivine, but the opposite is never the case, suggesting crystallization of olivine before orthopyroxene. The only clinopyroxene grain present in the thin section occurs as an intergrowth with kamacite having round plagioclase inclusions. Kamacite is smaller than olivine and orthopyroxene, 0.5 to 1 mm, and is subrounded to irregular in shape. Kamacite grains often have inclusions of graphite, which has ovoid, notched, or wedged shapes (Fig. 1d). Occurrence of graphite in Y-8002 is suggestive of simultaneous crystallization of graphite and kamacite from melt. Kamacite with abundant graphite coexists with schreibersite, but no schreibersite occurs as an isolated grain. Troilite is small (10–50 μm), subhedral, and often porous. The porous troilite transferred into a vein running through an olivine grain (Fig. 1e), which suggests that troilite melted *in situ* and the melt moved through pre-existing olivine. Only four spinel grains are present in the thin section, which are very small (less than 10 μm) except for one larger grain (50 μm). They are always associated with troilite and do not occur as isolated grains or as inclusions in olivine as is often the case in the related meteorites (*e.g.*, in Y-791493: NAGAHARA and OZAWA, 1986). Plagioclase occupies the interstices of other minerals and is distributed homogeneously over the thin section. There is no visible alignment or lineation of minerals at last in thin section studied, which is the case in Y-791493 (NAGAHARA and OZAWA, 1986). Plagioclase exhibits albite twinning (Fig. 1f).

Modal abundances of minerals obtained by the point counting technique are 32.3% olivine (25.6% by YANAI et al., 1984), 46.4% orthopyroxene (53.9), 1.3% clinopyroxene (2.5), 10.3% plagioclase (11.7), 6.3% Fe-Ni metal (6.0), 0.3% troilite (0.1), 0.4% graphite, 2.8% weathered area which may primarily have been metal, and trace amounts of chromite (0.1) and schreibersite (0.1). The difference of the analytical results between present work and YANAI et al. (1984) may be due to difference in the thin section studied, which have small areas in spite of the large grain size.

Textural and mineralogical characteristics are shown in elemental distribution maps (Fig. 2), where olivine is shown by pink, orthopyroxene by green, and Ca-pyroxene by blue (only one), where other phases shown by black are indistinguishable (Fig. 2a). The distribution of Mg well shows that ferromagnesian minerals are very

Fig. 1. Photomicrographs
showing texture of Y-8002.

(a) Texture of Y-8002. It con-
sists mainly of coarse-
grained olivine, ortho-
pyroxene with cleavages,
and opaque kamacite.
Transmitted light. Width
4 mm.
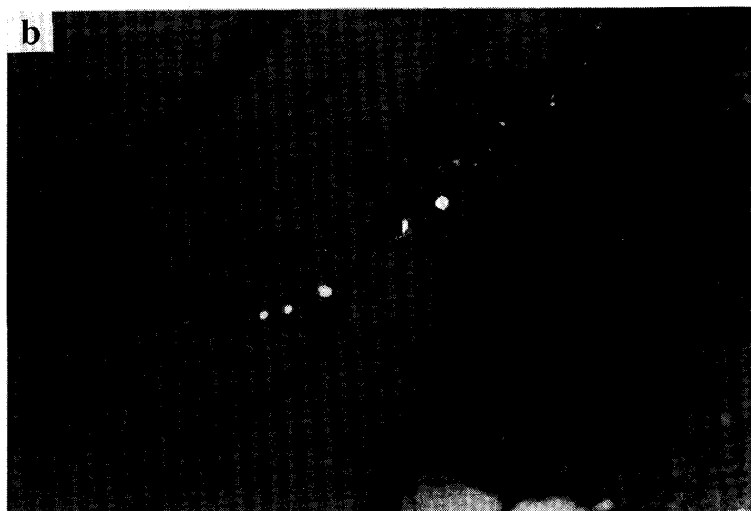
(b) Array of tiny metal and/
or troilite grains (bright
or dark spots) in olivine.
This texture is typically
developed in the "uni-
que" meteorites and
ureilites. Reflected light.
Width 0.6 mm.

(c) Array of tiny metal and/
or troilite (black dots) in
orthopyroxene. The array
develops independently
of crystal structure (direc-
tion of cleavages), sug-
gesting formation dur-
ing crystallization. Trans-
mitted light. Width 0.6
mm.

(d) Graphite inclusions (gray) in kamacite. This kamacite is the grain richest in graphite inclusions. Dark areas are voids. Reflected light. Width 0.6 mm.
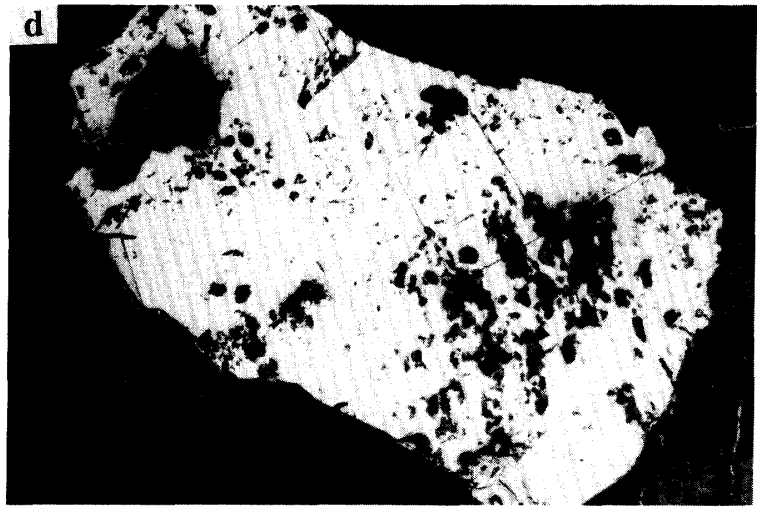
(e) Porous troilite and derived troilite vein running through an olivine grain (dark gray). Grayish material adjacent to troilite is a weathering product. Reflected light. Width 0.6 mm.

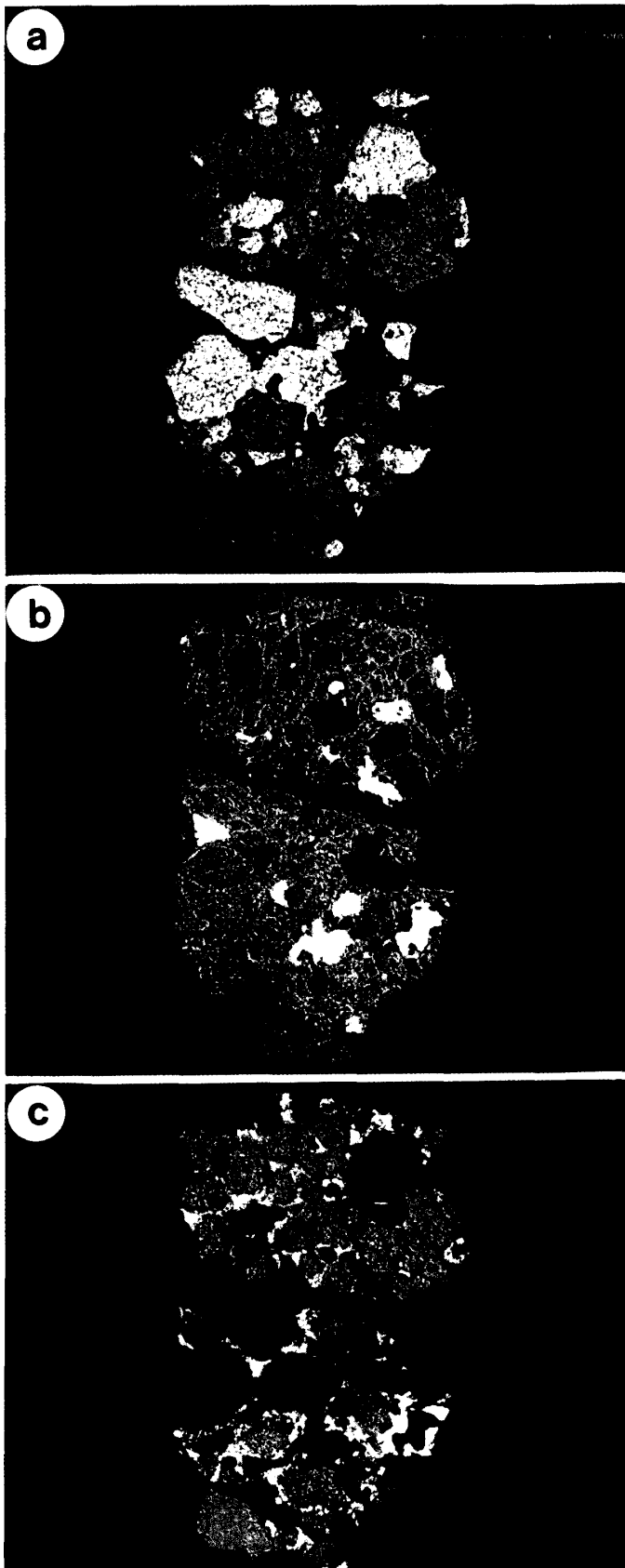(f) Twinning of plagioclase showing fairly high temperature crystallization. Transmitted light. Width 0.4 mm.

Fig. 2. Elemental distribution maps of Y-8002.

(a) Distribution of Mg. Olivine (pink) and orthopyroxene (green) are euhedral, coarse-grained, and highly homogeneous. Only one clinopyroxene grain (blue, in the lower right) is anhedral, surrounding irregular kamacite.

(b) Distribution of Fe. Kamacite (white) is subhedral to anhedral. Pink is weathered metal. Olivine (green) and orthopyroxene (blue) are penetrated by numerous veins. Some of the veins are weathering products, but some are troilite vein derived from porous troilite (see Fig. 1d).

(c) Distribution of Ca. Orthopyroxene is zoned from calcic core (yellow) to less calcic rim (blue). Interstitial plagioclase is also zoned from calcic core (white) to less calcic (pink) rim contacting with other minerals. Clinopyroxene (white) is visible.

homogeneous. Distribution of anhedral metal grains are shown by white color in the map for Fe; pink is weathered material, green is olivine, and blue is orthopyroxene (Fig. 2b). The map for Ca clearly shows the characteristics of the meteorite: orthopyroxene is zoned from Ca-rich core shown by yellow to Ca-poor rim shown by light-blue, plagioclase occupies almost all interstices of other phases forming a network over the meteorite, and is also zoned from calcic core (white) to less calcic rim (pink), and Ca-pyroxene (white) completely surrounds a kamacite grain (black) (Fig. 2c).

3.2. Mineralogy

Olivine is highly homogeneous throughout the thin section studied, Fo being between 96.5 and 97 (Fig. 3). Representative chemical compositions are listed in Table 1. The MnO content of olivine is about 0.30–0.35 wt%, which is lower than those in other related meteorites. Contents of other elements, such as CaO, NiO, and Cr_2O_3 are below 0.1 wt%, and they are homogeneously distributed within and

Table 1. Chemical composition of olivine, orthopyroxene, and clinopyroxene in Y-8002.

	Ol(1)	Ol(2)	Ol(3)	Ol(4)	Opx(5)	Opx(6)	Opx(7)	Cpx(8)	Cpx(9)	Cpx(10)
SiO_2	42.9	43.0	42.9	43.0	59.6	59.6	59.7	55.6	55.0	55.5
TiO_2	0.02	0.04	0.02	0.00	0.22	0.23	0.23	0.81	0.93	0.81
Al_2O_3	0.01	0.01	0.03	0.02	0.33	0.38	0.36	0.96	1.18	0.96
Cr_2O_3	0.02	0.03	0.01	0.02	0.19	0.24	0.22	0.38	0.51	0.37
V_2O_3	0.01	0.00	0.00	0.01	0.01	0.03	0.00	0.04	0.08	0.06
FeO	3.52	3.37	3.38	3.24	2.58	2.38	2.43	1.20	1.29	1.89
MnO	0.33	0.30	0.28	0.30	0.40	0.36	0.38	0.22	0.21	0.24
MgO	54.6	54.9	54.7	54.7	37.0	36.0	35.2	19.0	18.4	19.1
CaO	0.02	0.00	0.03	0.01	1.16	2.21	1.26	21.9	22.6	21.3
NiO	0.01	0.00	0.00	0.02	0.01	0.00	0.00	0.02	0.02	0.11
Na_2O	0.00	0.00	0.00	0.00	0.04	0.07	0.03	0.54	0.58	0.53
Total	101.4	101.7	101.4	101.3	101.5	101.5	99.8	100.7	100.8	100.9
	O=4				O=6					
Si	1.006	1.005	1.006	1.008	1.995	1.999	2.025	1.984	1.969	1.982
Ti	0	0.001	0	0	0.006	0.006	0.006	0.022	0.025	0.022
Al	0	0	0.001	0	0.013	0.015	0.014	0.040	0.050	0.041
Cr	0	0.001	0	0	0.005	0.006	0.006	0.011	0.015	0.011
V	0	0	0	0	0	0.001	0	0.001	0.002	0.002
Fe	0.069	0.066	0.066	0.064	0.072	0.067	0.069	0.036	0.039	0.056
Mn	0.007	0.006	0.006	0.006	0.011	0.010	0.011	0.007	0.006	0.007
Mg	1.909	1.916	1.913	1.913	1.844	1.799	1.780	1.011	0.980	1.014
Ca	0.001	0	0.001	0	0.042	0.079	0.046	0.839	0.867	0.816
Ni	0	0	0	0	0	0	0	0.001	0	0.003
Na	0	0	0	0	0.003	0.005	0.002	0.037	0.040	0.037
Total	2.992	2.995	2.993	2.991	3.991	3.987	3.959	3.989	3.993	3.991
Mg/(Mg+Fe)	0.965	0.967	0.967	0.968	0.962	0.964	0.963	0.966	0.962	0.948
Wo (mol%)					2.15	4.06	2.43	44.5	46.0	43.3

(1) Average of 33 analyses, (2) core of a large crystal, (3) rim of a large crystal, (4) inclusion in orthopyroxene, (5) average of 16 analyses, (6) core of a large crystal, (7) rim of a large crystal, (8) average of 23 analyses, (9) most Wo-rich composition, (10) most Wo-poor composition.

among grains. Olivine is rarely poikilitically enclosed by orthopyroxene, most of which have the same composition as euhedral grains (Table 1).

