

The Chemical Composition, Petrography and Mineralogy of the Japanese Chondrite Fukutomi

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I. Introduction

The chondrite Fukutomi fell at one o'clock in the afternoon on March 19, 1882 in Fukutomi-mura, Kishima-gun, Saga-ken, Kyushu, Japan at $33^{\circ}11'N.$, and $130^{\circ}12'E.$ The description of the circumstances of the fall, brief petrographical and mineralogical description and overall chemical analysis have been carried out by Miyashiro *et al.* (1966). Keil and Fredriksson (1974) and Van Schmus and Wood (1967) observed also thin sections of this chondrite, and the latter authors classified this chondrite to be L5. However this classification, especially subgroup is still doubtful.

Because, the chondrite Fukutomi has peculiar structure, inhomogeneous distribution of large minerals, coexistence of high and low temperature minerals and so on, it is necessary to reexamine the Fukutomi chondrites from various directions.

II. Chemical Composition

The chemical composition of the Fukutomi chondrite was determined by the same method employed in the determination of the chondrites Nagai (Murayama *et al.*, 1978), Kamiomi (Okada *et al.*, 1979) and Shibayama (Shima *et al.*, 1979).

The results of chemical fractional dissolution are shown in Table 1. Among L group chondrites, besides the Fukutomi, only two chondrites Nagai (Murayama *et al.*, 1978) and Shibayama (Shima *et al.*, 1979) have been subjected to this treatment. Because the Shibayama was highly weathered, the present results could be compared with that for Nagai.

All lithophile elements in Fukutomi examined are more soluble in Br_2 -aqua regia than those in Nagai. Further, 1.3% of Mg and Mn in Fukutomi were dissolved in

Table 1. Relative abundances from stepwise dissolution.

Element (%)	Fraction (%)		
	CuCl ₂ -KCl	Br ₂ +Aqua regia	Residue
Mg	14.65	1.29	68.54
Al	1.12	~0	13.7
Cr	0.342	~0	4.1
Mn	0.248	1.3	60.6
Ti	0.062	~0	39.0
Fe	22.78	37.5	53.1
Ni	1.25	65.1	34.9
Co	0.071	78.4	21.6
P	0.13	~0	100

CuCl₂-KCl solution while 0.35% of Mg and ~0% of Mn in Nagai was in the same solution. Compared with lithophile elements, the solubilities of siderophile elements such as Ni and Co in Fukutomi for CuCl₂-KCl solution were less than those in Nagai. Miyashiro *et al.* (1966) reported that extraordinary part of Fukutomi contains very high, about twice normal content, of sulfur. These facts suggest that soluble Mg and Mn and insoluble Ni and Co in CuCl₂-KCl solution may be composed of sulfide minerals such as Niningerite, Alabandite and Pentlandite like Type I enstatite chondrites Abee and Indarch (Mason, 1966; Shima and Honda, 1976; 1974) and Yamato (a) (Shima, 1974) rather than oxide minerals. This means that the chondrite Fukutomi contains low temperature minerals or rather primordial, less metamorphosed minerals.

The chemical composition in conventional form and CIPW norm are shown in Table 2. In this table, the contents of trace metals in Fukutomi obtained by emission spectrograph with an inductively coupled radio-frequency plasma, model ICPQ-100, manufactured by Shimadzu Seisakusho Ltd., are presented. These data of trace metals were taken from the meteorite solution without any chemical procedures, and gave reasonable values as ordinary chondrites.

Bulk chemical composition and the norm are compared with the values of Fukutomi No. 1 by Miyashiro *et al.* (1966). They were determined for three specimens, out of which two of them they call ordinary part and another one is extraordinary part. Further, they took as much as 4.6 g as sample for No. 1, 12.4 g for the ordinary part of No. 2 and 3.9 g for extraordinary part of No. 2. Usually if ~10 g samples were taken for analyses, the data should give mean value of the chondrite even if it is quite heterogeneous. When their two data for ordinary part of No. 1 and No. 2 are compared with each other, some elemental compositions, such as Al, K, P, Cr and S are quite different from each other. Their Co values are unbelievably low. These facts suggest that the chondrite Fukutomi is extremely inhomogeneous and unusual or their analyses are inconsistent in some extent.

