PETROLOGY AND MINERALOGY OF THE COOLIDGE METEORITE (CV4)

Takaaki NOGUCHI

Department of Earth Sciences, Ibaraki University,
1-1, Bunkyo 2-chome, Mito 310

Abstract: The Coolidge meteorite is important because it is the only equilibrated carbonaceous chondrite which does not belong to the CK group. The petrology and mineralogy of the Coolidge meteorite were investigated by SEM and EPMA. Coolidge has some similarities to the "reduced group" of CV3 chondrites. First, abundant Fe-Ni metal and sulfide are kamacite, taenite, and troilite. Second, AOIs, CAIs, and chondrules are depleted in Na and K. Third, the main constituent the mineral of matrix is olivine. In addition to these similarities, Coolidge also has some unique properties. First of all, matrix as well as AOIs, CAIs, and chondrules is depleted in Na. This corresponds to the depletions of Na abundances in bulk compositions of Coolidge relative to CV3 chondrites. The depletion was probably established before agglomeration of this meteorite. Second, there are some plagioclase pyroxene chondrules in this meteorite. Different from type IV chondrules, they include various amounts of silica mineral. Third, bulk compositions of CAIs are rich in FeO and Cr_2O_3. Fourth, compositional variations of spinel in Coolidge are very wide. The $Cr/(Cr+Al)$ ratio varies from 0.01 to 0.7, and the $Mg/(Mg+Fe)$ ratio varies from 0.45 to 0.77. Fifth, correlations between some elements in spinel and those in pyroxenes in Coolidge are not similar to the "reduced group" of CV3 chondrites but similar to EOCs. The last three differences were probably established during thermal metamorphism. Olivine-spinel geothermometry suggests that Coolidge was heated to temperatures as high as $\sim900°C$. The duration of thermal metamorphism was short enough to preserve many petrographic features such as Fe-Mg compositional zoning in low-Ca pyroxene. Distribution of Co between kamacite and taenite suggest fast cooling around $\sim500°C$.

1. Introduction

The Coolidge meteorite was once classified as H4 (MASON, 1963). Later, it was reclassified as C4, and it has been known as one of the equilibrated carbonaceous chondrites (*e.g.* VAN SCHMUS and WOOD, 1967; VAN SCHMUS, 1969; MCSWEEN, 1977; SCOTT and TAYLOR, 1985). Recent chemical, mineralogical, and petrologic studies of the equilibrated carbonaceous chondrites have shown that most of them belong to the CK group. Although CK chondrites include abundant magnetite instead of Fe-Ni metals, the Coolidge meteorite includes Fe-Ni metals. KALLEMEYN (1987) tentatively classified it as CV4. Therefore, it is important to investigate Coolidge to estimate the degree of metamorphism of the CV chondrite parent body. I studied this meteorite from the standpoint of petrology and mineralogy using a

scanning electron microscope (SEM) and an electron microprobe analyzer (EPMA).

2. Materials and Experimental Methods

A polished thin section (PTS) of the Coolidge meteorite (sample No. 4202-4) was investigated using a SEM and an EPMA. A detailed SEM observation was carried out with a JEOL JSM-840 SEM equipped with an energy dispersive spectrometer (Link system). EPMA analyses were performed by a JEOL JXA-733 microprobe operated at 15 kV accelerating voltage and 9 nA beam current for silicates, oxides, phosphates, and carbonates; and 15 kV and 20 nA for metals and sulfide. Bulk compositions of CAIs, AOIs, chondrules and matrix were analyzed by defocused beams ($\sim 50\,\mu$m in diameters for inclusions and chondrules, and $\sim 20\,\mu$m for matrix) at 15 kV and 9 nA. Correction by the BENCE and ALBEE method (BENCE and ALBEE, 1968) was used for the analyses of silicates, oxides, phosphates, carbonates, and bulk compositions of the inclusions, chondrules and matrix, and the ZAF method for Fe-Ni metal and sulfide. Special deconvolution programs were applied to correct for X-ray overlaps of K_β and K_α lines between some elements such as Ti and V, Ca and P, S and Co, and Co and Fe.

3. Petrography

Coolidge is composed of chondrules, amoeboid olivine inclusions (AOIs), calcium aluminum-rich inclusions (CAIs), metal and sulfide grains, isolated silicate minerals, and fine-grained matrix. The petrography of these components is described in the following sections.

3.1. Chondrules

The most abundant type of chondrule observed in this meteorite is porphyritic olivine pyroxene (POP) chondrule (Fig. 1a). Some barred olivine (BO) and radial pyroxene (RP) chondrules are also observed. Diameters of the chondrules range from 0.2 to 1.8 mm. The average diameter of 86 chondrules is 0.7 ± 0.3 (1σ) mm. This is smaller than the average chondrule diameter of CV3 chondrites, 1 mm (GROSSMAN et al., 1988). Many chondrules contain acicular to prismatic plagioclase crystals (Fig. 1a). The sizes of such plagioclase often reach $10-20\,\mu$m $\times$ $50-150$ μm. CV3 chondrites also contain chondrules which include coarse plagioclase crystals. The morphology and texture of plagioclase are very similar to those in Coolidge. However, the plagioclase-bearing chondrules are more abundant in Coolidge than in CV3 chondrites.

Together with chondrules which contain coarse plagioclase crystals, chondrules containing weakly devitrified brown glass (microcrystalline recrystallized) were also observed. SEM observation shows that there are acicular crystals (sub-μm thick), probably Ca-rich pyroxene, in the chondrule glass.

Most of the low-Ca pyroxenes in chondrules display polysynthetic twinning.
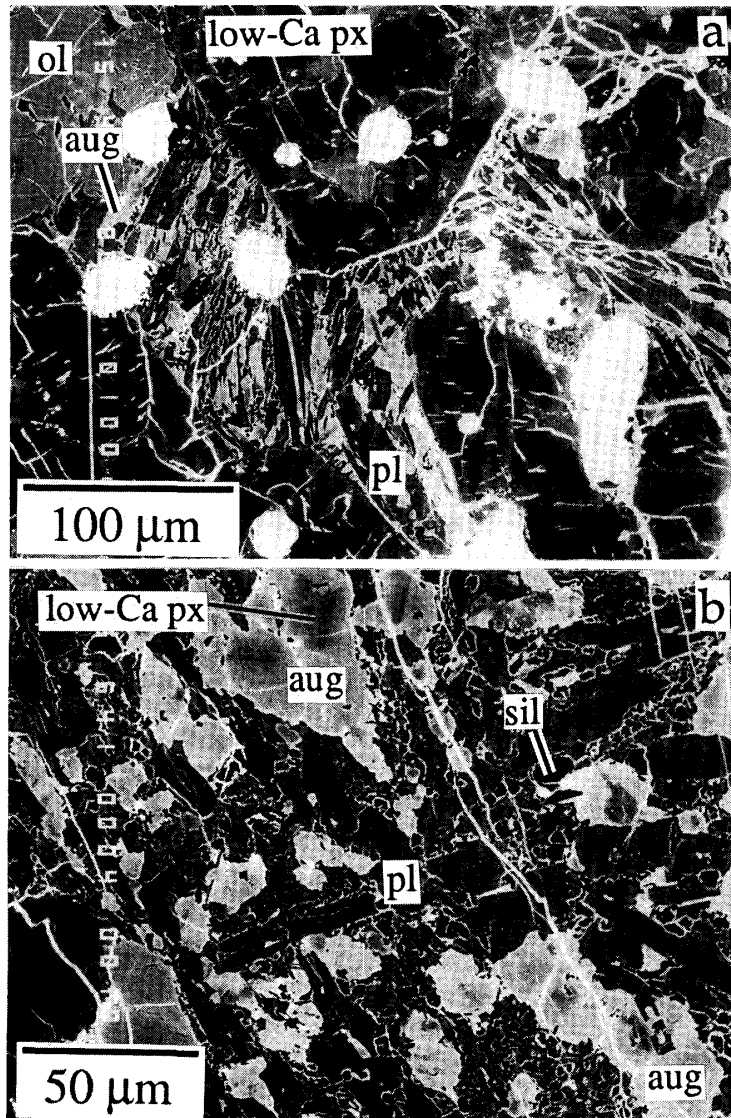
Fig. 1. Backscattered electron image (BEI) photographs of chondrules in Coolidge.