Orthopyroxene is the most abundant mineral in Y-8002. It has well developed cleavages in one or two directions. The Fe/Mg ratio of orthopyroxene is uniform throughout the thin section (En 96 to 97) (Fig. 3), but the grains are zoned with respect to Ca. The Wo content ranges from 3.7 mol% in the central portion to 1.2% in the outer rim. Calcium is homogeneously distributed in central portion and decreases outwards within 50 μm of the rim, which is well shown by difference in color in Fig. 2c. Orthopyroxene has numerous thin lamellae of clinopyroxene which are visible under an optical microscope. Sometimes the lamellae are as thick as 10 μm in width. The diffusion distance of Ca in orthopyroxene at subsolidus temperatures was, thus, less than 10 μm, indicating that the zonal structure of Ca in orthopyroxene grains was formed during crystallization. Pyroxene compositions are plotted on the quadrilateral with isotherms by LINDSLEY and ANDERSEN (1983), which shows that they cooled from about 1200°C to 800°C assuming that orthopyroxene and clinopyroxene crystallized simultaneously (Fig. 4). Orthopyroxene contains considerable amounts of MnO (0.34–0.44%, average 0.40%), Al_2O_3 (0.28–0.39%, 0.33%), TiO_2 (0.18–0.29%, 0.22%), and Cr_2O_3 (0.16–0.22%, 0.19%) (Table 1). Concentration of these elements roughly increases toward rim of a grain. Contents of other elements are below 0.1 wt%.

The sole clinopyroxene is not so homogeneous as olivine and orthopyroxene with respect to the X_{Mg} (= Mg/(Mg + Fe) mol) ratio, ranging 0.94–0.98 (Fig. 3). It is also heterogeneous as to Ca, the Wo content ranging from 43.2 to 46 mol%. Trace element contents are fairly high: the Al_2O_3 content ranging from 0.8 to
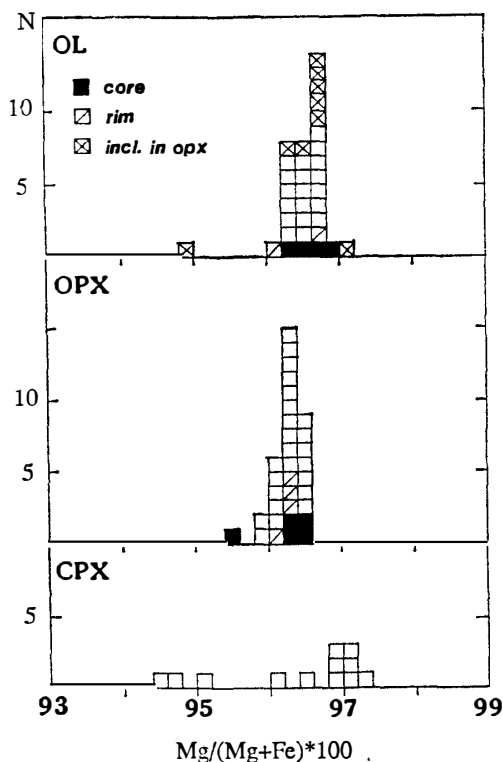

Fig. 3.   Histogram of X_{Mg} (Mg/(Mg+Fe) mol ratio) of olivine, orthopyroxene, and clinopyroxene of Y-8002. Olivine (OL) and orthopyroxene (OPX) are highly homogeneous and there is no compositional difference in occurrence. Clinopyroxene is heterogeneous.

1.0 wt% (average of 23 analyses is 0.96 wt%), TiO_2 from 0.7 to 0.9% (0.81), MnO from 0.14 to 0.25 (0.22), Na_2O from 0.5 to 0.6% (0.54), and Cr_2O_3 from 0.3 to 0.5% (0.38) (Table 1). The composition of clinopyroxene plotted on the LINDSLEY and ANDERSEN (1983) quadrilateral gives temperatures from 1150°C to 1000°C if two pyroxenes crystallized simultaneously (Fig. 4).

Plagioclase is relatively calcic when compared to those in ordinary chondrites, and is zoned from Ca-rich (An_{32}) central portion to Ca-poor margin (An_{28}) with the Or content ranging from 1 to 2 mol% (Fig. 5, Table 2).

Spinel is highly chromian (64 to 69 wt% Cr_2O_3) and low in Al_2O_3 (2–5 wt%); they also have considerable amounts of MnO (1.2–3.3 wt%), V (0.2–0.7 wt% as V_2O_3) and ZnO (0.4 to 1.2 wt%) (Table 2). The spinels are highly heterogeneous from grain to grain. Figure 6 shows the Mg/(Mg+Fe) and Cr/(Cr+Al) ratios of spinel, which shows that each spinel grain has a different composition. The higher the Fe content, the higher the Mn and Zn and the lower the Mg and Cr contents.
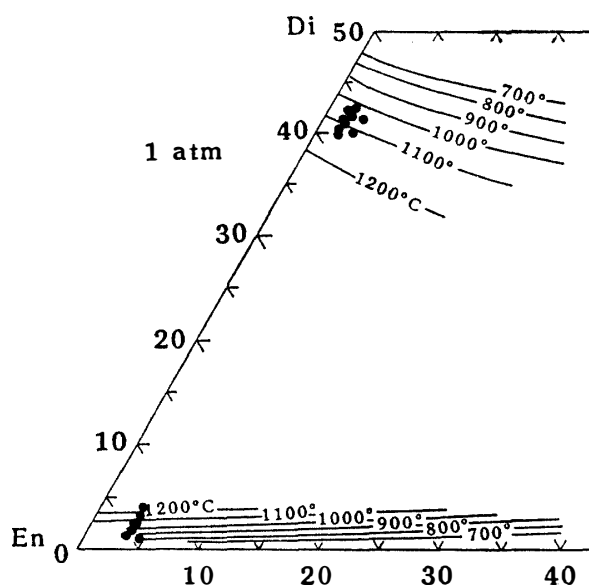

Fig. 4. Composition of pyroxenes in Y-8002 plotted on the LINDSLEY and ANDERSEN (1983) quadrilateral. The most Ca-rich and Ca-poor compositions are shown for both pyroxenes.
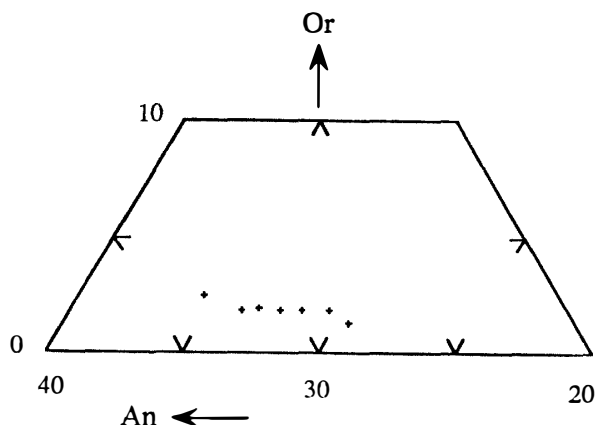

Fig. 5. Histogram showing composition of plagioclase. Plagioclase is heterogeneous within and among grains, although it is difficult to define a grain because of interconnection. One point represents one analysis.

Table 2.   *Chemical composition of plagioclase and spinel in Y-8002.*

	Pl(1)	Pl(2)	Pl(3)		Sp(4)	Sp(5)	Sp(6)	Sp(7)	Sp(8)	Sp(9)
SiO_2	64.0	65.2	63.2							
TiO_2	0.06	0.07	0.05		0.49	0.69	0.68	0.58	0.09	0.09
Al_2O_3	24.7	24.9	24.4		4.36	4.86	4.98	5.44	2.27	2.34
Cr_2O_3	0	0.01	0		67.4	67.9	66.9	64.4	67.2	64.4
V_2O_3	0	0	0.01		0.53	0.53	0.62	0.64	0.47	0.30
FeO	0.16	0.02	0.13		9.64	7.22	8.58	12.7	14.9	15.7
MnO	0.01	0	0.02		1.72	1.23	1.47	1.77	3.32	3.10
MgO	0.02	0	0.01		14.7	16.5	15.6	12.3	10.0	10.5
CaO	5.80	5.55	5.24							
ZnO					0.68	0.49	0.66	0.76	1.04	1.17
Na_2O	7.02	5.99	7.29							
K_2O	0.27	0.34	0.20							
Total	102.0	102.1	100.6		99.5	99.4	99.5	98.5	99.3	97.6
	O=8				O=4					
Si	2.764	2.796	2.771							
Ti	0.002	0.002	0.002		0.012	0.017	0.017	0.015	0.002	0.002
Al	1.259	1.258	1.258		0.169	0.186	0.192	0.217	0.093	0.097
Cr	0	0	0		1.763	1.748	1.734	1.717	1.842	1.796
V	0	0	0		0.014	0.014	0.016	0.017	0.013	0.008
Fe	0.006	0.001	0.005		0.270	0.197	0.235	0.359	0.432	0.465
Mn	0	0	0.001		0.049	0.034	0.041	0.051	0.098	0.093
Mg	0.002	0	0		0.721	0.802	0.761	0.616	0.518	0.554
Ca	0.269	0.255	0.246							
Zn					0.017	0.012	0.016	0.019	0.027	0.030
Na	0.588	0.498	0.619							
K	0.015	0.019	0.011							
Total	4.905	4.829	4.913		3.015	3.010	3.012	3.011	3.025	3.045
Ca/(Ca+Na)	0.314	0.339	0.284	Mg/(Mg+Fe)	0.728	0.803	0.764	0.632	0.545	0.206
Ca/(Ca+Na+K)	0.309	0.330	0.281	Cr/(Cr+Al)	0.913	0.904	0.901	0.888	0.952	0.949

(1) Average of 18 analyses, (2) most Ca-rich composition, (3) most Ca-poor composition, (4) average of 10 analyses, (5) core of grain #1, (6) rim of grain #1, (7) grain #2, (8) grain #3, (9) grain #4.

Aluminum, Ti, and V contents are related to each other, but they are not related to Mg and Fe.

Kamacite is homogeneous in the center of grains with Ni content of 5.0–5.4 wt%, and is zoned within marginal 10 to 20 μm of grains where Ni content decreases outward to about 4.5%. There is no compositional heterogeneity around graphite inclusions. Schreibersite is weekly zoned from Ni-richer core (36 wt%) to Ni-poorer rim (35%) at the contact with kamacite. The Co content is about 0.5 wt% and the Si content is below the detection limit of the EPMA analysis.

3.3. Thermal history

Crystallization sequence of Y-8002 was estimated from the texture. Olivine was the first because of its euhedral to subhedral shape and having no inclusion of other phases, followed by orthopyroxene which is euhedral to subhedral with olivine inclusions. Kamacite, schreibersite, and graphite crystallized next, which is shown by the fact that kamacite is subhedral to anhedral against olivine and orthopyroxene
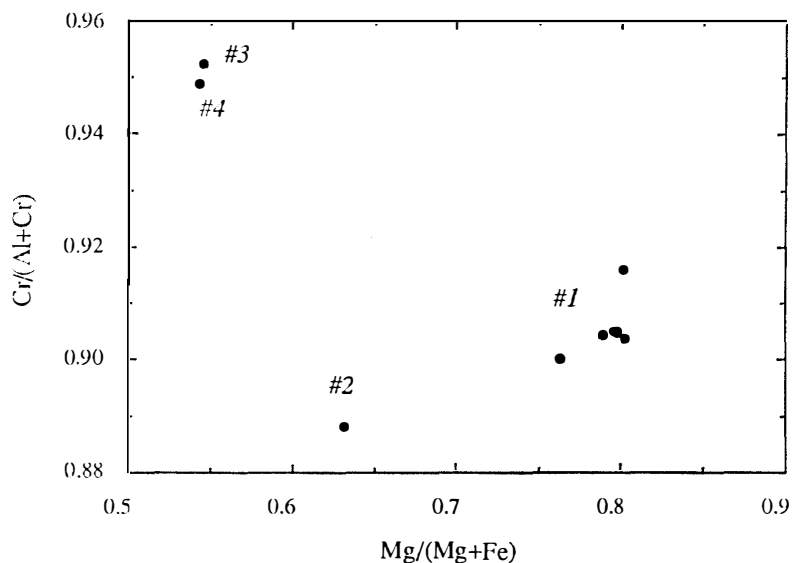
Fig. 6. The relationship between Mg/(Mg+Fe) and Cr/(Al+Cr) ratios of four spinels in Y-8002. They are highly heterogeneous from grain to grain. Grain #1 is coarse (50 μm), but other three are small (<10 μm). All the grains are coexisting with troilite. Compositions are shown in Table 2.

but is entirely enclosed by plagioclase and clinopyroxene. Plagioclase crystallized prior to or simultaneously with clinopyroxene. Crystallization sequence of troilite and chromite is hardly known because of their rare occurrence. During melting, small amount of metal and troilite melts tends to be rounded because of strong surface tension against silicate melt, which would not pervade olivine grains. The porous troilite and veins may have been formed by a shock after solidification. This event may be related to a thermal disturbance shown by the ^{39}Ar-^{40}Ar dating (KANEOKA et al., 1991).