Present data, however, shows fairly good agreements with the mean values or

Table 2. Chemical composition and the CIPW norm of the chondrite Fukutomi compared with previous work*.
(weight percent)

	this work	previous work		this work	previous work
SiO ₂	39.46	38.96	Olivine	44.76	41.20
MgO	24.29	25.33	Pyroxene	En	18.26
FeO	13.36	11.62		Fs	4.89
Al ₂ O ₃	2.11	2.30		Wo	2.35
CaO	1.70	1.65	Ilmenite	0.20	0.22
Na ₂ O	1.01	0.90	Chromite	0.74	0.90
K ₂ O	0.11 _s	0.18	Feldspar	Or	0.72
Cr ₂ O ₃	0.50	0.60		Ab	8.55
MnO	0.32	0.32		An	0.86
TiO ₂	0.10 ₃	0.11	Apatite	0.69	0.61
P ₂ O ₅	0.30	0.25	Nickel-Iron	9.82	11.18
H ₂ O	0.21	0.21	Troilite	5.96	6.37
Fe	8.50	9.83	Free Silica	1.26	—
Ni	1.25	1.33	H ₂ O	0.21	
Co	0.071	0.02			
FeS	5.96	6.37			
Total	99.26	99.98		99.27	99.84
Total Fe	22.78	22.92			
Trace Metals (ppm)			Ratios		
V	39.7		Fe _{total} /SiO ₂	0.577	0.588
Cu	95.8		Fe _{metal} /Fe _{total}	0.373	0.429
Zn	80.1		Fa	{ 26	24**
Sr	12.7			{ 23**	
Zr	9.73		SiO ₂ /MgO	1.62	1.54

* Miyashiro *et al.*, (1966). ** Estimated by optics.

one of the values of ordinary parts of No. 1 and No. 2 of their values. The only differences are the data for Co and the dividing of Fe to FeO and metallic Fe. They have not stated how their results were obtained in their paper. One reason for the discrepancy in Fe is that meteorite sample itself has been oxidized or weathered during these 15 years. Another reason, which is more likely, is that our dividing method is experimental so it should be more reliable than earlier work. In conclusion, it could be said that the chondrite Fukutomi is really heterogeneous but even so we could obtain the mean chemical composition with about 1~2 g of sample.

The norm in this paper has been calculated from chemical composition with modified method, that is, the ratio of olivine to pyroxene is obtained from the data of Mg and Fe in Table 1, data obtained by chemical fractional dissolution. By this method, one can account for free silica. This free silica may consist of tridymite in the following section.

In Table 2, the ratios of Fe_{total}/SiO_2 , Fe_{metal}/Fe_{total} , SiO_2/MgO and fayalite, are tabulated. When these values are compared with the indicator values of the chemical classification of chondrites by Van Schmus and Wood (1967), all of present values and the three values computed from the Miyashiro *et al.*, indicate that the chondrite Fukutomi belongs to L-group chondrite. The ratio of Fe_{metal}/Fe_{total} by Miyashiro *et al.*, is too high. This suggests also that their method of dividing Fe in two parts is not reasonable.

The fayalite ratio obtained by optics are also shown. The data are variable but all three data are in the range of L group chondrites. These variable fayalite ratios may suggest that this chondrite does not belong to subgroups 5 or 6.

III. Petrography an Mineralogy

Polished sections and polished thin sections of the meteorite specimen were prepared for the petrographical and mineralogical investigations using a polarizing microscope. The optic axial angles and optical orientations of transparent minerals were measured with an aid of a universal stage which was set on the microscope stage. Semiquantitative X-ray microanalysis was carried out with a JSM-U scanning electron microscope equipped with an X-ray detection system. Debye-Scherrer photographs of separated mineral grains and X-ray powder patterns of the pulverized meteorite sample were also investigated for the identification of mineral species.

According to works by Keil and Fredriksson (1964), Miyashiro *et al.* (1966) and Van Schmus and Wood (1967), this meteorite is an L-group chondrite, and is composed of olivine, orthopyroxene, clinopyroxene, plagioclase, nickel-iron, troilite, chromite and devitrified glass. In general, the present result agrees with these works, but further studies in this work showed that the Fukutomi meteorite is a chondrite including xenoliths, and that several accessory minerals, tridymite, apatite and native copper

Table 3. Modal composition of the Fukutomi chondrite*
(volume percent)

	Matrix	Chondrule
Olivine	61	57
Orthopyroxene	13	13
Clinopyroxene	10	10
Plagioclase	—	trace
Tridymite**	2	—
Opaque minerals	10	2
Others***	4	18
Total	100	100

* The volume ratio of the matrix and chondrules are 75% and 25% respectively.