(a) Low-Ca clinopyroxene phenocrysts in the POP chondrule show characteristic lamellae-like Fe-Mg zoning. Interstices between olivine and low-Ca clinopyroxene phenocrysts are filled by laths of plagioclase and Ca-rich pyroxene. Very bright spherules are Fe-Ni metal and troilite. ol: olivine; low-Ca px: low-Ca clinopyroxene; aug: augite; pl: plagioclase.

(b) A plagioclase pyroxene chondrule with many small silica grains. They are enclosed by thin (sub-μm) bright material, probably Ca-rich pyroxene. low-Ca px: low-Ca clinopyroxene; aug: augite; pl: plagioclase; sil: silica mineral.

SEM observation reveals that these pyroxenes have the characteristic Fe-Mg zoning (Fig. 1a), which is shown in some of the type 3–4 ordinary chondrites, Carlisle Lakes-type chondrites, and the Karoonda CK4 chondrite (TSUCHIYAMA et al., 1988; WEISBERG et al., 1991; NOGUCHI, 1993).

Some chondrules are composed mainly of such plagioclases, pyroxenes, and various amounts of silica minerals (Fig. 1b). They sometimes contain small amounts of olivine in addition to these minerals. McSWEEN (1977) and DODD (1981) displayed photomicrographs of Ca and Al-rich chondrules in this meteorite. They are type IV chondrules (McSWEEN, 1977). Type IV chondrules consist of plagioclase, melilite, and spinel, with local and minor perovskite. They have similar bulk composition to that of CAIs, and are undersaturated with SiO_2. BISCHOFF and KEIL (1983a, b, c, 1984) reported Ca and Al-rich chondrules in ordinary chondrites. Such Ca and Al-rich chondrules are composed of Al-Ti-rich pyroxene (fassaite) and olivine with Ca and Al-rich microcrystalline to glassy matrix. As described above,
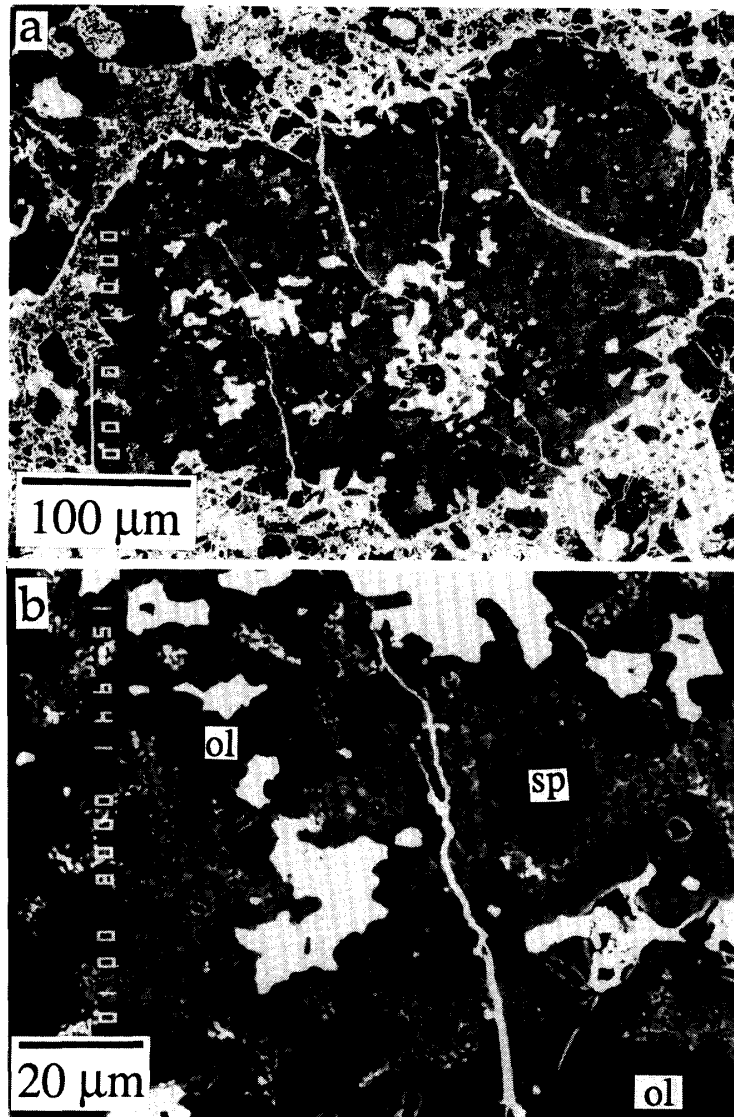
Fig. 2. BEI photographs of an amoeboid olivine inclusion (AOI) in Coolidge.

(a) This inclusion has the rim of olivine. It resembles rimmed olivine aggregates in Allende and Mokoia (KORNACKI et al., 1983).

(b) An enlarged view of the AOI in Fig. 2a. Fine grained materials fill the interstices of olivine and spinel. Very bright patches are troilite. ol: olivine; sp: spinel.

the mineralogy of the plagioclase pyroxene chondrules in this study is different from that of type IV chondrules and Ca and Al-rich chondrules, and suggests that the plagioclase pyroxene chondrules are saturated with SiO_2.