Crystallization temperature of olivine should have been above 1200°C, which is the temperature recorded in pyroxene. Pyroxene thermometry shows that the meteorite cooled from 1200°C to 800°C continuously. Cooling history of Y-8002 below 700°C is hardly known, because of absence of minerals having lower temperature information. Spinel is present in Y-8002, but the grains are associated with troilite, not included in olivine, making the olivine-spinel thermometry (FABRIES, 1979; ROEDER et al., 1979) or speedometry (OZAWA, 1984) difficult. The exchange distribution coefficient of Mg and Fe between orthopyroxene and spinel (MUKHERJEE nad BULATOV, 1989) gives a temperature of 870°C, which is roughly consistent with the closure temperature of Mg-Fe interdiffusion in olivine and pyroxenes.

The lower temperature records are generally obtained from metallic phases, but the absence of taenite in Y-8002 makes it impossible to estimate lower temperature cooling rate of the meteorite. Taenite is not present in the thin section studied, but low Ni content in kamacite (5.5 wt% in the center of grain and 4.8% at the contact between schreibersite) and high Ni content in schreibersite (35% at the

contact with kamacite) predict that taenite should be present in the meteorite and that the zoning profile was achieved during cooling at low temperatures ($<400°C$) (GOLDSTEIN and DOAN, 1972; ROMIG and GOLDSTEIN, 1981).

4.   Mineralogical Comparison with Other "Unique" Meteorites

Y-8002 is related to Y-74063, Y-75274, ALH-77081, Y-791493, and probably Y-74357 and Y-78230 on the basis of mineralogy, particularly trace element contents in pyroxenes and composition of plagioclase and spinel. They may be further related to non-Antarctic "unique" meteorites, Lodran and Acapulco. In order to investigate genetic relationship and origin of unique meteorites, mineral compositions are compared. The average compositions are used for comparison, when available. About half of the compiled data are averages of more than 10 or 20 analyses, but half of the literature data are either representative or arbitrarily chosen ones. The compiled data are listed in Table 3. As far as Y-78230 is concerned, there are too little data up to now, so it will be eliminated in the following discussion although it may be related to the meteorites studied here.

MAC88177 is considered to be a new type of achondrite (PRINZ et al., 1991; TAKEDA et al., 1991; WARREN and KALLEMEYN, 1991), but its literature data on texture, mineralogy, and oxygen isotopic compositions are within the range of the "unique" meteorites studied here. The olivine composition of Fo_{87} is the same as most of the "unique" meteorites. Oxygen isotopic compositions ($\delta^{17}O = 0.47‰$, $\delta^{18}O = 3.25‰$, PRINZ et al., 1991) are close to Y-8002. It essentially lacks metal-troilite and plagioclase (although a small amount of melt is present), suggesting that it may be the residue of a small degree of partial melting of the common precursor material with the "unique" meteorites from which metal-troilite and melt were almost completely fractionated. Detailed mineralogical data will be required for comparison.

4.1.   Magnesium and iron in olivine, pyroxenes, and spinel

Figure 7 shows the X_{Mg} relationships among olivine, orthopyroxene, clino-pyroxene, and spinel of the "unique" meteorites. Equilibrium has been achieved as to Mg and Fe among olivine, orthopyroxene, clinopyroxene, and spinel except for those of Y-74357. Olivine in Y-74357 is much more magnesian than coexisting pyroxenes, which is because Y-74357 olivine was reduced at subsolidus temperatures resulting in an appearance similar to ureilites (YANAI et al., 1984; HIROI and TAKEDA, 1991). Although olivines in Lodran and Y-791493 are also reduced as shown by "reverse zoning" (BILD and WASSON, 1976; NAGAHARA and OZAWA, 1986), the effect of reduction is restricted to rims of grains. The effect of reduction at subsolidus temperature is, thus, not significant in almost all the meteorites studied. Since the Mg/Fe distribution between olivine and orthopyroxene has been known to be independent of temperature at high temperatures ($>1000°C$) (LARIMER, 1968), the different Mg/Fe ratios of the meteorites represent different bulk composition of silicate portions (Fig. 7a). Olivines and clinopyroxenes in those meteorites have a nearly constant Mg-Fe distribution coefficient, and they should have been in equilibrium down to about 700–800°C for most meteorites, again

Table 3. Mineralogical comparison among the "unique" meteorites and ordinary chondrites.

Sample	Y-8002	Y-8002	75274	75274	791493	791493	77081	74357	74063	78230	Acapulco	Lodran	H	L	LL
Data source	[1]	[2]	[1]	[2]	[2]	[3]	[4,5]	[2]	[5,6]	[7]	[8]	[9,10]	[11–15]	[11–15]	[11–14]
OI(Fo)	97	96	96	96	88	87	89	92	89	90	88	87	81	75	71
OI(MnO)	0.33		0.37			0.56	0.48	0.5	0.48		0.56	0.47	0.50	0.47	
Opx(X_{Mg})	97	96	96	96	87	88	90	86	87	90	86	86	83	79	74
Opx(MnO)	0.4	0.34	0.43	0.48	0.53	0.56	0.53	0.52	0.55		0.6	0.51	0.52	0.48	0.44
Opx(TiO_2)	0.22	0.19	0.25	0.27	0.08	0.2	0.18	0.14	0.18		018		0.19	0.2	0.21
Opx(Cr_2O_3)	0.19	0.46	0.38	0.47	0.35	0.35	0.24	0.43	0.21		0.2	0.39	0.09	0.11	0.14
Cpx(X_{Mg})	97	98	97	98	93	91	92	88	92		92	88	89	86	83
Cpx(MnO)	0.22	0.26	0.22	0.3	0.37	0.3	0.32	0.41	0.29		0.37	0.32	0.23	0.21	0.22
Cpx(TiO_2)	0.81	0.81	0.97	0.75	0.4	0.35	0.53	0.31	0.65		0.49		0.48	0.45	0.46
Cpx(Cr_2O_3)	0.38	0.77	0.84	0.74	1.37	1	1.41	1.25	1.14	1.35	1.08	1.55	0.59	0.61	0.6
Pl(An)	31	28		28	18	19	16	15	15	14	15		12	10	11
Sp(X_{Mg})	0.73					0.26	0.49		0.33		0.37	0.34	0.13	0.1	0.08
Sp(Cr/(Al+Cr))	0.90					0.87	0.87		0.86		0.87	0.87	0.87	0.88	0.87
Sp(MnO)	1.72					1.36	1.25		1.23		1.34	1.1	0.94	0.74	0.65
Sp(TiO_2)	0.49					0.66	1.2		1.22		1.36	1.16	2.33	2.81	3.23
Sp(Cr_2O_3)	67					62	60		60		62	60	57	56	54

Oxides (wt%), $X_{Mg}=Mg/(Mg+Fe)$

[1]: this study, [2]: YANAI et al. (1984), [3]: NAGAHARA and OZAWA (1986), [4]: SCULTZ et al. (1982), [5]: KIMURA (unpubl. data), [6] YANAI and KOJIMA (1991), [7] HIROI and TAKEDA (1991), [8]: PALME et al. (1981), [9]: PRINZ et al. (1978), [10]: BILD and WASSON (1976), [11] VAN SCHMUS (1969), [12] BUNCH et al. (1967), [13] BUNCH and OLSEN (1973), [14] RUBIN (1990), [15] NOGUCHI (unpubl. data)
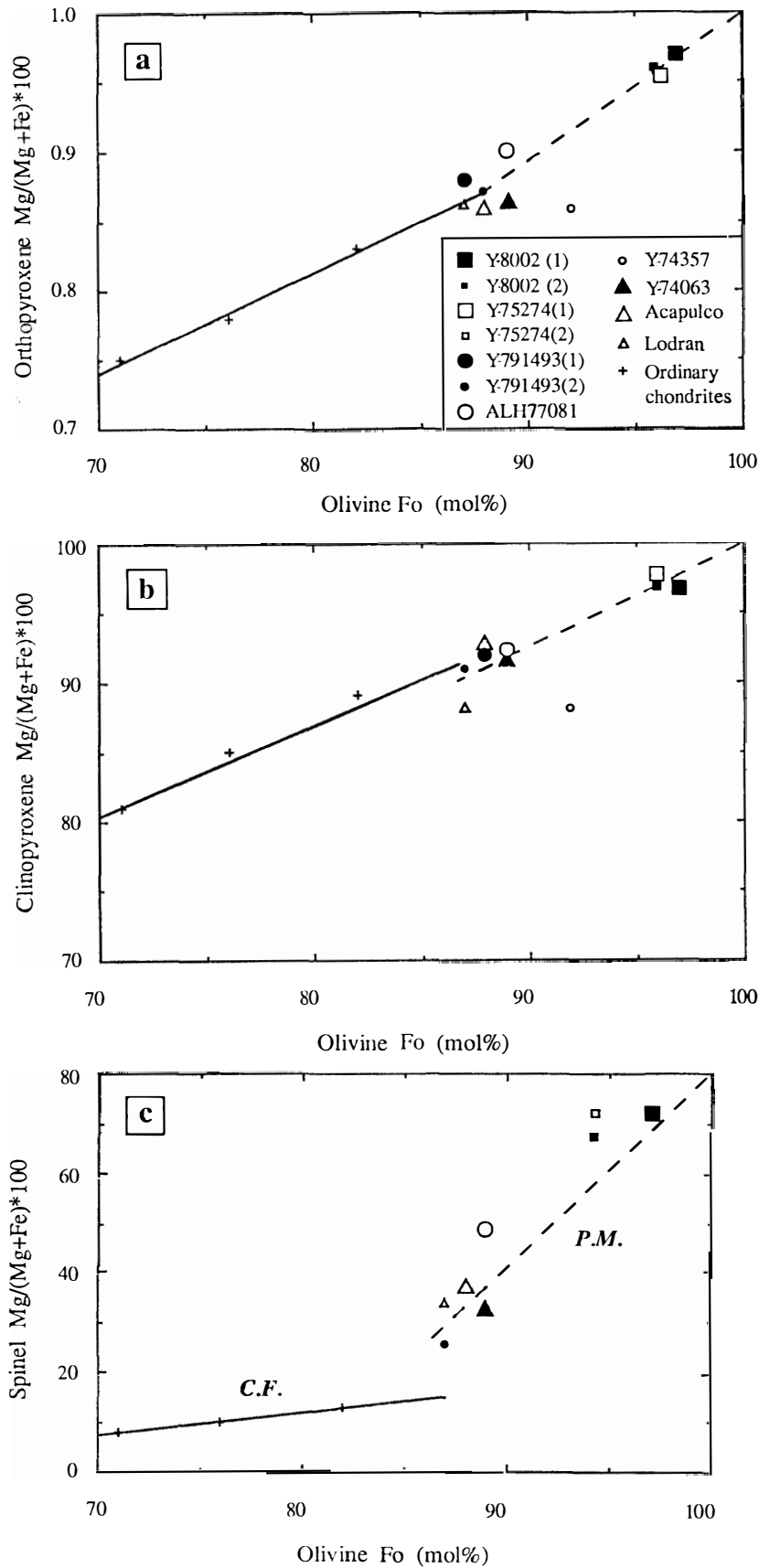
Fig. 7. X_{Mg} relationships between Fo content of olivine and (a) orthopyroxene, (b) clinopyroxene, and (c) spinel. Composition of minerals are averages of more than 10 or 20 analyses or representative value listed in literatures. Composition and data source are listed in Table 3. Y-8002 (1) and Y-75274 (1) are from present study, Y-8002 (2), Y-75274 (2), and Y-791493 (1) are from YANAI et al. (1984), and Y-791493 (2) is from NAGAHARA and OZAWA (1976).

except for Y-74357 (Fig. 7b). Olivine and spinel are plotted on a similar distribution line although the data of ALH-77081 show a little scatter (Fig. 7c), and give similar "equilibration" temperatures (800–900°C). As will be discussed later, highly magnesian nature of Y-8002 and Y-75274 is due to reduction during melting at high temperatures.

The relationships shown above indicate that the meteorites are highly equilibrated as to Mg and Fe, that the equilibration temperature is almost the same although the Mg/Fe ratio varies widely, and that they suffered a similar cooling history at subsolidus temperatures. The Mg/Fe ratio of ferromagnesian minerals were controlled by the bulk composition (Mg/Fe ratio) of the rock.

4.2. Manganese, chromium and titanium in olivine, pyroxenes, and spinel

The relationships between Fo content of olivine and MnO contents of olivine, pyroxenes, and spinel are shown in Fig. 8. The MnO contents of olivine, orthopyroxene, and clinopyroxene increase with decreasing Fo content, which is general in most geochemical processes although it is not the case for ureilites where the effect of reduction at subsolidus temperatures is large (GOODRICH and BERKLEY, 1985; GOODRICH et al., 1987). The MnO content of spinel shows the opposite relation in that the MnO content increases with increasing Fo content. This is due to high MnO content of Y-8002 spinel, but it is doubtful whether the value is representative of Y-8002 spinels or not. If we take the MnO content of the large spinel grain (1.2 wt%) instead of the average value, which is shown by a diamond in Fig. 8d, the MnO content of spinel in the "unique" meteorites makes an inverse relationship with the Fo number, being consistent with the relationships of olivine and pyroxenes. Further analysis for big spinels coexisting with silicates will be required. In spite of the uncertainty of internal relationship, the MnO contents of spinels in the "unique" meteorites are a little higher than those of ordinary chondrites.