** Tridymite is a rare mineral in ordinary chondrites, but significant amounts of tridymite are present in the thin section used in the present work.

*** Devitrified glass and very fine-grained material.



Fig. 1. A thin section of the Fukutomi meteorite. Chondritic structure is well-developed. Scale bar is 0.5 mm.

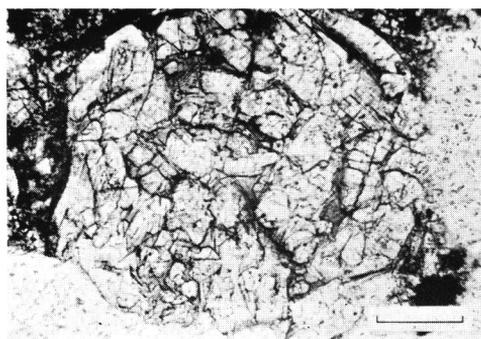


Fig. 2. A microporphyritic chondrule consisting of olivine, orthopyroxene and clinobronzite crystals. The interstices between the crystals are filled with devitrified glass, pale brown in color. Scale bar is 0.1 mm.

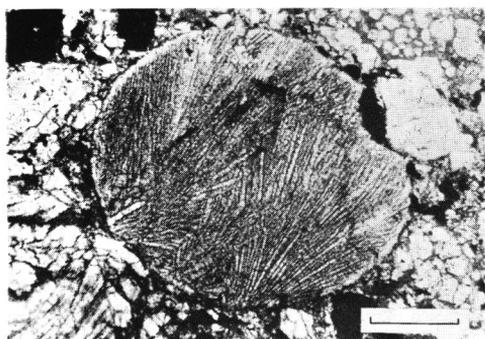


Fig. 3. A chondrule consisting of radially elongated olivine crystals. Scale bar is 0.2 mm.

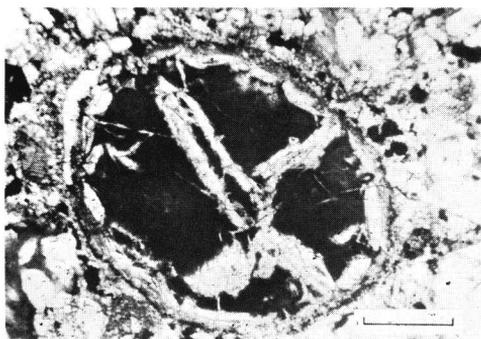


Fig. 4. This chondrule is mainly composed of devitrified material, dark brown in color, and olivine crystals. Scale bar is 0.2 mm.

are present. The modal composition of minerals in the thin section, about 1 cm² in total area, is shown in Table 3. This meteorite shows a distinct chondritic structure (Fig. 1). Chondrules generally range from 0.2 mm to 2 mm in size, and reach 4 mm in maximum size. Their outlines are easily recognized in the matrix, and several of them are surrounded by a dark, fine-grained rim. In the thin section, microporphyritic chondrule is most common (Fig. 2). In this type of chondrule, mineral assemblages of phenocrysts are olivine only, olivine-clinobronzite-bronzite, olivine-clinobronzite and bronzite-clinobronzite. Radial chondrules consisting of preferentially elongated olivine, orthopyroxene and clinopyroxene are also common (Fig. 3). In addition, devitrified glass-rich chondrules are also present (Fig. 4). The interstitial materials between phenocrysts in chondrules are devitrified glass, brown and dark brown

in color, microcrystalline fibrous material, dark brown in color, and finely acicular pyroxene-like crystals (Fig. 5). Micron-sized opaque grains consisting of chromite, troilite and nickel-iron are scattered in the interstitial materials.