3.2 Amoeboid olivine inclusions

AOIs are relatively small ($<$a few hundreds μm across) as shown in Fig. 2a. Although many small fragments of AOIs ($\sim$several tens of μm across) are observed in the matrix, there are several AOIs which were not fragmented. They are mainly composed of olivine and small amounts of Fe-Ni metal and troilite and aluminous spinel. They have rims of olivine. There are fine-grained ($<$a few μm in diameters) porous materials among the olivine grains (Fig. 2b). Their textures are similar to rimmed olivine aggregates (KORNACKI et al., 1983). Because it is difficult to identify each fine-grained mineral by a microprobe, mineral species of the fine-grained materials are estimated based on many single spot analyses of the fine-
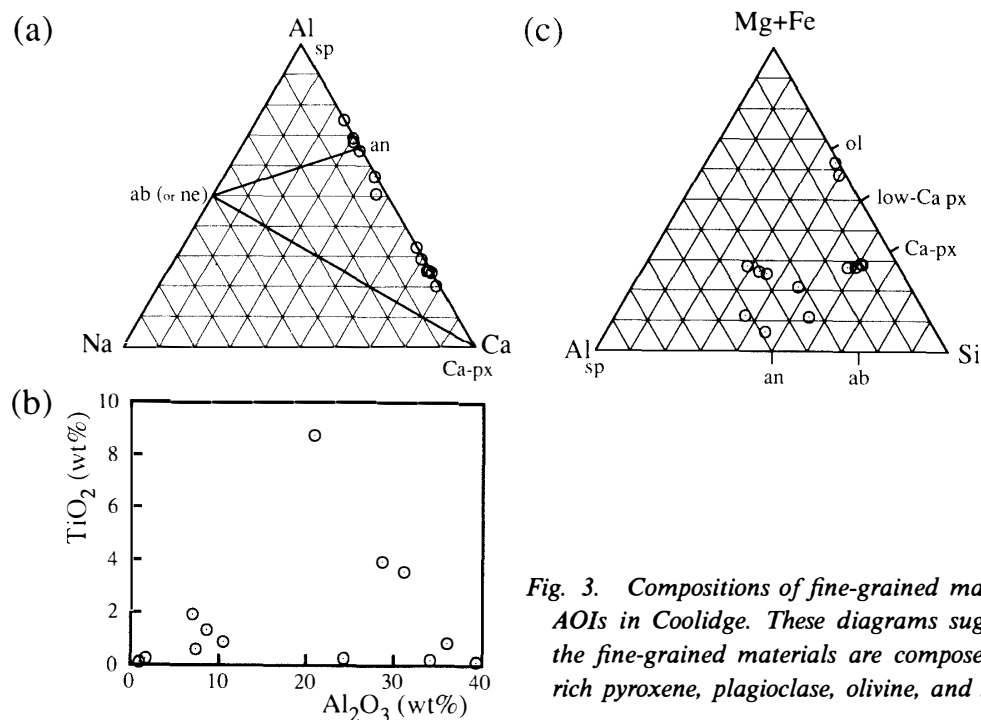
Fig. 3. Compositions of fine-grained materials in AOIs in Coolidge. These diagrams suggest that the fine-grained materials are composed of Ca-rich pyroxene, plagioclase, olivine, and spinel.

grained materials. Figure 3 shows the single spot analyses of the AOIs investigated. This figure suggests that the materials are composed of a mixture of Ca-rich pyroxene, plagioclase, olivine, and spinel. Although nepheline is included in the rimmed olivine aggregates in Allende and Mokoia (KORNACKI et al., 1983), EPMA analysis does not suggest the existence of Na-bearing phases (Fig. 3c). Spinels in the fine-grained materials are probably aluminous, because Cr_2O_3 concentration of the fine-grained materials is < 2 wt%. There are some analyses with high TiO_2 contents (3–9 wt%) in Fig. 3. Their compositions suggest that they are mixtures of spinel and olivine, and of spinel and Ca-rich pyroxene, respectively. Because Ca-rich pyroxene in AOIs contain < 1 wt% TiO_2 (Fig. 12), TiO_2 contents of the fine-grained Ca-rich pyroxene are thought to be < 1 wt%. Therefore, some analyses with high TiO_2 contents (3–9 wt%) in Fig. 3 result from high TiO_2 spinel and/or high TiO_2 inclusions in the spinels.

3.3. Calcium aluminum rich inclusions

CAIs are various in size, 100–500 μmm across (Figs. 4a–4c). All of them are fine-grained CAIs. Small fragments of CAIs also appear in the matrix. The boundaries of such fragments and matrix are often difficult to delineate. CAIs investigated here contain aggregates of subhedral-anhedral spinel crystals ($<$ a few tens of μm in diameter). It is remarkable that many spinels in CAIs show compositional zoning (Fig. 4a). They have aluminous cores and chromian rims. Plagioclases occur as small (< 10 μmm) crystals. Ca-rich pyroxenes are observed in the rims of CAIs and are rare in the interiors of them. In addition to these minerals, calcite, Ca-phosphate, and fine-grained opaque minerals are often found. Ca-phosphate often occurs on the
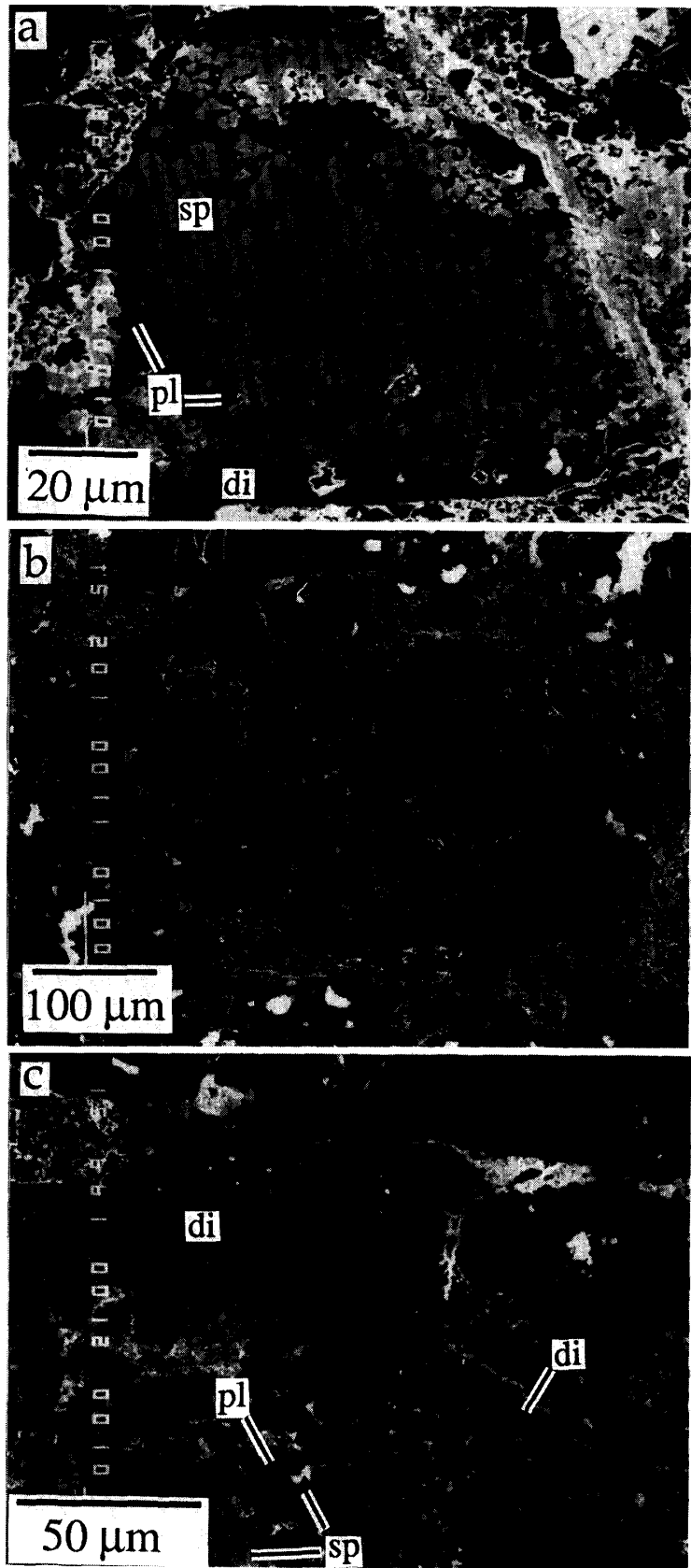
Fig. 4. BEI photographs of cal-
cium aluminum rich inclusions
(CAIs) in Coolidge.