The TiO_2 contents of orthopyroxene are weakly related to the Fo contents (Fig. 9a) and those of clinopyroxenes are positively related to the Fo contents (Fig. 9b). On the contrary, those of spinels are negatively correlated (Fig. 9c). The TiO_2 contents of orthopyroxene are nearly the same level as those in the ordinary chondrites, those in clinopyroxene are much higher, and those of spinels are significantly lower.

The Cr_2O_3 contents of orthopyroxene are scattered in the diagram (Fig. 10a), and it is difficult to see any internal relationship. The Cr_2O_3 content of clinopyroxene is negatively related to the Fo content (Fig. 10b), suggesting the same trend for orthopyroxene. If it is the case, the Cr_2O_3 data for Y-75274 and Y-8002 except for the present study (large solid square) are too high. The Cr_2O_3 contents of spinel are inversely related to the Fo content (Fig. 10c), which are significantly higher than those in ordinary chondrites.

4.3. Plagioclase

The An content of plagioclase is positively related to the Fo content of olivine (Fig. 11a), and is higher than that of ordinary chondrites. The An contents of
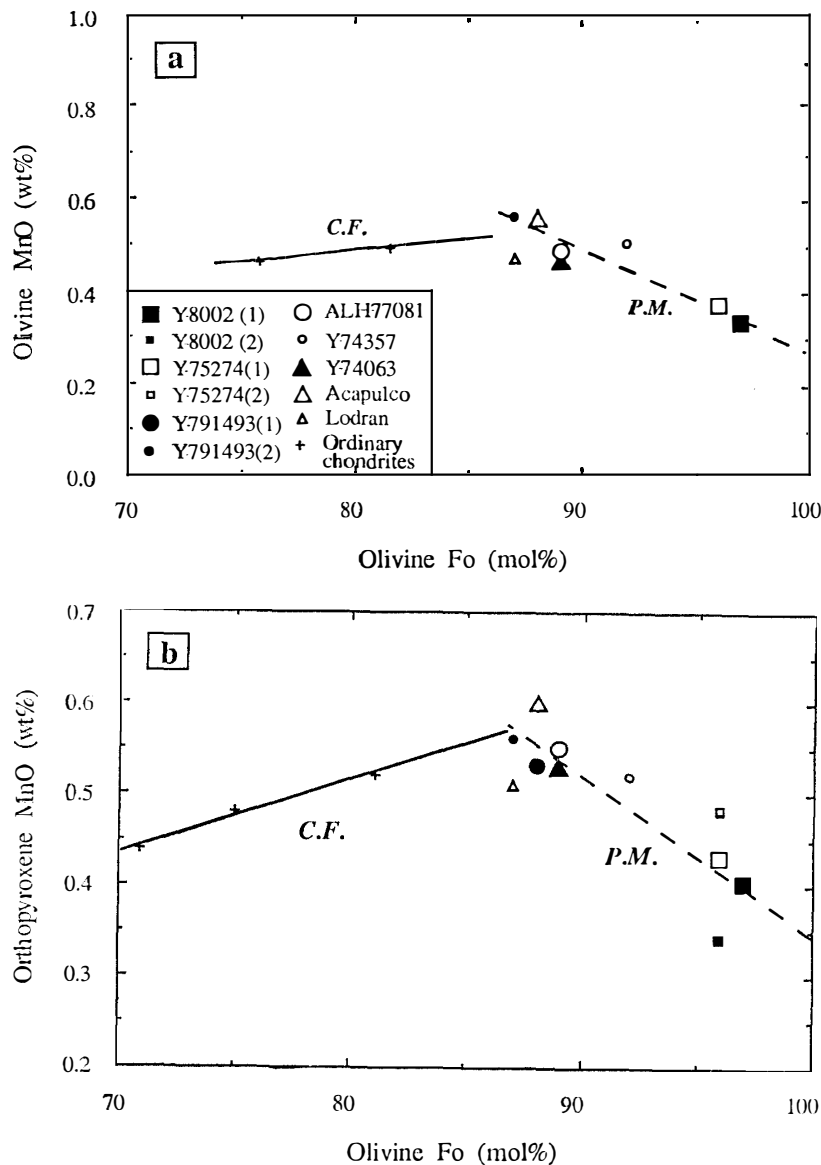
Fig. 8. The relationships between Fo contents of olivine and MnO contents of (a) olivine, (b)
orthopyroxene, (c) clinopyroxene, and (d) spinel in the "unique" meterorites. P.M. shows the
partial melting trend of the "unique" meteorites and C.F. shows the cosmochemical fractionation
trend responsible for the origin of chemical group of ordinary chondrites. The MnO contents of
olivine, orthopyroxene, and clinopyroxene decrease with increasing Fo contents of olivine. See
discussion in text for spinel. Primary compositions of clinopyroxene and spinel in the precursor are
richer in MnO than those in ordinary chondrites. P.M. trends for spinel based on the "average"
composition of Y-8002 (black solid) and that based on composition of the spinel #1 (diamond) are
shown (Fig. 8d)

plagioclase, as well as X_{Mg} of ferromagnesian minerals, are classified into two
groups: Y-8002 and Y-75274 have calcic plagioclase, and those in other meteorites
have less calcic plagioclase. The An content of plagioclase is positively related to
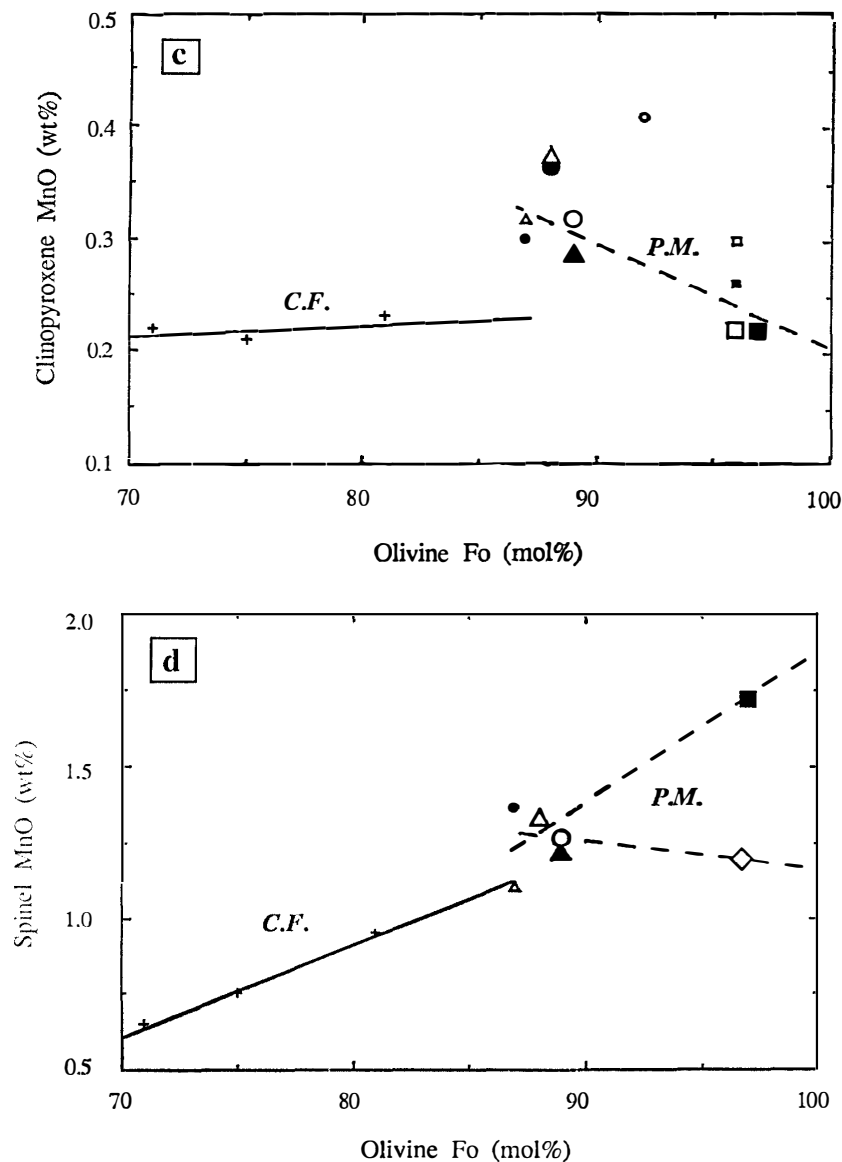the X_{Mg} of pyroxenes and spinels and the Cr_2O_3 content of spinel (Fig. 11b).

Fig. 8. (Continued).

4.4. Oxygen isotopic compositions

The oxygen isotopic compositions of the "unique" meteorites have been reported (CLAYTON et al., 1984; MAYEDA et al., 1987, MAYEDA and CLAYTON, 1989), and are summarized in Fig. 12a. They are plotted in a wide area on the three isotope plot below the terrestrial fractionation line. There is no clear relationship between mineralogy and oxygen isotopic composition, but Y-8002 and Y-75274, which were formed through high degree of partial melting as will be discussed later, and Y-74357, which was reduced at subsolidus temperatures, in the lower portion of the three isotope plot than other meteorites.