The matrix surrounding chondrules presents a clastic appearance. It consists of inequigranular aggregate composed mainly of olivine, pyroxene and opaque grains, and encloses many chondrule fragments, varying in size and in shape. In places, the matrix material is microcrystalline and microgranular, devitrified glassy material being scattered. In the Fukutomi specimen for making thin sections, two kinds of lithic fragments were observed. One is dark and nearly black in color, and the other is white-colored. Fig. 6 shows the microscopic appearance of the thin section of the

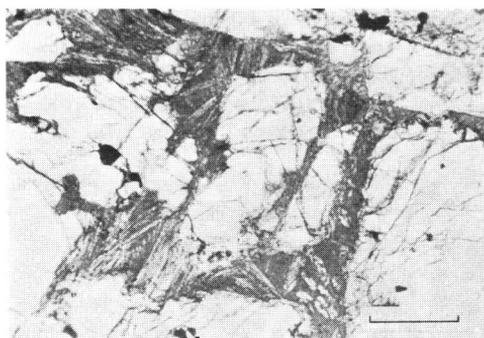


Fig. 5. The interstices of an olivine chondrule. They are filled with acicular, pyroxene-like materials, showing oblique extinction. Scale bar is 0.1 mm.

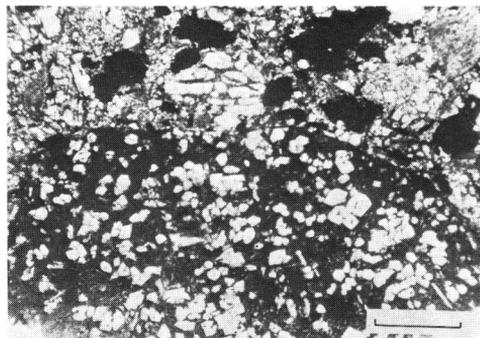


Fig. 6. A dark-colored lithic fragment occurs in the lower part of the photograph. It is chiefly composed of euhedral olivines and dark fine grained matrix. Scale bar is 0.5 mm.

former material. This fragment is a few millimeters or more in size, and possesses a similar mineralogical and textural feature to that of an olivine microporphyritic chondrule. This is chiefly composed of olivine crystals and dark, fine-grained matrix. The olivine phenocrysts are generally euhedral in shape, and some of them are corroded in their margin. Clinobronzite which is polysynthetically twinned is rarely seen in the phenocrysts. The dark matrix consists of minute olivine granules, fine aggregate of acicular pyroxene crystals and dark-colored devitrified glass which includes very fine-grained opaque grains (Fig. 7). Very fine, skeletal crystals of clinopyroxene-like material showing oblique extinction are also present in places between phenocrysts. Opaque grains are generally very small in size, but ovoidal form of troilite grains are present in some dark fragments, testifying the liquid immiscibility of sulfide and silicate. The other kind of lithic fragment, white in color, is composed of an aggregate of tridymite crystals, and is marginally surrounded with lamellar clinopyroxenes (Fig. 8).

Olivine: Olivine is the dominant mineral species both in the matrix and chondrules.

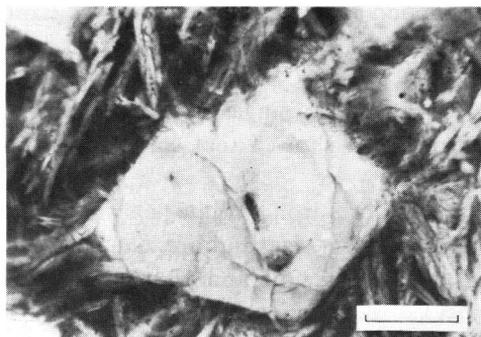


Fig. 7. Enlargement of a part of the dark-colored lithic fragment. The matrix surrounding olivine grains is composed of very fine-grained pyroxenes and dark brown, devitrified glass including tiny opaque grains. Scale bar is 0.03 mm.

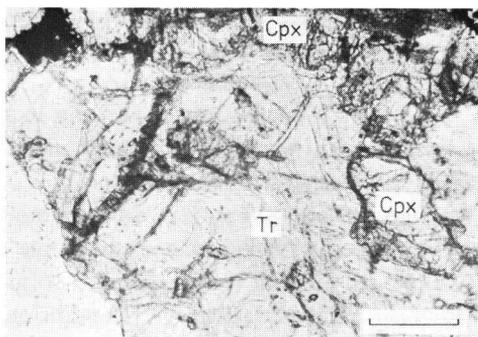


Fig. 8. A lithic fragment consisting of tridymite (Tr). The outer part is rimmed with clinopyroxenes (Cpx). Scale bar is 0.1 mm.