(a) An aggregate of spinel crys-
tals and small plagioclase
crystals also shown in the
interior of the inclusion. The
inclusion is surrounded by a
rim of diopside. This struc-
ture is basically common to
the other CAIs investigated.
It is remarkable that spinels
display compositional zon-
ing. The interstices of spinel
and plagioclase crystals are
filled by fine-grained (sub-
μm) materials. sp: spinel; pl:
plagioclase; di: diopside.

(b) The interior of this CAI is
filled by spinel, plagioclase
and fine-grained materials.
Relatively bright material on
the periphery of the CAI
(near the right edge of this
photograph) is Ca-phosphate.

(c) An enlarged view of the CAI
in Fig. 4b. Diopside also rims
this inclusion. sp: spinel; pl:
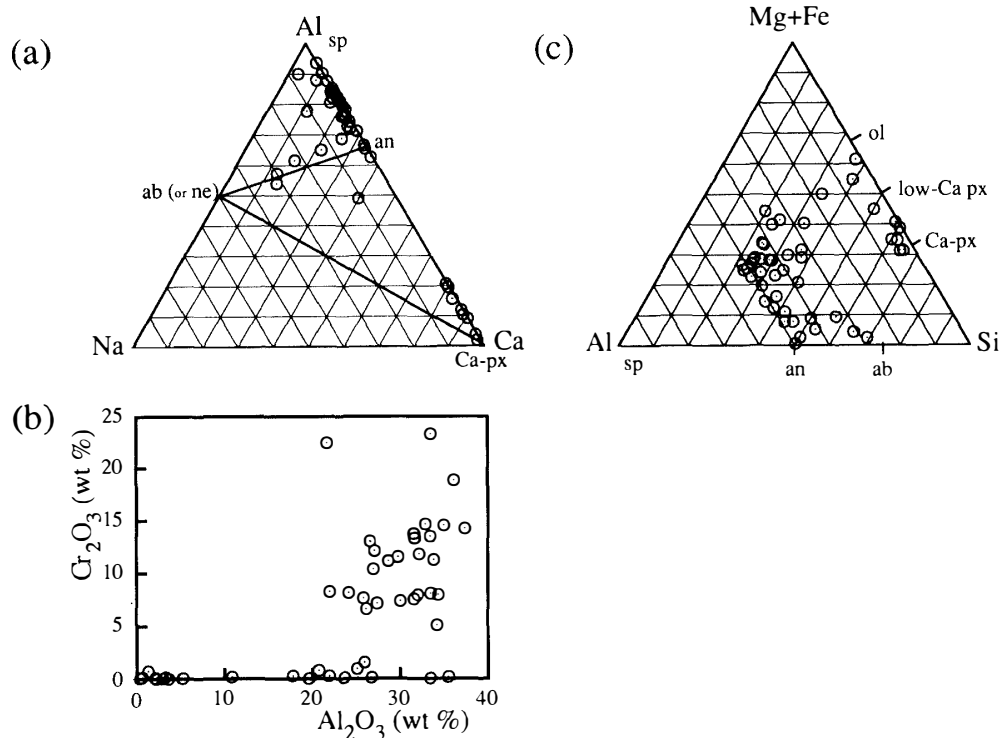plagioclase; di: diopside.

Fig. 5.   Compositions of fine-grained materials in CAIs in Coolidge. These figures suggest
          that the fine-grained materials are composed mainly of spinel, plagioclase, and
          Ca-rich pyroxene.

rims of CAIs.

Interstices of spinel, plagioclase, and Ca-rich pyroxene crystals are filled by
fine-grained (sub-μm size) materials (Figs. 4a and 4c). Because of their small grain
sizes, constituent minerals of the fine-grained material in CAIs are estimated by
considering many single spot analyses of them. Figure 5 shows the analyses and
suggests that they are composed mainly of spinel, plagioclase, and Ca-rich pyroxene.
Figure 5b clearly shows that spinels in the fine-grained materials are rich in Cr.
Their high Cr_2O_3 contents probably resulted from the exchange of Cr and Al
between aluminous spinels and chromian spinels. Occurrences and compositions of
the chromian spinels will be described in Section 4.4. Their small grain sizes
probably facilitated their becoming more chromian than the larger spinel grains,
which have aluminous cores (Figs. 4a and 4c).

3.4.   Matrix

Most portions of the matrix appear opaque under an optical microscope. The
matrix of this meteorite is composed of fine-grained (sub-μm) materials and
fragmental olivine and pyroxene ($>$a few μm across) which were probably derived
from chondrules (Fig. 6). Except for the fragmental silicate minerals, most of the
fine-grained mineral grains in matrix are too fine to be identified by SEM and
EPMA. As shown in the previous sections, constituent minerals of the fine-grained

Fig. 6. A BEI photograph of matrix in Coolidge. The matrix is composed of fine-grained (sub-μm) materials with many fragments of olivine and low-Ca pyroxene which were probably derived from chondrules.
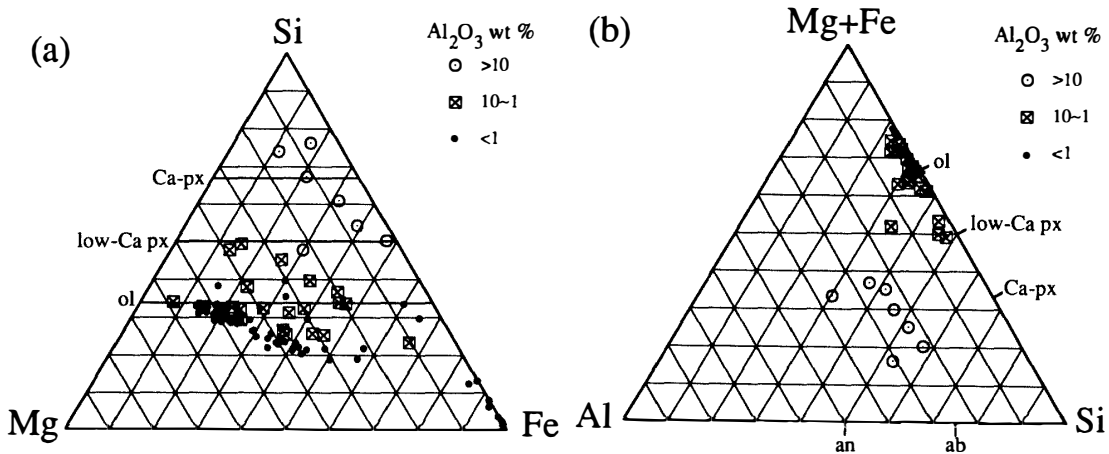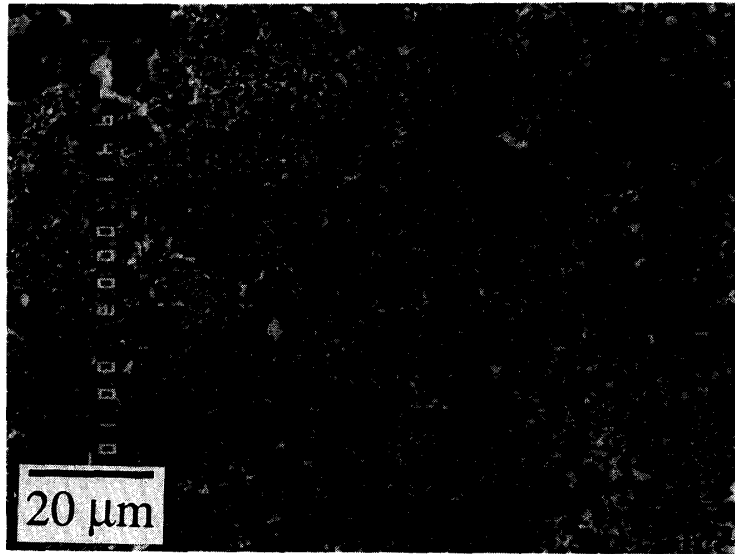

Fig. 7. Compositions of fine-grained matrix in Coolidge. These figures suggest that the matrix is composed of olivine and an Fe-bearing phase (probably Fe-oxide weathering products) with small amounts of pyroxene, and plagioclase.