The problem is similar to the oxygen isotopic compositions of ureilites (CLAYTON and MAYEDA, 1988). Oxygen isotopic compositions of ureilites are plotted around the "Allende mixing line" with slope of 1 with significant scatter. CLAYTON and
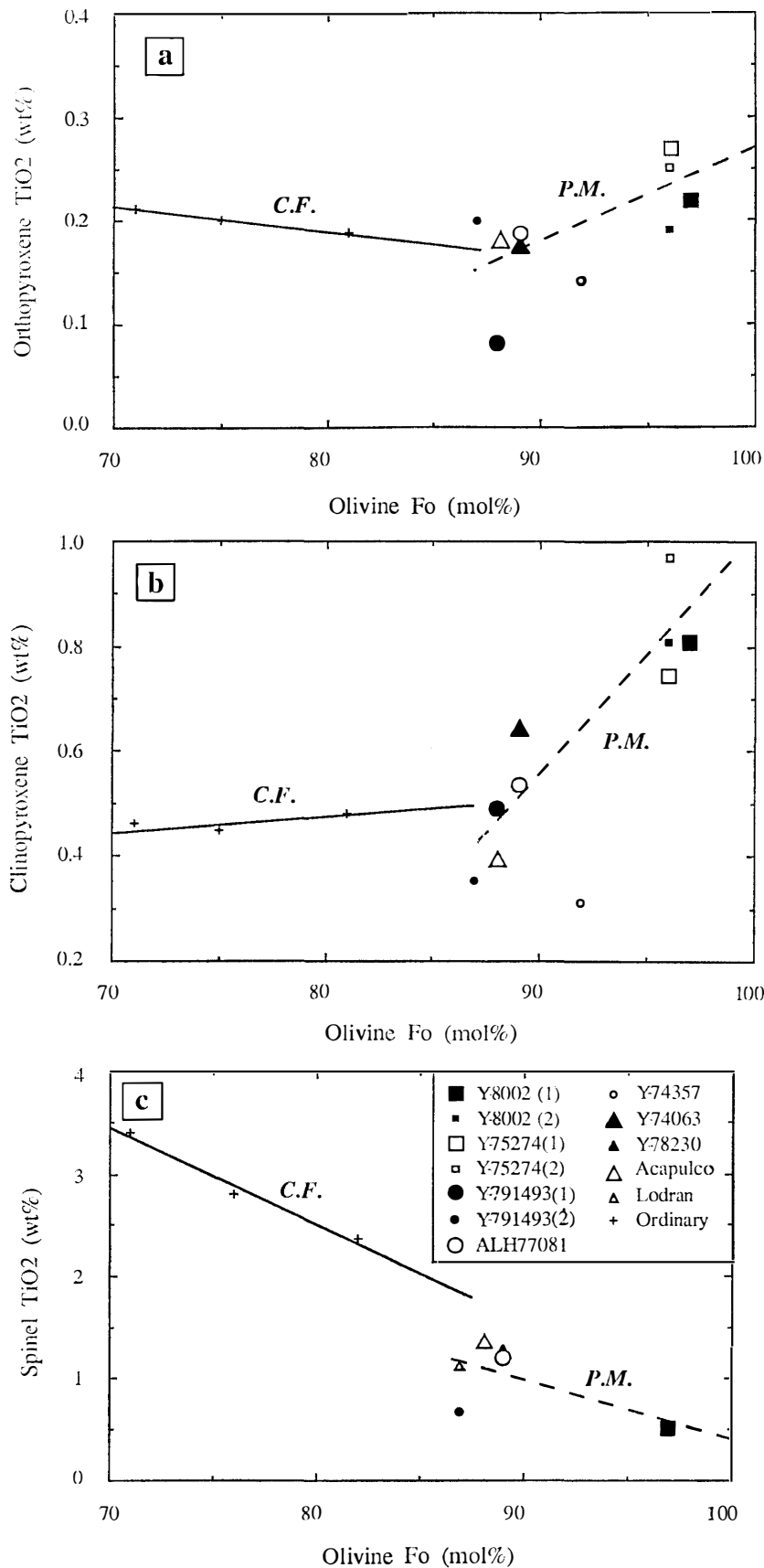
Fig. 9. The relationships between Fo contents of olivine and TiO₂ contents of (a) orthopyroxene, (b) clinopyroxene, and (c) spinel in the "unique" meteorites. Diversity of TiO₂ contents of orthopyro-xene and an apparent increase of TiO₂ of clinopyroxene with Fo are not well understood. Primary compositions before partial melting are nearly the same as those estimated from cosmo-chemical fractionation.
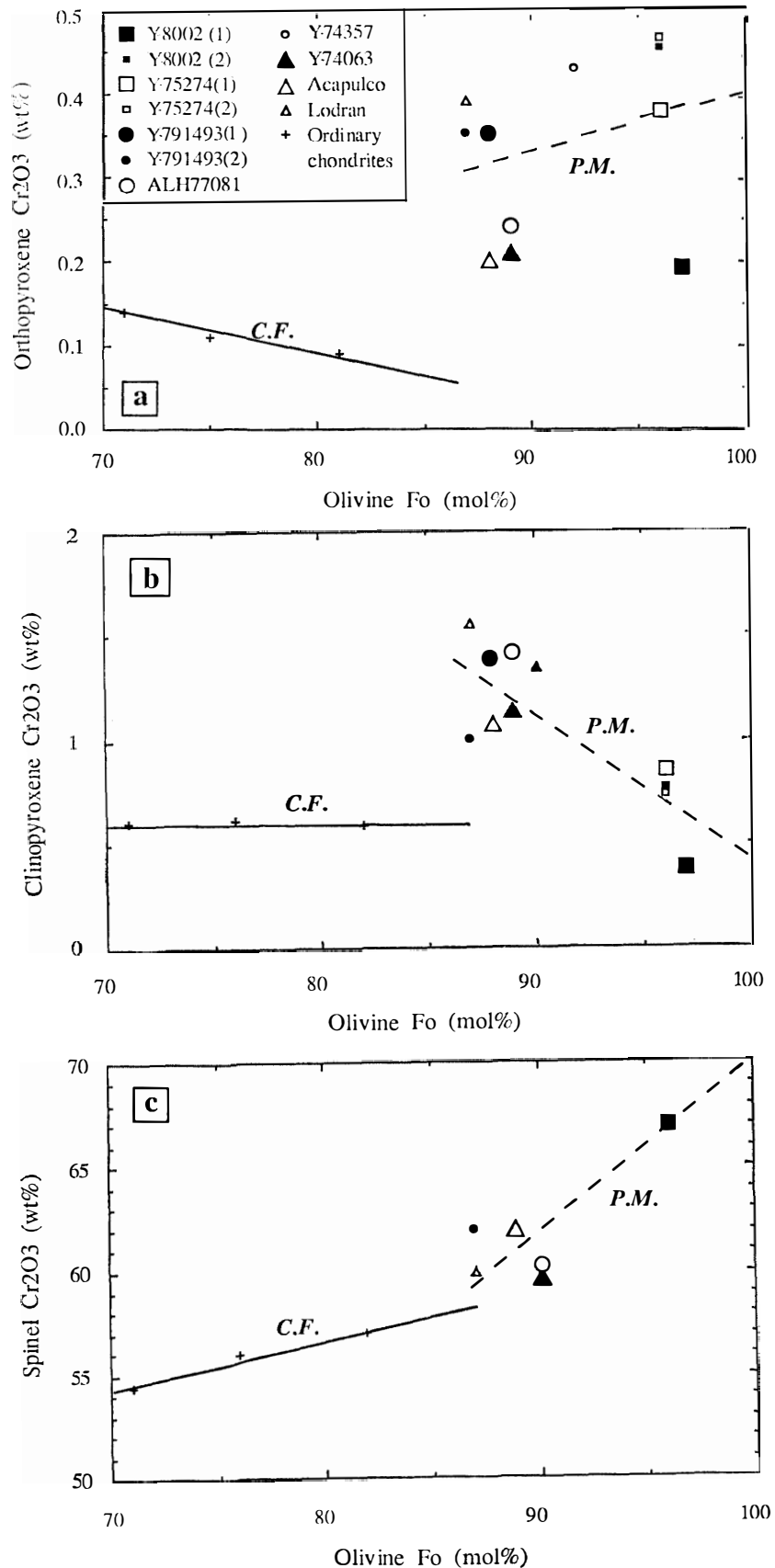
Fig. 10. The relationships between Fo contents of olivine and Cr2O3 contents of (a) orthopyroxene, (b) clinopyroxene, and (c) spinel in the "unique" meteorites. Cr2O3 in orthopyroxene does not show a clear trend, that in clinopyroxene decreases, and that in spinel increases with increasing Fo content. Decrease in Cr2O3 in clinopyroxene with increasing Fo content may be due to crystallization of highly chromian spinel.
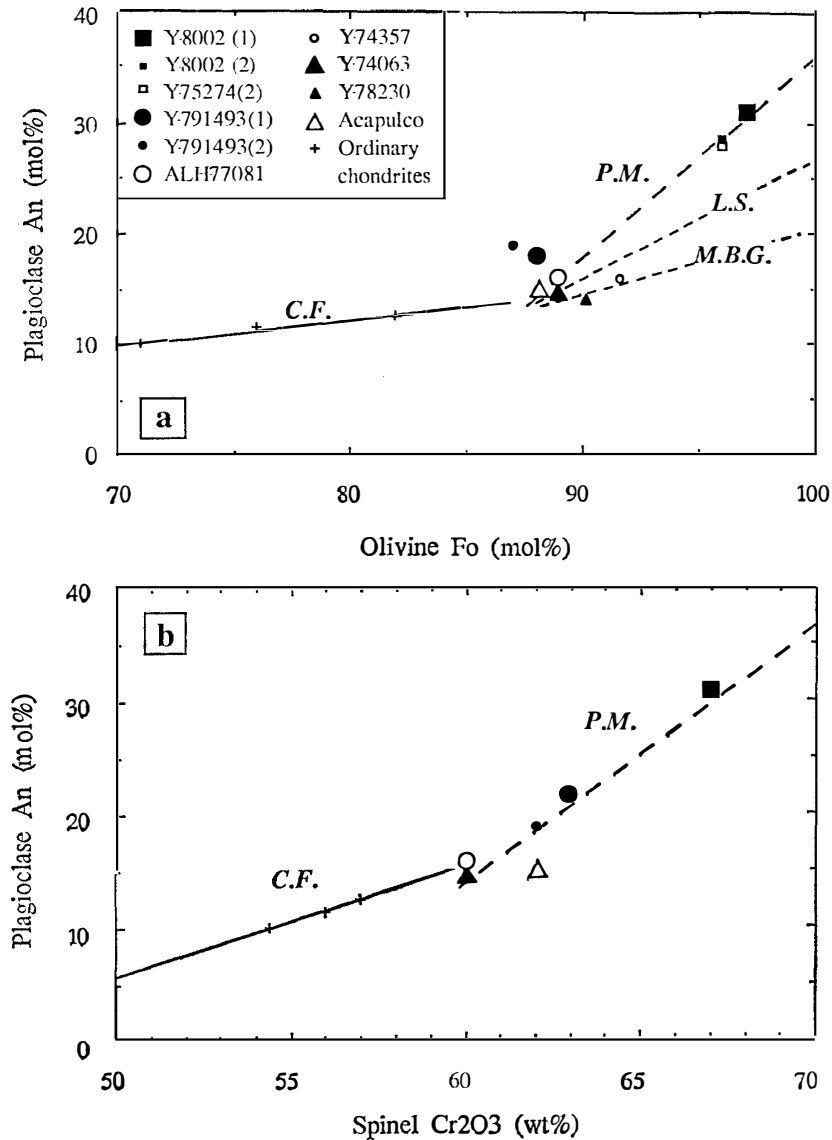
Fig. 11. The relationships between the An contents of plagioclase and (a) Fo contents of olivine and (b) Cr_2O_3 contents of spinel in the "unique" meteorites. L.S. and M.B.G. in (a) show the trends parallel to the fractional crystallization trends of gabbros in the Layered Series (L.S.) and Marginal Border Group (M.B.G.) of Skaergaard Intrusion. Absolute compositions of L.S. and M.B.G. are different from the trends shown in the figure, but the slope are the same. They are more gentle slope than the P.M. trend by the "unique" meteorites. Average composition is shown for M.B.G. Data from WAGER and BROWN (1968) and HOOVER (1989).

MAYEDA (1988) concluded that the diversity of oxygen isotopic compositions of ureilites originated in nebular processes, which was not homogenized through igneous processes in the parent body. GOODRICH *et al.* (1991), however, on the basis of Sm-Nd and Rb-Sr dating, proposed a mixing of the ureilite parent body with a
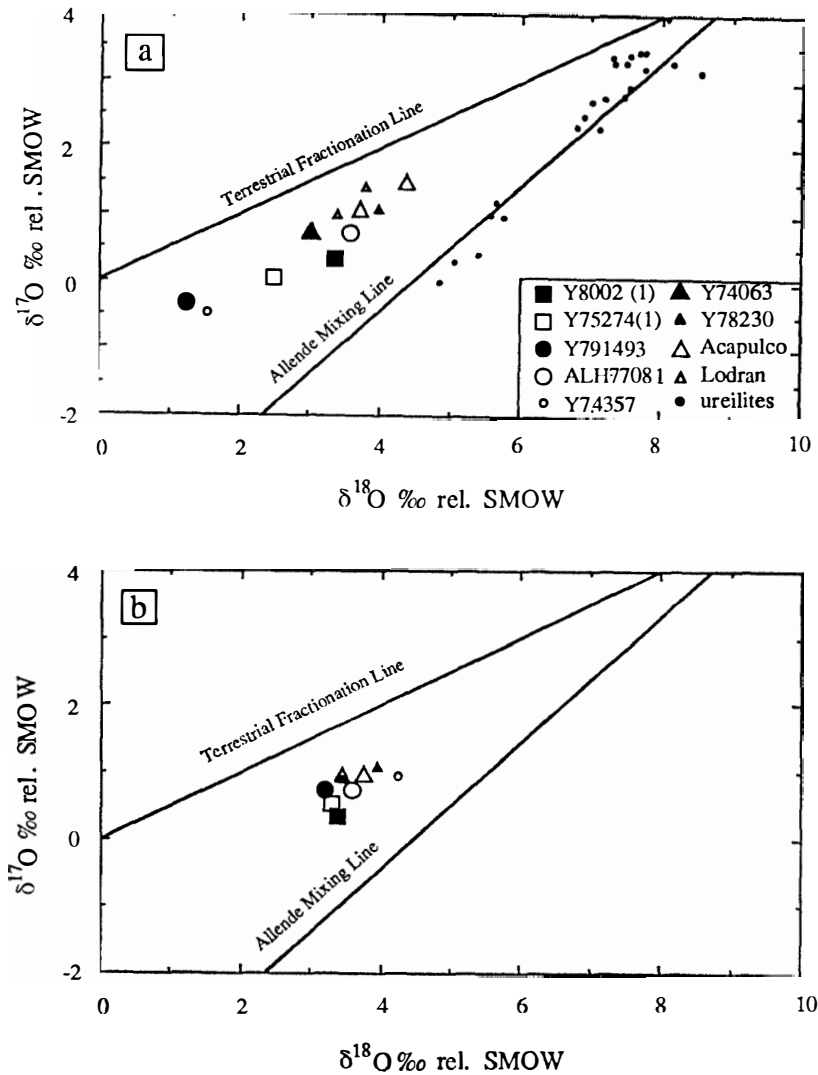
Fig. 12. (a) Oxygen isotopic compositions of the Antarctic and non-Antarctic "unique" meteorites and ureilites. The "unique" meteorites are plotted along a line with a slope of 1/2 with large scattering, and reduced meteorites, Y-8002, Y-75274, and Y-74357, are plotted below the line. Data from CLAYTON and MAYEDA (1978, 1988), CLAYTON et al. (1984), and MAYEDA and CLAYTON (1980, 1989) and MAYEDA et al. (1987).

(b) Oxygen isotopic compositions of the calculated values for Y-74063, Y-74357, Y-75274, and Y-791493 (see text for calculation), data for bulk Lodran and Acapulco along with original data for Y-78230, Y-8002 and ALH-77081. They plot in a narrow area, suggesting derivation from a common precursor.

material like carbonaceous chondrites matrix after igneous differentiation.

The reason for the great diversity of oxygen isotopic compositions of the "unique" meteorites may be partial oxidation of metallic iron in Antarctica, which moves the isotopic compositions toward the lower left of the T. F. line. We can estimate the primary oxygen isotopic compositions of Y-74063, Y-74357, Y-75274, and Y-791493

assuming that all oxygen in Fe_2O_3 obtained by the wet chemical analysis was derived from Antarctic ice through weathring processes. The published and unpublished Fe_2O_3 contents in those meteorites are 0.91, 7.55, 2.8, and 6.0 wt%, respectively (HARAMURA et al., 1983; NAGAHARA and OZAWA, 1986; YANAI and KOJIMA, 1991; HARAMURA unpub. data). From these values, we get 3.45‰ of $\delta^{18}O$ (3.01 before recalculation) and 0.98‰ of $\delta^{17}O$ (0.76) for Y-74063, 4.26 (1.52) and 0.94 (−0.43) for Y-74357, 3.3 (2.4) and 0.53 (0.08) for Y-75274, and 3.2 (1.19) and 0.73 (−0.28) for Y-791493, respectively. The calculated oxygen isotopic compositions are shown in Fig. 12b with those of bulk compositions for Lodran and Acapulco (MAYEDA et al., 1987; CLAYTON, unpublished data) instead of mineral data. They plot in a narrow area; thus the meteorites could be derived from a common parent body.

The calculated compositions except for those of Y-74357 and Y-8002 are along a line with slope of 1. The diversity is 0.8‰ for $\delta^{18}O$, 0.54‰ for $\delta^{17}O$, and −0.93 to −1.19 for $\delta^{17}O$, which are no larger than twice of those for mineral separates from Lodran and Acapulco (0.4 to 0.7‰ for $\delta^{18}O$, 0.4‰ for $\delta^{17}O$, and −0.8 to −0.94 for $\delta^{17}O$) (CLAYTON and MAYEDA, 1978; MAYEDA and CLAYTON, 1980). Since those meteorites are coarse-grained, small amount of samples provided for oxygen isotopic analysis can result in apparent large diversities, which is well shown in the data for mineral separates from Lodran and Acapulco. The relatively large diversity of Y-74357 may be due to overestimation of effect of weathering and that of Y-8002 to data of which effect of weathring has not been evaluated. Therefore, the "unique" meteorites including Lodran and Acapulco have similar oxygen isotopic compositions, which is a strong evidence for their derivation from a common precursor.