The grains, 0.8 mm in maximum size, are mostly angular. In chondrules and dark lithic fragments, the olivines exhibit a euhedral crystal form, and several grains adjacent to the devitrified glass and dark brown, microcrystalline material possess a thin reaction rim of pyroxene. Between the crossed nicols, olivine grains show undulatory extinction, mosaic extinction and kink bands. The optic axial angle ($2V$) is -86° on average, and refractive indices are $\alpha=1.675$ and $\gamma=1.719$. The molar composition estimated by optics is Fa_{23} .

Pyroxene: In the pyroxene species, orthopyroxene (bronzite) and clinobronzite are two major constituents. Orthopyroxene is the most abundant mineral next to olivine, and exhibits euhedral to subhedral crystal shape. The grains are colorless, and range up to 0.4 mm in size. In the cross polarized light, undulatory extinction is distinct. Optic axial angle ($2V$) is -75° on average, and refractive indices are $\alpha=1.678$ and $\gamma=1.690$. According to the determinative curve by Kuno (1954), the molar composition is estimated to be Fs_{22} . Clinobronzite is distinguished by the presence of polysynthetic twin lamellae. The width of lamellae ranges from 0.4 to 10 μm . The extinction angle ($c \wedge Z$) is about 44° . Another clinopyroxene species is augite. It is a minor constituent, and is generally present as a rim around the host crystals of orthopyroxene and clinobronzite. The extinction angle ($c \wedge Z$) is $36^\circ-39^\circ$.

Plagioclase: Plagioclase is a very minor mineral constituent of this chondrite, and occurs as tiny grains, less than 10 μm in size, in the interstices between olivine and pyroxene grains.

Tridymite: The distribution of tridymite is very inhomogeneous. It is only found in the lithic fragment forming a cluster-like aggregate, ranging up to 1 mm in size. The individual grains are 0.3 mm in maximum size, and most of them are twinned. The refractive indices are $\alpha=1.470$ and $\gamma=1.474$. The d-spacings (\AA) and visually

estimated intensities of the distinct X-ray diffraction lines in the Debye photograph are as follows; 4.32 (vs), 4.11 (s), 3.80 (m), 3.25 (vw), 3.01 (vw), 2.97 (w), 2.78 (vw), 2.50 (w), 2.39 (vw) and 2.30 (w).

Apatite: Apatite, about 0.1 mm in size, is present as a single grain in the matrix. It is uniaxially negative, and electron microprobe scan showed that chlorine is contained in the chemical composition.

Opaque minerals: The opaque minerals are nickel-iron, troilite, chromite and native copper. Nickel-iron grains range up to 0.4 mm in size, and are xenomorphic. By chemical etching using 5% nital solution, kamacite, taenite and plessite areas are clearly distinguished (Fig. 9). In addition, most kamacite grains show a well-developed Neumann structure produced by the shock effect. Troilite occurs as xenomorphic grains ranging up to 0.4 mm in size. Between crossed nicols, several grains show that they are composed of multiple sets of differently oriented crystals. Chromite ranges from micron-sized grains to 0.05 mm in size. In this meteorite, very minute grains of metallic mineral, less than 0.01 mm in size and showing a pink color with a high reflectivity, were found in the plessitic area of nickel-iron in contact with troilite (Fig. 10). In the atmosphere, this mineral tarnishes into brown. The X-ray microanalysis showed that it is essentially composed of Cu.

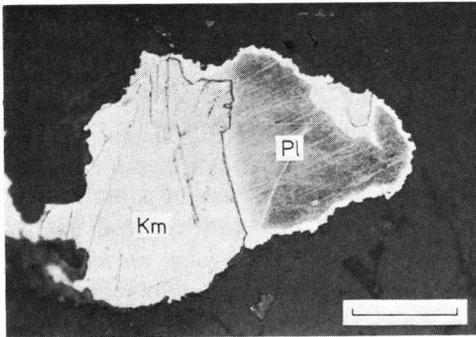


Fig. 9. A nickel-iron grain etched by 5% nital solution. The fine plessite area (Pl) is rimmed with taenite, and the kamacite (Km) shows Neumann lines. Scale bar is 0.05 mm.

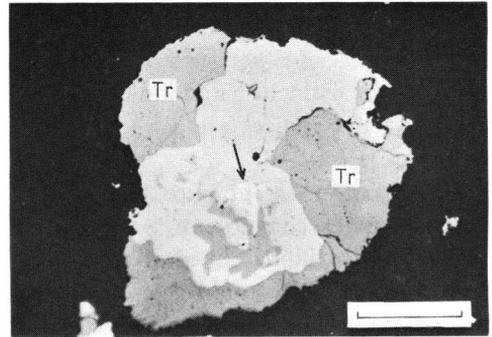


Fig. 10. Native copper (\downarrow) in the nickel-iron associated with troilite grains (Tr). A red filter was used for taking a photograph. Scale bar is 0.05 mm.