(sub-μm) matrix are estimated by considering many spot analyses of them. Although SCOTT and TAYLOR (1985) noted that the matrix in Coolidge was highly weathered for EPMA analysis of minerals, the constituent minerals can be estimated by EPMA analysis. The fine-grained matrix is composed of olivine and Fe-bearing materials with smaller amounts of pyroxene and plagioclase (Fig. 7). In addition to these minerals, perhaps small amounts of spinels also exist. Some fractions of spinel and plagioclase were probably derived from fragments of CAIs. Figure 7a suggests that the $Mg/(Mg+Fe)$ ratio of olivine in the matrix varies from 0.7 to 0.8. The Fe-bearing materials are probably Fe-oxide weathering products which were formed from Fe-Ni metals by terrestrial weathering.

3.5. Opaque minerals

Fe-Ni metal and troilite grains occur as spherules in chondrules (Fig. 8). Most
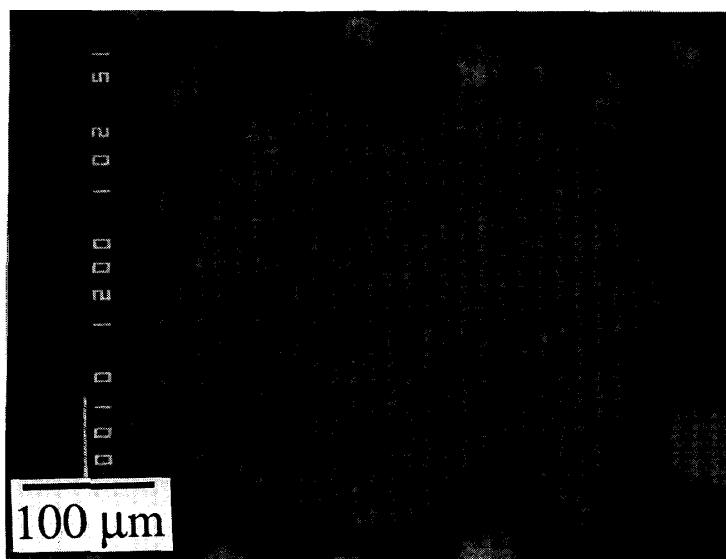
Fig. 8. A BEI photograph of Fe-Ni metal spherules in a chondrule in Coolidge. The spherules are composed of kamacite with small amounts of taenite. Taenite occurs as relatively bright thin (<10 μm) inclusions in kamacite. Slight difference of brightness within kamacite in the largest spherule in this figure suggest that it is a composite grain of kamacite with different crystal orientations.
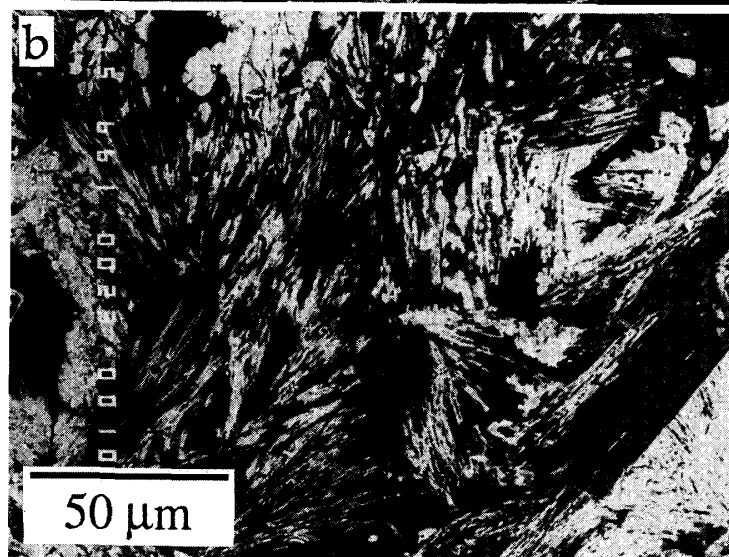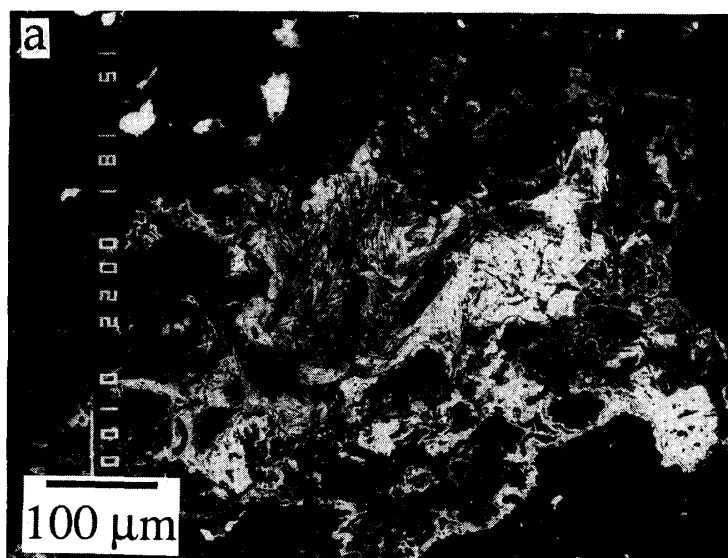

Fig. 9. BEI photographs of an altered metal grain in the matrix in Coolidge.

(a) This inclusion is composed of Fe-oxide weathering products. There is no relic Fe-Ni metal or troilite.

(b) An enlarged view of the inclusion in Fig. 9a. This inclusion is composed of at least two ~three kinds of materials with different brightness on a BEI photograph.

of the metal grains are kamacite with small amounts of taenite. Taenite patches are scattered over kamacite grains. The widths of taenite patches are narrow ($<$ several μm). Fe-bearing sulfide in this meteorite is troilite. Pentlandite is not observed. Chromian spinel and Ca-phosphate sometimes occur as inclusions in the Fe-Ni metal spherules and troilite grains. Peripheries of the spherules are altered to various extents by terrestrial weathering.

On the other hand, Fe-Ni metals in the matrix are highly altered. In the matrix, there are irregularly-shaped inclusions which are composed of aggregates of Fe-bearing materials (Fig. 9). BEI photographs show that there are two or three kinds of materials with different brightness (*e.g.* Fig. 9b). They are composed of acicular materials. Their compositions will be discussed later.