5. Genetic Relationships among the "Unique" Meteorites

Major and minor elemental relationships among olivine, pyroxenes, spinel, and plagioclase in the "unique" meteorites can be explained by 1) fractional crystallization in a common parental body, or 2) partial melting in a common parental body. NAGAHARA and OZAWA (1986) discussed that Y-791493, ALH-77081, Lodran, and Acapulco were formed through different degrees of partial melting (difference in temperature) from a common chondritic precursor material in a small body where metal-troilite separation from silicates hardly took place but extraction of partial melt easily occurred. The present study supports their conclusion; the difference in mineralogy of the "unique" meteorites can be well understood by the difference in degree of partial melting. A model of multiple parent bodies may be possible, but the similar oxygen isotopic compositions of those meteories (Fig. 12b) prefer a single parent body. Accordingly, there is no necessity to consider a multiple parent body model in this paper.

Experimental studies have shown that X_{Mg} of olivine, orthopyroxene, and Ca-pyroxene becomes higher with higher degree of partial melting although the distribution coefficient is nearly constant over a wide compositional, temperature, and pressure range (e.g., DUKE, 1976; LONGHI et al., 1978; JAQUES and GREEN, 1980;

TAKAHASHI and KUSHIRO, 1983). In this scheme, Y-8002 and Y-75274 are thought to have been formed through the highest degree of melting and the others have suffered lower degree of melting. There should be a relationship between relative abundances of silicates and metallic phases and degree of partial melting (that is, X_{Mg} of mafic minerals or An content of plagioclase), if the meteorites were formed through different degrees of partial melting. However, as mentioned before, modal compositions or bulk compositions of those meteorites are not representative because of very small sample sizes in spite of fairly coarse grain sizes of the constituent minerals. The MnO distribution between olivine and melt or between pyroxenes and melt have been well studied for various systems and at various temperatures and pressures. With increasing temperature, the MnO contents of olivine and clino-pyroxene decrease (e.g., WATSON, 1977; IRVING, 1978; LINDSTROM and WEILL, 1978; JAQUES and GREEN, 1980; RAY et al., 1983; DUNN, 1987). The relationship between MnO contents of olivine, orthopyroxene, and clinopyroxene and Fo contents in the "unique" meteorites support the idea that Y-8002 and Y-75274 were formed through higher degree of partial melting and that the others were formed through lower degree of partial melting. As stated before, the MnO content of spinels appears to be inconsistent with the partial melting model, but it may be due to unrepresentative data of Y-8002 spinel.

The TiO_2 contents in orthopyroxenes and clinopyroxenes in the "unique" meteorites are positively related to the Fo content of olivines (Fig. 9a, b), on the other hand that of spinel is inversely related (Fig. 9c). Experimental works have shown that distribution of Ti in Ca-pyroxene becomes smaller with increasing tem-perature (LINDSTROM and WEILL, 1978; RAY et al., 1983) or constant (DUKE, 1976; DUNN, 1987), which contradicts the partial melting model for the "unique" meteorites. The reason for this discrepancy is not known, but may be due to crystallization under highly reducing conditions where all or part of Ti was reduced to T^{3+} and behaved as a compatible element.

Experimental results by JAQUES and GREEN (1980) showed that the $Cr/(Cr+Al)$ ratio of residual spinel increases and that the Cr_2O_3 content of orthopyroxene increases with increasing degree of partial melting. Aluminum is highly compatible with melt, is lost from the residue, and the residual minerals become enriched in Cr. The relationship between Fo content and Cr_2O_3 contents of spinel in the "unique" meteorites is consistent with the partial melting model, though that between Fo and Cr_2O_3 in clinopyroxene is not. This inconsistency may be due to crystallization of highly chromian spinel prior to clinopyroxene, which reduced the concentration of Cr in silicate melt.

Plagioclase compositions of the "unique" meteorites can be explained also by the partial melting model. It is evident from texture that plagioclase was an inter-stitial phase, that is, it was liquid during melting. Melt was almost completely lost from Lodran. Extraction or movement of partial melt is easy in a short time laboratory experiments (e.g., TAKAHASHI, 1983), and is also easy under the presence of temperature gradient in the system (WALKER et al., 1988). Very small amount of melt was lost from others. On the contrary, it is difficult to retain partial melt at high degree of melting which is estimated to be more than 10% from modal

composition of plagioclase. Existence of interstitial melt implies that the meteorite melted in a place where gravitational separation of liquid was not effective. It does not favor a large magma chamber, and rather prefers a small melt stream, which will be discussed later.

When a chondritic material melts in a closed system, the melt composition should become more calcic with increasing degree of melting because Na is much more compatible with melt than Ca. The Fo content of olivine and the Cr_2O_3 content of spinel are positively related to the An content of plagioclase in the "unique" meteorites (Fig. 11), showing that Y-75274 and Y-8002 were formed through the highest degree of melting, followed by Acapulco, Y-791493, ALH-7781, Y-74063 and Lodran, of which relative order is not resolvable. This order appears to be partly inconsistent with textural observation; Y-74063 and ALH-77081 have relict chondrules; on the contrary, Y-791493 and Lodran are coarse-grained magmatic rocks like terrestrial peridotites. Furthermore, Lodran does not have plagioclase. This discrepancy comes mainly from heterogeneity of plagioclase (up to 5 mol% An content in Y-8002) although the average values are shown in Fig. 11. The literature data for degree of heterogeneity of plagioclases of other samples are 1–2 mol%, but further detailed study will be required. The discrepancy comes also from uncertainty of the compiled mineral data where relative error among laboratories may be 1 mol% for Fo and X_{Mg} of other ferromagnesian minerals, from small sample sizes which makes analytical uncertainty larger, and from insensitivity of mineral compositions against partial melting with low degree of partial melting. Thus, the unclear relationship among Fo content of olivine, An content of plagioclase, and texture of Y-74063, Y-78230, Y-791493, ALH-77081, Acapulco, and Lodran is due to quality of the compiled data.

The REE patterns of the "unique" meteorites (Fig. 13) support the idea that the "unique" meteorites were produced through different degrees of melting from a common precursor which were estimated from mineralogical data. We should be careful in interpreting the REE patterns of those meteorites. At first, although the meteorites are residue of partial melting, partial melts were not completely lost, but yet melt (plagioclase) remained in the interstices in most meteorites. This is probably because melt extraction was not effective in a small body due to small gravity and melting in a small melt stream. It means that the stage of crystallization of cumulus phases and interstitial phases were not the same; it shows a possibility that melt network transported interstitial melt after solidification of cumulus phases, which is general for cumulate rocks. Furthermore, the samples provided for REE analyses are not necessarily representative of bulk composition of rocks because of small size, as well as modal compositions and major elemental compositions.

Y-74063 and ALH-77081 are primitive as shown by flat chondritic patterns (Schultz et al., 1982; Yamamoto et al., 1991); Y-8002 is residue of partial melting where small amount of melt rich in plagioclase component was added after extraction of partial melt, resulting in enrichment of heavy REE with small positive Eu anomaly (Yamamoto et al., 1991). Acapulco is rich in a melt component with significant light REE enrichment and flat heavy REE (Palme et al., 1981). Lodran is primitive having weak heavy REE enrichment and minor depletion of Eu
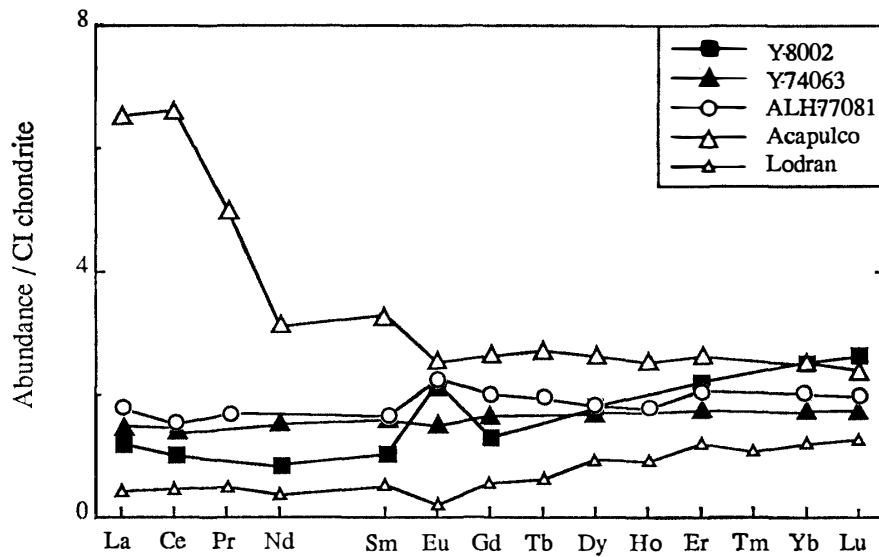
Fig. 13. REE patterns of the Antarctic and non-Antarctic "unique" meteorites. Y-74063 is flat, suggesting that it is primitive, ALH77081 is similar to Y-74063 with slight addition of plagioclase, Y-8002 and Lodran are residue with slight enrichment or loss of plagioclase (melt), and Acapulco is enriched in melt fraction. Data from FUKUOKA et al. (1978), PALME et al. (1981), SCHLUTZ et al. (1982), and YAMAMOTO et al. (1991).

(FUKUOKA et al., 1978). The REE concentration of Y-8002 is higher than Lodran. Although this is consistent with the fact that Lodran does not have interstitial plagioclase and that Y-8002 has abundant plagioclase, it is inconsistent with the model of higher degree of partial melting for Y-8002 than Lodran. The reason for this discrepancy is known.

FUKUOKA et al. (1978) proposed disequilibrium melting for the origin of REE pattern of Lodran. PALME et al. (1981) also suggested "some disequilibrium melting" for the origin of Acapulco REE pattern. However, disequilibrium melting would not be the only one possibility for the Lodran and Acapulco REE patterns. They assumed melting in closed systems; however, the systems should have been open as mentioned before. The chemical compositions of plagioclase and REE patterns suggest that the abundance of interstitial melt changed during melting. The modal composition of these meteorites does not show the degree of melting because both melt extraction and addition took place.

Mineralogical and textural characteristics of the "unique" meteorites can also be achieved through a fractional crystallization process. Generally, it is not easy to distinguish the effect of fractional crystallization and partial melting, as has been debated for the origin of the HED parent body. Bulk compositions of eucrites plot near the peritectic point of the system olivine-pyroxene-plagioclase, which is suggestive of formation of eucrites as a partial melt (STOLPER, 1977; STOLPER et al., 1979; SACK et al., 1991). Partial melting model, however, did not well explain the compositional continuum from diogenites to eucrites, which is rather favorable to the fractional crystallization model (IKEDA and TAKEDA, 1985; HEWINS, 1986;

WARREN, 1985). In the case of the "unique" meteorites, we have residues but do not have meteorites corresponding to a melt fraction, which makes the understanding of the origin ambiguous.

The compositional relationship between olivine and plagioclase may distinguish the process of partial melting and fractional crystallization. In the process of fractional crystallization, change of Fo content is larger than that of An content at the early stage of fractionation when olivine and plagioclase crystallize simultaneously; with progress of fractional crystallization, the An mol% decreases more slowly than Fo mol% although the absolute Fo and An content are dependent on magma compositions. The lines labelled LS or MBS in Fig. 11a are the trends of the Layered Series (LS) and Marginal Border Series (MBS) of the Skaergaard gabbroic Intrusion (WAGER and BROWN, 1968; HOOVER, 1989). These lines have more gentle slope than the line defined by the "unique" meteorites, which suggests that the "unique" meteorites were not formed through fractional crystallization, although plagioclase of the Skaergaard Intrusion is cumulus and those in the "unique" meteorites are interstitial.

The largest problem for the fractional crystallization model is that Y-74063 and ALH-77081, which have the lowest X_{Mg} of ferromagnesian minerals, have chondritic texture. If they were formed through fractional crystallization from more magnesian precursor, they should have igneous texture. This excludes the possibility of fractional crystallization for the origin of "unique" meteorites.

Formation of the "unique" meteorites through partial melting does not necessarily mean crystallization in a large magma chamber. Abundant concentration of rare gases in those meteorites (TAKAOKA and YOSHIDA, 1991) is inconsistent with melting and crystallization in a big magma chamber which should have caused loss of rare gases. They should have formed in a small melt pocket or melt stream in a fairly short time. NAGAHARA and OZAWA (1986) discussed on the basis of olivine petrofabric analysis that Y-791493 melted in a small melt stream. Olivine in Y-791493 has a fabric of Y(c) concentration, which suggests formation in a magmatic laminar flow rather than by crystal accumulation in a stagnant magma chamber. Petrofabric analysis for Y-8002 has not yet done, but a preliminary result shows that Y-74063 olivine also shows the Y(c) concentration (TSUCHIYAMA, private communication), supporting the conclusion that the "unique" meteorites melted in a small magma stream. Heterogeneous melting in a parental body may be important for the origin of the "unique" meteorites.

6.   Precursor Material of the "Unique" Meteorites

As discussed in the previous chapter, the "unique" meteorites were formed through partial melting from a common precursor. Compositional relationships among minerals in the "unique" meteorites and those among ordinary chondrites make it possible to estimate the primary composition of the "unique" meteorites before melting. Since mineral compositions from meteorites with different degrees of partial melting are linearly related to each other (Figs. 7–11), the precursor composition before melting should be on the extension of the lines. The line

defined by minerals in the "unique" meteorites were formed during partial melting, and it will be called "partial melting trend (P.M.)", hereafter. Data of Y-74357 were not used, because of reduction of olivine.