IV. Discussion

The present work on the petrography and mineralogy of the Fukutomi meteorite supports the earlier result that this meteorite belongs to the olivine-hypersthene chondrite. The most conspicuous feature of the Fukutomi meteorite is the occurrence of lithic fragments.

Native Cu is generally an accessory species in the meteorite constituents. Among the 280 stones examined, the 170 contain native Cu (Buseck *et al.*, 1973;

Ramdohr, 1973; Okada, 1975; Yabuki, 1978). This corresponds to ~60%. The rate of finding native Cu in H4~6 is more than 70%. The L-group chondrites are a little less, but still the rate account more than 60% for sub-group 3~5. It is interesting that the H-group chondrites, higher in the number for subgroup has higher percentage of these rate whereas for the L-group chondrite, the rate remains similar between 3 to 5 and for H3 and L6 it is the level in 50%. For carbonaceous and enstatite chondrites, the number of stone examined is small, but none has been found in E4, 5 and C2 groups and only one case out of three has been in E6 group. The C3 group has the similar rate, a little over 60%, to ordinary chondrites. For achondrites, only 2 cases out of 10 pyroxene-, one case out of 6 enstatite- and none out of 3 hypersthene-achondrites have reported. The formational process of native Cu is now under question. One possible mechanism is a precipitation of Cu from the Fe-Ni-Cu solid solution, and the other is a decomposition of Cu-Fe sulfide mineral (Ramdohr, 1973). In the Fe-Cu equilibrium system at 1 atm, the solubility of Cu in α -Fe is 330 ppm at 500°C and 3ppm at 300°C, and that in γ -Fe is 2430ppm at 500°C and 170ppm at 300°C (Speich *et al.*, 1966). In the Ni-Fe grain including Cu metal in Fig. 10, the modal composition of native Cu in the metal phase was estimated to be ~2.7%. Assuming the Fe-Cu solid solution containing the same amount of Cu as the above, the precipitation of Cu metal is expected during the course of the slow cooling. Addition of Ni to the Fe-Cu system is likely to change the solubility of Cu, but it is unfortunate that there are no available data on solubility under various conditions. According to the chemical analyses of four ordinary chondrites by Hey *et al.* (1966), kamacite and taenite of olivine-hypersthene chondrites contain 23–65 ppm Cu and 2360–2610 ppm Cu, respectively. This result indicates that much Cu concentrates in the taenite, and that the small amount of Cu contained usually in ordinary chondrites, 50–205 ppm, may entirely dissolve in the taenite phase. In the Fukutomi meteorite, the occurrence of native Cu is inhomogeneous, and Ni-Fe grains including native Cu are very small in number. These facts suggest that metallic Cu was produced by some specific reaction rather than the precipitation from the Fe-Ni-Cu equilibrium system. On the other hand, Ramdohr (1973) stated that native Cu is related to the decomposition of chalcopyrrhotite. Unfortunately, the nature of this mineral has not yet been fully understood by mineralogists. However, it is more probable that the native Cu in the Fukutomi meteorite was produced from a certain copper-bearing sulfide mineral, considering the inhomogeneous distribution of the metallic Cu and its coexistence with troilite. This could be supported with the observation of high contents of sulfide in Fukutomi by Miyashiro *et al.* (1966).

The occurrence of tridymite is interesting, because this mineral is a rare constituent in ordinary chondrites, and because it composes a lithic fragment in this meteorite. Although the presence of free silica is expected from the CIPW norm calculation based on bulk chemical composition, the coexistence of olivines (Fa 23–26) and tridymite is not presumed from the MgO-FeO-SiO₂ equilibrium system. It could be assumed that tridymite was introduced by the mixing process at some stage in the

history of this meteorite.