4. Mineralogy

4.1. Olivine

Olivines are homogeneous in this meteorite (Fig. 10 and Table 1). Average Fo content is $86\pm0.6(1\sigma)$, which is consistent with the previous data (85.8 ± 0.4: VAN SCHMUS, 1969; SCOTT and TAYLOR, 1985; 85.6 ± 0.7: GEIGER and BISCHOFF, 1989). Some olivine oikocrysts in low-Ca pyroxene phenocrysts in POP chondrules have higher Fo contents on average, up to 90 mol%. The average CaO content is well below 0.05 wt%, which is consistent with the data reported by SCOTT and TAYLOR (1985). The average MnO content is 0.23 wt%. This value is about half of the average MnO content of olivine in H chondrites in which olivine has similar average Fo content to that in Coolidge. For example, average MnO wt% and Fo mol% of olivine in Allegan H5 chondrite are 0.50 and $Fo_{82.2}$, respectively (NO-GUCHI, unpublished data). This is consistent with lower bulk abundance of Mn in Coolidge than that in the equilibrated H chondrites (KALLEMEYN, 1987; PALME *et al.*, 1988). Olivine in Coolidge does not contain Ni, as described by GEIGER and BISCHOFF (1989).

4.2. Pyroxene

Low-Ca pyroxenes in chondrules are relatively heterogeneous (Fig. 11 and Table 1). Most low-Ca pyroxenes are from $\sim En_{97}$ to $\sim En_{85}$, which is consistent with that by GEIGER and BISCHOFF (1989). They reported that En content of low-Ca pyroxene ranges from En_{95} to En_{82}. Average En content is $88.9\pm3.7(1\sigma)$. VAN SCHMUS (1969) reported that average En mol% is $88.8\pm1.23(1\sigma)$. Because most magnesian and ferroan portions of each low-Ca pyroxene crystal were selected on BEI photographs in this study, the standard deviation in this study is larger than that in VAN SCHMUS (1969). Although the points of analysis were selected on the basis of BEI photographs, low-Ca pyroxenes with $>En_{97}$ and $<En_{80}$ are not observed. The average Wo content of low-Ca pyroxenes is 2.1 mol%. Figure 11 suggests that there are compositional orthopyroxene and pigeonite, in addition to low-Ca clinopyroxene.

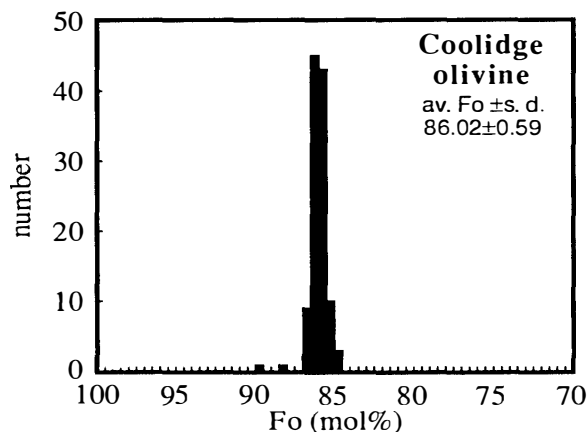Wo mol% of Ca-rich pyroxenes varies. However, Mg values are relatively

Fig. 10. A histogram of Fo mol% of olivine in chondrules in Coolidge.

Table 1. Average compositions of olivine and pyroxene in Coolidge.

| Mineral	Olivine	s.d.	low-Ca pyroxene	s.d.	Ca-pyroxene	s.d.
SiO_2	40.24	0.38	57.28	1.49	52.29	2.70
TiO_2	0.02	0.06	0.12	0.13	0.89	0.60
Al_2O_3	0.03	0.08	1.13	1.96	3.96	3.56
Cr_2O_3	0.02	0.11	0.29	0.22	0.59	0.58
V_2O_3	0.01	0.09	0.00	0.01	0.00	0.01
FeO	13.35	0.54	6.31	2.21	3.05	1.79
NiO	0.03	0.05	0.03	0.04	0.07	0.13
MnO	0.21	0.06	0.16	0.14	0.16	0.20
MgO	46.06	0.64	33.92	1.95	18.71	3.20
CaO	0.02	0.09	0.93	1.09	19.85	3.48
ZnO	0.00	0.01	0.00	0.01	0.00	0.01
Na_2O	0.00	0.01	0.01	0.03	0.23	0.20
K_2O	0.00	0.00	0.00	0.01	0.00	0.01
P_2O_5	0.01	0.06	0.01	0.02	0.02	0.06
Total	100.01		100.17		99.83	

s.d.: standard deviation.
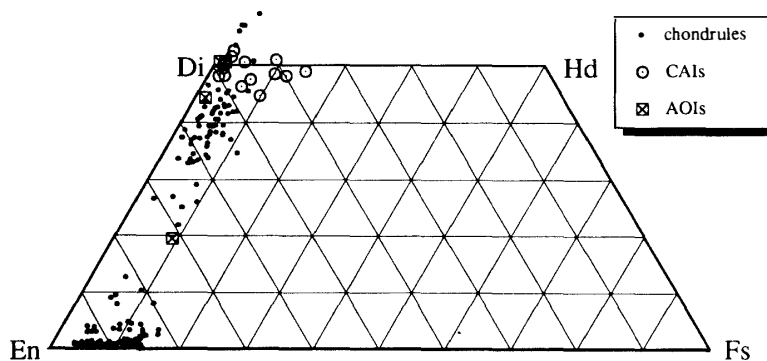
Fig. 11. A pyroxene quadrilateral diagram showing compositions of pyroxene in CAIs, AOIs, and chondrules in Coolidge.

homogeneous (Fig. 11). Ca-rich pyroxenes in chondrules, AOIs, and CAIs have different chemical compositions. Wo and Fs contents of Ca-rich pyroxenes in CAIs tend to be higher than those in chondrules and similar to those in AOIs. En contents of Ca-rich pyroxenes in CAIs are lower than those in chondrules and AOIs.

Average Al_2O_3, TiO_2, MnO, and CaO contents in low-Ca pyroxenes and Ca-rich pyroxenes in chondrules are similar to those in CV3 and CO3 chondrites (*e.g.* NOGUCHI, 1989). Al_2O_3 and TiO_2 contents of most of the Ca-rich pyroxenes in CAIs are <5 and <0.8 wt%, respectively. They are lower than those in CAIs in CV 3 chondrites (*e.g.* TESHIMA and WASSERBURG, 1985; WAEK, 1987). Average Cr_2O_3 content in pyroxenes in chondrules (0.26 wt% for low-Ca pyroxene, and 0.56 wt%
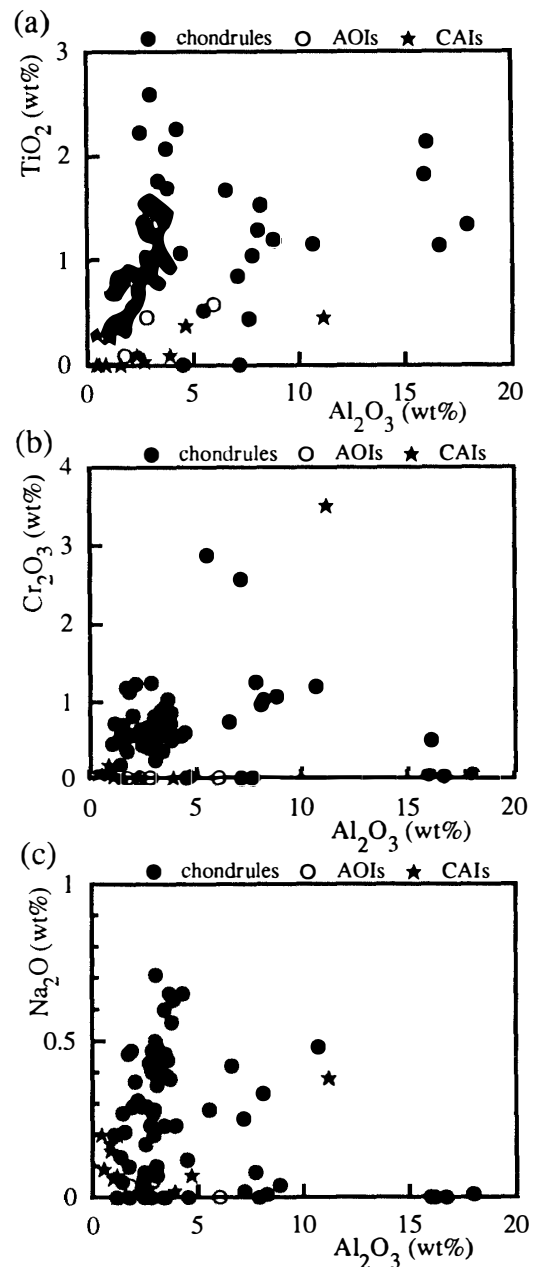
Fig. 12. Relationships between minor elements in Ca-rich pyroxenes in Coolidge.