Although the "unique" meteorites shows various textures, the chondritic texture with relict of chondrules in ALH-77081 (SCHULTZ et al., 1982) and in Y-74063 (YANAI and KOJIMA, 1991) indicates that the precursor was a kind of chondrite and that the equilibrium Fo content should be 87. The precursor chondrite is tentatively called as an "U(nique)" chondrite. The relationships among X_{Mg} and the Mn, Ti and Cr contents of ordinary chondrite minerals are investigated (Fig. 7–11). The trends are linear, which should have been established through cosmochemical fractionation by which chemical group of ordinary chondrites were achieved. Tine will be called "cosmochemical fractionation line (C. F.)", hereafter. If the composition of minerals in "U" chondrite are on the extension of the C. F. lines, it means that the "U" chondrites were formed in a similar region in the solar nebula as ordinary chondrites where a common cosmochemical fractionation process played a role on the bulk composition of the precursor materials.

X_{Mg} of ortho- and clinopyroxene at Fo_{87} on the P. M. are 0.87 and 0.92, respectively, which are on the extension of C. F. X_{Mg} of spinel (0.3) is higher than that at Fo_{87} on C. F. (0.15). Other compositions thus obtained are: An of plagioclase 15, MnO in olivine 0.5, MnO in orthopyroxene 0.58, MnO in clinopyroxene 0.34, MnO in spinel 1.18, TiO_2 in orthopyroxene 0.15, TiO_2 in clinopyroxene 0.42, TiO_2 in spinel 1.3, Cr_2O_3 in orthopyroxene 0.3, Cr_2O_3 in clinopyroxene 1.3, and Cr_2O_3 in spinel 60. The An contents of plagioclase (Fig. 11a), and the MnO contents of orthopyroxene (Fig. 8) on P. M. and C. F. cross at Fo_{87}, but those for X_{Mg} of spinel (Fig. 7) and MnO of clinopyroxene and spinel (Fig. 8), TiO_2 and Cr_2O_3 in pyroxenes and spinel (Figs. 9 and 10) do not. The estimated values are summarized in Table 5. Numbers in parenthesis are the values estimated from the cosmochemical fractionation which formed the chemical group of ordinary chondrites.

The higher P. M. values of MnO and particularly Cr_2O_3 contents at Fo_{87} than C. F. values at Fo_{87} should reflect higher contents of those elements in the "U" chondrite. High Cr content in Acapulco has been previously pointed out by PALME et al. (1981) and that of Y-791493 by NAGAHARA and OZAWA (1986). SCHULTZ et al. (1982) thought that the high Cr content of ALH-77081 is due to insufficient sample allocated to the bulk chemical analysis in which chromite is heterogeneously

Table 4. Estimated mineral compositions of the "U" chondrite.

	X_{Mg} or An	MnO	TiO_2	Cr_2O_3
Olivine	87	0.5		
Orthopyroxene	86	0.58	0.15(0.18)	0.3(0.1)
Clinopyroxene	92	0.34(0.23)	0.42(0.48)	1.3(0.6)
Spinel	0.3(0.16)	1.18(1.08)	1.3(1.8)	60(58)
Plagioclase	15			

Numbers in parenthesis are values estimated from the cosmochemical fractionation trend responsible for chemical group of ordinary chondrites

distributed. They should have agreed to the high concentration of Cr if they investigated the high abundance of Cr in pyroxenes and spinel. High concentration of moderately volatile siderophile elements, P, Mn, Zn, and Ga as well as Cr, have been pointed out for Y-791493 (NAGAHARA and OZAWA, 1986), Acapulco (PALME et al., 1981) and for ALH-77081 (SCHULTZ et al., 1982). Although there is not sufficient analytical data for the "unique" meteorites, the data for Y-74063 (FUKUOKA and KIMURA, 1990) show also high concentration of Mn. The Cr concentration is, however, unexpectedly low, which is even lower than that of ordinary chondrites. This may be actually because of insufficient sample allocated to bulk analysis. The MnO data for Y-74063 by YANAI and KOJIMA (1991) seems to be too low; the MnO contents in olivine, pyroxenes, and spinel shown in their paper are higher than the bulk data. The data are also inconsistent with the analytical result by FUKUOKA and KIMURA (1990). Thus the MnO content of the precursor of the "unique" meteorites is higher than the value reported for Y-74063. Difference in trace element concentrations in the "U" chondrite and ordinary chondrites exhibits that the "U" chondrites were formed at the different region in the solar nebula where ordinary chondrites were formed. This is supported by difference in oxygen isotopic compositions of the "unique" meteorites from those of ordinary chondrites (Fig. 12).

High abundance of rare gases in both the "unique" meteorites (PALME et al., 1981; SCHULTZ et al., 1982; TAKAOKA and YOSHIDA, 1991) and ureilites (GÖBEL et al., 1978) is a problem. Recent study on rare gases have revealed that rare gases are enriched not in grain boundary but in all minerals including olivine, pyroxenes, kamacite, and troilite in Y-74063 (TAKAOKA et al., 1991) and in a magnetic separate in Acapulco (MARTI et al., 1991). This enabled the high abundance of rare gases even after magmatic processes, but the duration of melting should have been short enough to retain gases.

7. Role of Reduction

The compositions of minerals indicate that Y-8002 and Y-75274 should have formed through a high degree of partial melting from a precursor "U" chondrite with a composition similar to Y-74063. The distribution of Mg and Fe between melt and olivine, however, shows that Fo_{97} olivine cannot be formed from Fo_{87} of Y-74063. It has been well established that distribution of Mg and Fe between olivine and melt $(K_D = (Fo/Fa) \cdot (X_{Fe}^{Liq}/X_{Mg}^{Liq}))$ is about 0.3 regardless of the composition of the system and temperature and pressure (ROEDER and EMSLIE, 1970; LONGHI et al., 1978; TAKAHASHI, 1978). The most magnesian olivine which can be formed from Fo_{87} olivine is calculated to be $Fo_{95.7}$, which is a little less magnesian than Fo_{97} of Y-8002 and Y-75274. Or conversely, the Mg/(Mg+Fe) ratio of the melt which coexists with Fo_{97} olivine is 0.91, which is more magnesian than the 0.87 of the "U" chondrite. This means that olivine in Y-8002 and Y-75274 cannot be formed even through total melting of the "U" chondrite. Reduction of the liquid during melting would be important for the origin of highly magnesian olivine. The model of crystallization of Y-8002 and Y-75274 under reducing conditions is consistent with the presence of graphite inclusions in kamacite in Y-8002 and Y-

75274. Forsterite 95 olivine can coexist with orthopyroxene and metallic iron at 10^{-14}–10^{-17} atm f_{O_2} at magmatic temperatures (1000°–1200°C) (NITSAN, 1974). Graphite is stable at this f_{O_2} even at 1 bar, and will be stable at higher pressure plausible for a magma stream in a small body. The estimated f_{O_2} is a little above the measured f_{O_2}-T curve for pallasites which is in turn a little higher than that for EL6 chondrites (BRETT and SATO, 1984). This f_{O_2} condition is 1 to 1.5 orders of magnitude lower than that for ordinary chondrites.

Reduction of liquid during partial melting should have been quite different from that at subsolidus temperatures. Reduced metal could have been separated from silicate liquid by dissolving into metallic Fe-Ni, resulting in the apparent absence of evidence for reduction in solid phases.

Reduction of partial melt may affect the Mg-Fe distribution between minerals. The relationship between Fo content of olivine and Cr_2O_3 content of spinel of the "U" group meteorites is inconsistent with previous studies on peridotite melting (JAQUES and GREEN, 1980; DICK and BULLEN, 1984). Partial melting largely changes the Cr/Al ratio of spinel, but changes slightly the Mg/Fe ratio of olivine when the degree of melting is not extremely high (>50%). The Cr/(Cr+Al) ratio of spinel varies from 0.87 of Y-74063 and ALH-77081 to 0.91 of Y-8002, on the contrary, the Fo content from 0.87 to 0.97. Variation of Fo content is too big compared to that of spinel Cr/(Al+Cr) ratio, which may be also due to crystallization of highly magnesian olivine.

8.   Conclusions

Petrological study of Y-8002, a "unique" meteorite, has shown that it is a group of non-classified meteorites which includes Y-74063, Y-75274, ALH-77081, Y-791493, Acapulco, and Lodran. Y-74357, Y-78230, and MAC88177 may belong to this clan, though data are insufficient to be conclusive. X_{Mg} of pyroxenes and spinel, the An content of plagioclase, and the Cr_2O_3 contents of spinel and orthopyroxene in these meteorites are positively related to the Fo content of olivine, and the MnO and TiO_2 contents of pyroxenes and spinel are inversely related to the Fo content. These relationship are formed through a partial melting process and the differences in mineral compositions are due to the difference in degree of partial melting from a common precursor. Y-8002 and Y-75274 were formed through a high degree of melting under low f_{O_2} conditions intermediate between ordinary chondrites and pallasites, resulting in crystallization of highly magnesian olivine, pyroxenes, and spinel. The other meteorites were formed through lower degrees of partial melting. This model is consistent with the REE patterns. Extraction of a partial melt was effective during melting because of an easily formed melt network. Melting took place in flows of melt heterogeneously in a parent body, rather than in a stagnant floor of a magma reservoir. The precursor is a kind of chondrite having chondrules with composition more magnesian than any ordinary chondrites but less magnesian and E chondrites. Because of unique composition of minerals, the precursor chondrite is tentatively called as "U" chondrite. Mineral composition of the "U" chondrite was estimated from the mineral trends during partial melting.

The "U" chondrite is richer in Cr and Mn than the composition estimated from fractionation trends responsible for chemical group of ordinary chondrites, thus, precursor material of the "U" chondrite was formed in a different region in the solar nebula from ordinary chondrites. Oxygen isotopic compositions of the "U" meteorites plot in a narrow range below the terrestrial fractionation line, supporting their formation in a location different from that for ordinary chondrites.

Acknowledgments

References

BERKLEY, J. L., TAYLOR, G. J., KEIL, K. HARLOW, G. E. and PRINZ, M. (1980): The nature and origin of ureilites. Geochim. Cosmochim. Acta, **44**, 1579–1597.

BILD, R. W. and WASSON, J. T. (1976): The Lodran meteorite and its relationship to the ureilites. Mineral. Mag., **40**, 721–735.

BRETT, R. and SATO, M. (1984): Intrinsic oxygen fugacity measurements on seven chondrites, a pallasite, and a tektite and a redox state of meteorite parent bodies. Geochim. Cosmochim. Acta, **48**, 111–120.

BUNCH, T. E. and OLSEN, E. (1973): Ortho-clinopyroxene compositions in ordinary chondrites and related Blander model calculation procedures. NASA Tech. Memo. NASA TMX-62, **259**, 16 p.

BUNCH, T. E., KEIL, K. and SNETSINGER, K. (1967): Chromite composition in relation to chemistry and texture of ordinary chondrites. Geochim. Cosmochim. Acta, **31**, 1569–1582.

CLAYTON, R. N. and MAYEDA, T. K. (1978): Genetic relations between iron and stony meteorites. Earth Planet. Sci. Lett., **40**, 168–174.

CLAYTON, R. N. and MAYEDA, T. K. (1988): Formation of ureilites by nebular processes. Geochim. Cosmochim. Acta, **52**, 1313–1318.

CLAYTON, R. N., MAYEDA, T. K. and YANAI, K. (1984): Oxygen isotopic compositions of some Yamato meteorites. Mem. Natl Inst. Polar Res., Spec. Issue, **35**, 267–271.

DICK, H. J. B. and BULLEN, T. (1984): Chromian spinel as a petrogenetic indicator in abyssal and alpine-type peridotites and spatially associated lavas. Contrib. Mineral. Petrol., **86**, 54–76.

DUKE, J. M. (1976): Distribution of the period four transition elements among olivine, calcic clinopyroxene and mafic silicate liquid: Experimental results. J. Petrol., **17**, 499–521.

DUNN, T. (1987): Partitioning of Hf, Lu, Ti, and Mn between olivine, clinopyroxene and basaltic liquid. Contrib. Mineral. Petrol., **96**, 476–484.

FABRIES, J. (1979): Spinel-olivine geothermometry in peridotites from ultramafic complexes. Contrib. Mineral. Petrol., **69**, 329–336.

FUKUOKA, T. and KIMURA, M. (1990): Chemistry of Y-74063, -74357 and ALH-78230 unique meteorites (abstract). Papers Presented to the 15th Symposium on Antarctic Meteorites, May 30–June 1, 1990. Tokyo, Natl Inst. Polar Res., 99–100.

FUKUOKA, T., MA, M.-S., WAKITA, H. and SCHMITT, R. A. (1978): Lodran: The residue of limited partial melting of matter like a hybrid between H and E chondrites. Lunar and

Planetary Science IX. Houston, Lunar Planet. Inst., 356–358.

GÖBEL, R., OTT, U. and BEGEMANN, F. (1978): On trapped noble gases in ureilites. J. Geophys. Res., 83, 855–867.

GOLDSTEIN, J. I. and DOAN, A. S., JR. (1972): The effect of phosphorus on the formation of the Widmanstätten pattern in iron meteorites. Geochim. Cosmochim. Acta, 36, 51–69.

GOODRICH, C. and BERKLEY, J. L. (1985): Minor elements in ureilites: Evidence for reverse fractionation and interstitial silicate liquid. Lunar and Planetary Science XVI. Houston, Lunar Planet. Inst., 280–281.