Of two kinds of lithic fragments, the texture, mineralogy and the coexistence of ovoidal troilite suggest that the dark brown fragment consisting of olivine microporphyry is a rapidly crystallized material of a mafic liquid produced by the shock melting of chondritic material. Its dark color results from the occurrence of finely dispersed opaque grains and the microcrystalline, devitrified material, dark brown in color. Similar fragments are described in other xenolithic chondrites, *i.e.*, Ghubara (Binns, 1968) and St. Mesmin (Dodd, 1974). In this fragment, the outline adjacent to the matrix of the host chondrite is well-defined, and further recrystallization of the dark matrix between the olivine phenocrysts are not noticed under the microscope. These observations imply that this fragment has been subjected to little effect by heating since its redeposition on the parent body.

On the other hand, the origin of the tridymite-bearing fragment is problematic. In stony meteorites, tridymite is often found in the pyroxene-plagioclase achondrites (*e.g.*, Hess *et al.*, 1949; Duke *et al.*, 1965) and in the enstatite chondrites (Binns, 1967a; Mason, 1966), but it is rare in the ordinary chondrites. The existence of cluster-like aggregate of tridymite rimmed with clinopyroxene can not be compared with the occurrence of tridymite in other stony meteorites. A rare sample including free silica xenoliths is the Farmington meteorite, in which cristobalite xenoliths rimmed with diopsidic clinopyroxene are present (Binns, 1967b). The occurrence of free silica xenoliths suggests that there was a silica-rich body on the surface of the parent body during the formation of this meteorite (Binns, 1967b). Another hypothesis is that free silica was supplied from other silica-bearing meteorites, because the recent studies of the xenolithic chondrites revealed the occurrence of various kinds of meteoritic fragments derived from achondrites, enstatite-, ordinary- and carbonaceous chondrites (Fodor *et al.*, 1976; Van Schmus, 1967; Binns, 1968; Kurat *et al.*, 1969; Kallemejn *et al.*, 1978). It is believed that these fragments were produced by brecciation and melting induced by the projectile impacts on the surface of the parent body, and that they were enclosed in this stone by remixing. However, it is a problem whether the mixing of the dark brown fragment and tridymite-bearing one happened at the same time. Clinopyroxene crystals which surround the tridymite fragment were probably formed by the reaction with olivines in the host material at high temperature, while the dark brown fragment does not show any recrystallized feature produced by thermal effect. As a result, it is suggested that the tridymite-bearing fragments were enclosed when the environment was rather hot, and that the dark brown fragment was formed and enclosed at the later and more cooled stage. It is believed that the brecciation and remixing by repeated impactings on the parent body occurred during the long formational history of the Fukutomi meteorite. The deformation structures of minerals, *e.g.*, distorted Neumann bands in kamacite grains and kink bands and mosaicism in olivines, support also the evidence of repeated deformational events.

V. Conclusion

The chemical composition was redetermined by a more refined method than that of former authors. The data indicate that even though the chondrite Fukutomi has large crystals and heterogeneous and peculiar structures, good sampling gives reliable mean chemical compositions. The ratios Fe_{total}/SiO_2 , (0.577), Fe_{metal}/Fe_{total} (0.373), SiO_2/MgO (1.62) and fayalite (23–26) indicate that the chondrite Fukutomi belongs to L-group chondrite.

Further petrographical and mineralogical investigation of the Fukutomi meteorite showed that two kinds of lithic fragments are included. One is a dark brown fragment chiefly consisting of euhedral olivine and dark, fine-grained matrix. This fragment could be produced by the rapid crystallization of shock-melted chondritic material by impact. The other fragment consists of tridymite, and its outer part is rimmed with clinopyroxenes. The petrography of the Fukutomi meteorite indicates that the meteorite is a xenolith-bearing olivine-hypersthene chondrite, and that the brecciation and remixing by repeated impacts on the surface of the parent body occurred during the formational history of this meteorite.