(a) A plot of Al_2O_3 vs. TiO_2 in Ca-rich pyroxene showing a positive correlation between these values.

(b) A plot of Al_2O_3 vs. Cr_2O_3 in Ca-rich pyroxene.

(c) A plot of Al_2O_3 vs. Na_2O in Ca-rich pyroxene. Most of Ca-rich pyroxene show a positive correlation between them. However, some Ca-rich pyroxene crystals with high Al_2O_3 show very low Na_2O content.

Table 2. *Representative compositions of high-Al₂O₃ pyrox-*
enes in Coolidge.

Mineral	low-Ca pyroxene	Ca-pyroxene
SiO_2	47.88	43.68
TiO_2	0.32	1.33
Al_2O_3	16.80	17.99
Cr_2O_3	0.00	0.07
V_2O_3	0.00	0.00
FeO	3.11	0.53
NiO	0.05	0.07
MnO	0.02	0.03
MgO	30.42	12.98
CaO	1.35	23.51
ZnO	0.00	0.00
Na_2O	0.00	0.01
K_2O	0.00	0.00
P_2O_5	0.00	0.00
Total	99.96	100.20

for Ca-rich pyroxene) is lower than those in CV3, CO3, and CM2 chondrites (0.6–0.7 and 0.9–1.1, respectively; NOGUCHI, 1989). On the other hand, average Na_2O content of Ca-rich pyroxene (0.21 wt%) is much higher than that in CV3, CO3, and CM2 chondrites (<0.07 wt%; NOGUCHI, 1989). TiO_2 contents of Ca-rich pyroxene in chondrules are positively correlated with Al_2O_3 contents (Fig. 12a). On the other hand, Cr_2O_3 contents are not well correlated with Al_2O_3 contents (Fig. 12b). Figure 12c shows the relationships between Al_2O_3 and Na_2O contents. Many Ca-rich pyroxenes in chondrules show a weak positive correlation between Al_2O_3 and Na_2O contents (Fig. 12c). But there are also some Ca-pyroxenes in chondrules which include almost no Na_2O.

Ca-rich pyroxenes including >15 wt% Al_2O_3 and almost no Na_2O occur in a BO chondrule. They coexist with low-Ca pyroxene including extremely high Al_2O_3 (12.72–16.80 wt%) (Table 2). Such subsilicic aluminian pyroxenes were reported in Y-82038 (H3) by WATANABE and MORIMOTO (1987). In Y-82038, such pyroxenes were discovered from a coarse grained lithic fragment. High Al_2O_3 contents in the low-Ca pyroxenes exceed the range of equilibrium solubility of Al_2O_3 even under high pressures (*e.g.* GASPARICK, 1984). Therefore, such pyroxenes probably crystallized under disequilibrium conditions.

4.3. Plagioclase

Plagioclase is observed in chondrules, CAIs, and AOIs. Plagioclase compositions in chondrules and the inclusions are heterogeneous. An mol% varies from 100 to 27. GEIGER and BISCHOFF (1989) reported a similar range of An mol% of plagioclase (100–30). Compositions of plagioclase in chondrules vary from $\sim An_{95}$ to $\sim An_{45}$ (Fig. 13a). Compositional variation of plagioclase within a chondrules is relatively small (2–6 An mol%), except for plagioclase in a porphyritic plagioclase
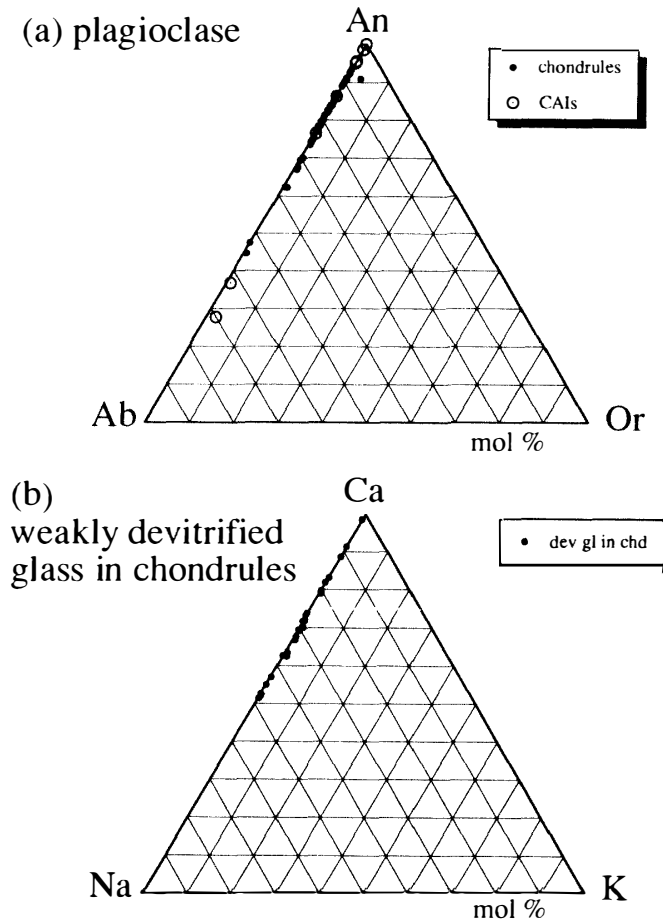
Fig. 13. (a) An-Ab-Or plot showing compositions of plagioclase in CAIs and chondrules in Coolidge. (b) Ca-Na-K plot showing compositions of weakly devitrified glass in chondrules in Coolidge.

pyroxene chondrule, in which plagioclase shows wide compositional variation (An_{93}–An_{68}). Therefore, compositions of plagioclase were probably controlled by the bulk composition of each chondrule. Compositional variations of plagioclase in CAIs are larger than that in chondrules (Fig. 13a). Although most of plagioclases in them have >An_{95}, the largest variation reaches 68 mol% An (An_{95}–An_{27}). Plagioclase in AOIs could not be analyzed by an EPMA due to their small sizes.