GOODRICH, C. A., JONES, J. H. and BERKLEY, J. L. (1987): Origin and evolution of the ureilite parent magmas: Multi-stage igneous activity on a large parent body. Geochim. Cosmochim. Acta, 51, 2255–2273.

GOODRICH, C. A., PATCHETT, J. J., LUGMAIR, G. W. and DRAKE, M. J. (1991): Sm-Nd and Rb-Sr isotopic systematics of ureilites. Geochim. Cosmochim. Acta, 55, 829–848.

HARAMURA, H., KUSHIRO, I. and YANAI, K. (1983): Chemical compositions of Antarctic meteorites I. Mem. Natl Inst. Polar Res., Spec. Issue, 30, 109–121.

HEWINS, R. H. (1986): Serial melting or magma ocean for the HED achondrites? (abstract). Meteoritics, 21, 396–397.

HIROI, T. and TAKEDA, H. (1991): Reflectance spectroscopy and mineralogy of primitive achondrites-lodranites. Proc. NIPR Symp. Antarct. Meteorites, 4, 163–177.

HOOVER, J. D. (1989): Petrology of the Marginal Border Series of the Skaergaard Intrusion. J. Petrol., 30, 399–439.

IKEDA, Y. and TAKEDA, H. (1985): A model for the origin of basaltic achondrites based on the Yamato-7308 howardite. Proc. Lunar Planet. Sci. Conf., 15th, Pt. 2, C649–C663 (J. Geophys. Res., 90, Suppl.).

IRVING, A. J. (1978): A review of experimental studies of crystal/liquid trace element partitioning. Geochim. Cosmochim. Acta, 42, 743–770.

JAQUES, A. L. and GREEN, D. H. (1980): Anhydrous melting of peridotites at 0–15 kb pressure and the genesis of tholeiitic basalts. Contrib. Mineral. Petrol., 73, 287–310.

KALLEMEYN, G. W. and WASSON, J. T. (1985): The compositional classification of chondrites: IV. Ungrouped chondritic meteorites and clasts. Geochim. Cosmochim. Acta, 49, 261–270.

KANEOKA, I., TAKAOKA, N. and YANAI, K. (1991): ^{40}Ar-^{39}Ar analyses on antarctic meteorites Y-74063 and ALH-78230: Consortium study on unique chondrites from Antarctica (abstract). Papers Presented to the 16th Symposium on Antarctic Meteorites, June 5–7, 1991. Tokyo, Natl Inst. Polar Res., 90–91.

KIMURA, M. (1987): Petrology of unique Antarctic chondrites, Y-74025 and Y-74063 (abstract). Papers Presented to the 12th Symposium on Antarctic Meteorites, June 8–10, 1987. Tokyo, Natl Inst. Polar Res., 47.

KIMURA, M., TSUCHIYAMA, A., FUKUOKA, T. and IIMURA, Y. (1991): Mineralogy and chemistry of Antarctic winonaites: Y-74025, Y-75300 and Y-75305 (abstract). Papers Presented to the 16th Symposium on Antarctic Meteorites, June 5–7, 1991. Tokyo, Natl Inst. Polar Res., 85–86.

KRACHER, A. (1988): Metal with anomalously low Ni and Ge concentrations in the Allan Hills A-77081 winonaite. Proc. Lunar Planet. Sci. Conf., 18th, 485–491.

LARIMER, J. W. (1968): Experimental studies on the system Fe-MgO-SiO₂-O₂ and their bearing on the petrology of chondritic meteorites. Geochim. Cosmochim. Acta, 32, 1187–1209.

LINDSLEY, D. H. and ANDERSEN, D. J. (1983): A two-pyroxene thermometer. Proc. Lunar Planet. Sci. Conf., 13th, Pt. 2, A887–A906 (J. Geophys. Res., 88 Suppl.).

LINDSTROME, D. L. and WEILL, D. F. (1978): Partitioning of transition metals between diopside and coexisting silicate liquid. I. Nickel, cobalt, and manganese. Geochim. Cosmochim. Acta, 42, 817–831.

LONGHI, J., WALKER, D. W. and HAYS, J. F. (1978): The distribution of Fe and Mg between olivine and lunar basaltic liquids. Geochim. Cosmochim. Acta, 42, 1545–1558.

MARTI, K., KIM, J. S., KIM, Y., LAVIELLE, B. and PRINZ, M. (1991): Progress in the Acapulco consortium (abstract). 54th Annual Meeting of the Meteoritical Society. July 21–26, Montrey, California, 148.

MATSUMOTO, Y., YANAI, K. and KOJIMA, H. (1983): A stoney-iron meteorite: Yamato-75274 (abstract). Papers Presented to the 8th Symposium on Antarctic Meteorites, March 17–19, 1982. Tokyo, Natl Inst. Polar Res., 7.

MAYEDA, T. K. and CLAYTON, R. N. (1980): Oxygen isotopic compositions of aubrites and some unique meteorites. Proc. Lunar Planet. Sci. Conf., 11th, 1145–1151.

MAYEDA, T. K. and CLAYTON, R. N. (1989): Oxygen isotopic compositions of unique Antarctic meteorites (abstract). Papers Presented to the 14th Symposium on Antarctic Meteorites, June 6–8, 1989. Tokyo, Natl Inst. Polar Res., 172.

MAYEDA, T. K., CLAYTON, R. N. and YANAI, K. (1987): Oxygen isotopic compositions of several Antarctic meteorites. Mem. Natl Inst. Polar Res., Spec. Issue, 46, 144–150.

MORI, H., TAKEDA, H., PRINZ, M. and HARLOW, G. E. (1984): Mineralogical and crystallographic studies of lodranite and primitive achondrite groups bearing on their genetic link. Lunar and Planetary Science XV. Houston, Lunar Planet. Inst., 567–568.

MUKHERJEE, A. B. and BULATOV, V. (1989): New high P-T experimental results on orthopyroxene-chrome spinel equilibrium and revised orthopyroxene-spinel cosmothermometer. Lunar and Planetary Science XX. Houston, Lunar Planet. Inst., 731–732.

NAGAHARA, H. and OZAWA, K. (1986): Petrology of Yamato-791493, "lodranite": Melting, crystallization, cooling history, and relationship to other meteorites. Mem. Natl Inst. Polar Res., Spec. Issue, 41, 181–205.

NAGAHARA, H., FUKUOKA, T., KANEOKA, I., KIMURA, M., KOJIMA, H., KUSHIRO, I., TAKEDA, H., TSUCHIYAMA, A. and YANAI, K. (1990): Petrology of unique meteorites, Y-74063, Y-74357, Y-75261, Y-75274, Y-75300, Y-75305, A-77081, A-78230 and Y-8002 (abstract). Papers Presented to the 15th Symposium on Antarctic Meteorites, May 30–June 1, 1990. Tokyo, Natl Inst. Polar Res., 92–94.

NITSAN, U. (1974): Stability field of olivine with respect to oxidation and reduction. J. Geophys. Res., 79, 706–711.

OZAWA, K. (1984): Olivine-spinel geospeedometry; Analysis of diffusion-controlled Mg-Fe^{2+} exchange. Geochim. Cosmochim. Acta, 48, 2597–2611.

PALME, H., SCHULTZ, L., SPETEL, B., WEBER, H. W., WÄNKE, H., CHRISTOPHE MICHEL-LEVY, M. and LORIN, J. C. (1981): The Acapulco meteorite: Chemistry, mineralogy, and irradiation effects. Geochim. Cosmochim. Acta, 45, 727–752.

PRINZ, M., KLIMENTIDIS, R., HARLOW, G. E. and HEWINS, R. H. (1978): Petrologic studies bearing on the origin of the Lodran meteorite. Lunar and Planetary Science IX. Houston, Lunar Planet. Inst., 919–921.

PRINZ, M., NEHRU, C. E., DELANEY, J. S. and WEISBERG, M. (1983): Silicates in IAB and IIICD irons, winonaites, lodranites and Brachina: A primitive and modified-primitive group. Lunar and Planetary Science XIV. Houston, Lunar Planet. Inst., 616–617.

PRINZ, M., CHATTERJEE, N., WEISBERG, M. K., CLAYTON, R. N. and MAYEDA, T. K. (1991): MAC88177: A new type of achondrite? Lunar and Planetary Science XXII. Houston, Lunar Planet. Inst., 1099–1100.

RAY, G. L., SHIMIZU, N. and HART, S. R. (1983): An ion microprobe study of the partitioning of trace elemental between clinopyroxene and liquid in the system diopside-albite-anorthite. Geochim. Cosmochim. Acta, 47, 2131–2140.

ROEDER, P. L. and EMSLIE, R. F. (1970): Olivine-liquid equilibrium. Contrib. Mineral. Petrol., 29, 275–289.

ROEDER, P. L., CAMPBELL, I. and JAMIESSON, E. (1979): A re-evaluation of the olivine-spinel geothermometer. Contrib. Mineral. Petrol., 68, 325–334.

ROMIG, A. D., JR. and GOLDSTEIN, J. I. (1981): Low temperature phase equilibria in the Fe-Ni and Fe-Ni-P systems: Application to the thermal history of metallic phases in meteorites. Geochim. Cosmochim. Acta, 45, 1187–1197.

RUBIN, A. (1990): Kamacite and olivine in ordinary chondrites: Intergroup and intragroup relationships. Geochim. Cosmochim. Acta, 54, 1217–1232.

SACK, R. O., AZZEREDO, W. J. and LIPSCHUTZ, M. E. (1991): Olivine diogenites: The mantle of the eucrite parent body. Geochim. Cosmochim. Acta, 55, 1111–1120.

SCHULTZ, L., PALME, H., SPETTEL, B., WEBER, H. W., WÄNKE, H., CHRISTOPHE MICHEL-LEVY, M. and LORIN, J. C. (1982): Allan Hills 77081—an unusual meteorite. Earth Planet. Sci.

Lett., **61**, 23–31.

STOLPER, E. (1977): Experimental petrology of eucritic meteorites. Geochim. Cosmochim. Acta, **41**, 587–611.

STOLPER, E., MCSWEEN, H., JR. and HAYS, J. F. (1979): A petrogenetic model of the relationships among achondritic meteorites. Geochim. Cosmochim. Acta, **43**, 589–602.

TAKAHASHI, E. (1978): Partitioning of Ni^{2+}, Co^{2+}, Fe^{2+}, Mn^{2+}, and Mg^{2+} between olivine and silicate melts: compositional dependence of partition coefficient. Geochim. Cosmochim. Acta, **42**, 1829–1844.

TAKAHASHI, E. (1983): Melting of a Yamato L3 chondrite (Y-74191) up to 30 kbar. Mem. Natl Inst. Polar Res., Spec. Issue, **30**, 168–180.

TAKAHASHI, E. and KUSHIRO, I. (1983): Melting of dry peridotite at high pressures and basalt magma genesis. Am. Mineral., **68**, 859–879.

TAKAOKA, N. and YOSHIDA, Y. (1991): Noble gas composition in unique meteorite, Yamato-74063. Proc. NIPR Symp. Antarct. Meteorites, **4**, 178–186.

TAKAOKA, N. NAGAO, K. and MIURA, Y. (1991): Noble gas study of unique meteorite Yamato-74063 by laser extraction (abstract). Papers Presented to the 16th Symposium on Antarctic Meteorites, June 5–7, 1991. Tokyo, Natl Inst. Polar Res., 92.

TAKEDA, H., SAITO, J., MIYAMOTO, M. and HIROI, T. (1991): Mineralogy of MAC88177 and comparison with augite-bearing lodranite, Yamato-74357. Lunar and Planetary Science XXII. Houston, Lunar Planet. Inst., 1375–1376.

VAN SCHMUS, W. R. (1969): The mineralogy and petrology of chondritic meteorites. Earth Sci. Rev., **5**, 145–184.

WAGER, L. R. and BROWN, G. M. (1968): Layered Igneous Rocks. London, Oliver & Boyd, 588 p.

WALKER, D., JUREWICZ, S. and WATSON, E. B. (1988): Adcumulus dunite growth in a laboratory thermal gradient. Contrib. Mineral. Petrol., **99**, 306–319.

WARREN, P. H. (1985): Origin of howardites, diogenites, and eucrites: A mass balance constraint. Geochim. Cosmochim. Acta, **51**, 713–725.

WARREN, P. H. and KALLEMEYN, G. W. (1991): Geochemistry of unique achondrite MAC88177: comparison with polymict ureilite EET87720 and "normal" ureilites. Lunar and Planetary Science XXII. Houston, Lunar Planet. Inst., 1467–1468.

WATSON, E. B. (1977): Partitioning of manganese between forsterite and silicate liquid. Geochim. Cosmochim. Acta, **41**, 1363–1374.

YAMAMOTO, K., NAKAMURA, N., MISAWA, K., KOJIMA, H. and YANAI, K. (1991): Lithophile trace elements and Sr isotopes in unique meteorites, Y-75063, ALH-78230, Y-74357, Y-8002 and Y-75300 (abstract). Papers Presented to the 16th Symposium on Antarctic Meteorites, June 5–7, 1991. Tokyo, Natl Inst. Polar Res., 87–89.

YANAI, K. and KOJIMA, H. (1991): Yamato-74063: Chondritic meteorite classified between E and H chondrite groups. Proc. NIPR Symp. Antarct. Meteorites, **4**, 118–130.

YANAI, K., KOJIMA, H., PRINZ, M., NEHRU, C. E., WEISBERG, M. K. and DELANEY, J. S. (1984): Petrologic studies of three primitive achondrites from the Yamato meteorites collection, Antarctica (abstract). Papers Presented to the 9th Symposium on Antarctic Meteorites, March 22–24, 1984. Tokyo, Natl Inst. Polar Res., 24–28.