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References

- BINNS, R. A., 1967a. Olivine in enstatite chondrites. *Am. Mineral.* **52**: 1549–1554.
- BINNS, R. A., 1967b. Farmington meteorite: Cristobalite xenoliths and blackening. *Science* **156**: 1222–1226.
- BINNS, R. A., 1968. Cognate xenoliths in chondritic meteorites: Examples in Mezö-Madaras and Ghubara. *Geochim. Cosmochim. Acta* **32**: 299–317.
- BUSECK, R. B., E. HOLDSWORTH, and G. R. SCOTT, 1973. The question of Eaton: Terrestrial versus meteoritic copper. *Geochim. Cosmochim. Acta* **37**: 1249–1254.
- DODD, R. T., 1974. Petrology of the St. Mesmin chondrite. *Contrib. Mineral. Petrol.* **46**: 129–145.
- DUKE, M. B. and L. T. SILVER, 1967. Petrology of eucrites, howardites and mesosiderites. *Geochim. Cosmochim. Acta* **31**: 1637–1665.
- FODOR, R. V. and K. KEIL, 1976. Carbonaceous and non-carbonaceous lithic fragments in the Plainview, Texas, chondrite: Origin and history. *Geochim. Cosmochim. Acta* **40**: 177–189.
- HESS, H. H. and E. P. HENDERSON, 1949. The Moore County meteorite: a further study with comment on its primordial environment. *Am. Mineral.* **34**: 494–507.
- HEY, M. H. and A. J. EASTON, 1968. Copper in various phases of several olivine-hypersthene and olivine-bronzite chondrites. *Mineral. Mag.* **36**: 855–858.
- KALLEMEYN, G. W., W. V. BOYNTON, J. WILLIS and J. T. WASSON, 1978. Formation of the Bencub-

- bin polymict meteorite breccia. *Geochim. Cosmochim. Acta* **42**: 507–515.
- KEIL, K. and K. FREDRIKSSON, 1964. The iron magnesium, and calcium distribution in coexisting olivines and rhombic pyroxenes of chondrites. *J. Geophys. Res.* **69**: 3487–3515.
- KUNO, H., 1954. Study of orthopyroxenes from volcanic rocks. *Am. Mineral.* **39**: 30–46.
- KURAT, G., K. FREDRIKSSON and J. NELEN, 1969. Der Meteorit von Siena. *Geochim. Cosmochim. Acta*, **33**: 763–773.
- MASON, B., 1966. The enstatite chondrites. *Geochim. Cosmochim. Acta.* **30**: 23–39.
- MIYASHIRO, A., S. MURAYAMA and H. HARAMURA, 1966. The Fukutomi, Japan, chondrite. *Japan. J. Geol. Geogr.* **37**: 39–44.
- MURAYAMA, S., Masako SHIMA and A. OKADA, 1978. The chemical composition, petrography and mineralogy of the Japanese chondrite Nagai. *Bull. Natl. Sci. Mus. Ser. E.* **1**: 19–29.
- OKADA, A., 1975. Petrological studies of the Yamato meteorites. Part 1. Mineralogy of the Yamato meteorites. *Mem. Natl. Inst. Polar Res. Spec. Issue. No. 5*, 14–66.
- OKADA, A., Masako SHIMA and S. MURAYAMA, 1979. Mineralogy, petrography and chemistry of the chondrite, Kamiomi, Sashima-gun, Ibaraki-ken Japan. *Meteoritics* **14**: 177–191.
- RAMDOHR, P., 1973. "The opaque minerals in stony meteorites." Akademie-Verlag, Berlin. pp. 25–26.
- SHIMA, Masako, 1974. The chemical compositions of the stone meteorites Yamato (a), (b), (c), and (d), and Numakai. *Meteoritics* **9**: 123–135.
- SHIMA, Masako and M. HONDA, 1967. Distributions of alkali, alkaline earth and rare earth elements in component minerals of chondrites. *Geochim. Cosmochim. Acta* **31**: 1995–2006.
- SHIMA Masako and M. HONDA, 1974. The distribution of lithophile elements in separated phases of enstatite chondrites. *Meteoritics* **9**: 405–408.
- SHIMA, Masako S. MURAYAMA and A. OKADA, 1979. Chemical composition, petrography and mineralogy of the Shibayama chondrite found in Shibayama-machi, Sanbu-gun, Chiba-ken, Japan. *Meteoritics* **14**: 317–330.
- SPEICH, G. R., J. A. GULA and R. M. FISHER, 1966. Diffusivity and solubility limit of copper in alpha and gamma iron. In *Electron Microprobe*. pp. 525–542. Wiley, New York.
- VAN SCHMUS, W. R., 1967. Polymict structure of the Mezö-Madaras chondrite. *Geochim. Cosmochim. Acta* **31**: 2027–2042.
- VAN SCHMUS, W. R. and J. A. WOOD, 1967. A chemical-petrologic classification for the chondritic meteorites. *Geochim. Cosmochim. Acta.* **31**: 747–765.
- YABUKI, H., 1978. Petrological studies on Yamato-74 meteorites. *Sci. Papers Inst. Phys. Chem. Res.* **72**: 14–40.