4.4. Spinel

Spinel group minerals in Coolidge show large compositional variation. Their $Cr/(Cr+Al)$ ratios vary from 0.01 to 0.7, and $Mg/(Mg+Fe)$ ratios from 0.45 to 0.77 (Fig. 14 and Table 3). Aluminous spinels occur in all the CAIs investigated. Their compositional range of $Cr/(Cr+Al)$ varies from 0.01 to 0.49, and the $Mg/(Mg+Fe)$ ratios from 0.53 to 0.77 (Fig. 14). Their wide compositional variation resulted from compositional zoning as described in Section 3.3. Mg-Al spinel component decreases from their cores to rims, and vice versa for the chromite component.
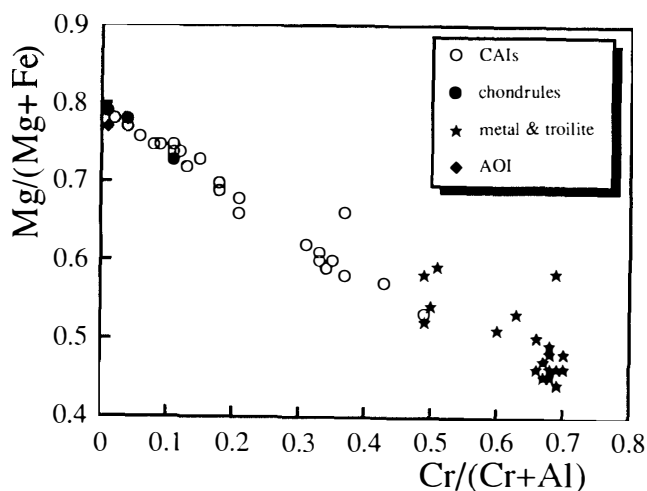
Fig. 14. A plot of $Cr/(Cr + Al)$ vs. $Mg/(Mg + Fe)$ ratios of spinels in various occurrences in Coolidge. There is a definite negative correlation between these values.

Table 3. Representative compositions of spinel in Coolidge.

Occur.	AOI	CAI	CAI	Chondrule	Metal
SiO_2	0.13	0.00	0.00	0.00	0.00
TiO_2	0.06	0.00	0.34	0.01	0.63
Al_2O_3	66.98	67.13	27.80	64.21	15.60
Cr_2O_3	1.04	1.30	40.20	3.74	52.78
V_2O_3	0.09	0.09	0.42	0.15	0.58
FeO	10.59	9.83	18.37	10.28	19.99
NiO	0.05	0.00	0.01	0.01	0.20
MnO	0.10	0.04	0.32	0.12	0.50
MgO	20.31	20.73	11.51	20.19	9.56
CaO	0.18	0.01	0.00	0.00	0.00
ZnO	0.20	0.45	0.19	0.47	0.00
Na_2O	0.00	0.01	0.00	0.00	0.00
K_2O	0.00	0.00	0.00	0.00	0.00
P_2O_5	0.00	0.00	0.00	0.00	0.00
Total	99.73	99.59	99.17	99.19	99.84

Aluminous spinels are also observed in some chondrules and AOIs. In contrast to the case of aluminous spinels in CAIs, their compositional zonings are not remarkable. Although their grain sizes are small ($< 20 \mu m$ across), their $Cr/(Cr + Al)$ ratios of them are low (< 0.1) and their $Mg/(Mg + Fe)$ ratios are high (Fig. 14). Because they are enclosed in olivine crystals, interdiffusion of Al and Cr cations among aluminous and chromites were probably inhibited.

Chromian spinels as well as aluminous ones occur in chondrules. They are associated with Fe-Ni metal and/or troilite grains within chondrules or on their rims. Chromian spinels are observed on the peripheries of Fe-Ni metal or troilite grains. Their grain sizes are small ($< 20 \mu m$ across). Most of them are plotted around $Cr/(Cr + Al) \approx 0.67$ and $Mg/(Mg + Fe) \approx 0.46$ (Fig. 14).
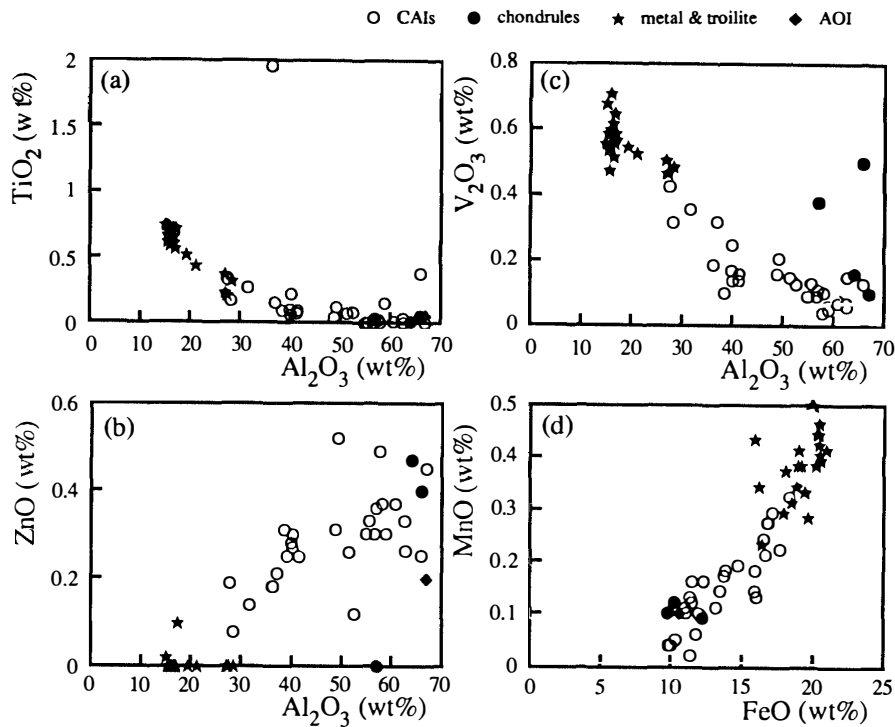
Fig. 15. Relationships between major and minor elements of spinels.

    (a) A plot of Al_2O_3 vs. TiO_2 in spinels. Spinels with $<30 wt\%$ Al_2O_3 show a negative correlation between these values. One spinel grain in a CAI contains high TiO_2 content.

    (b) A plot of Al_2O_3 vs. ZnO in spinels. Although the scattering of data points is relatively large, a positive correlation can seen between these values.

    (c) A plot of Al_2O_3 vs. V_2O_3 in spinels. There is a negative correlation between these values. Some spinels in chondrules do not obey the negative correlation.

    (d) A plot of FeO vs. MnO in spinels. There is a positive correlation between them.

Figure 15 shows the relationships between some major and minor elements in spinels. There are negative correlations of TiO_2 and V_2O_3 with Al_2O_3. On the other hand, there are positive correlations of ZnO with Al_2O_3, and of MnO with FeO. In Fig. 15a, one spinel grain in a CAI shows high TiO_2 content. The high TiO_2 content may result from fine-grained (sub-μm?) inclusion of high-TiO_2 phase (perovskite or ilmenite). In Figs. 15b and c, spinels in chondrules and an AOI tend to deviate from the trends shown in these figures. Olivines including them perhaps inhibited diffusion of V and Zn cations among aluminous and chromian spinels. Chromites in EOCs and CO3s of higher petrologic subtypes (3.3–3.6) contain >0.1 and 0.1–0.3 wt% ZnO, respectively (JOHNSON and PRINZ, 1991). On the other hand, most of the chromian spinels in Coolidge have well below 0.1 wt% ZnO.

4.5. Metal

CV3 chondrites of the reduced group (McSWEEN, 1977) contain abundant kamacite with small amounts of taenite, and sulfide in them is troilite. On the other hand, the CV3 chondrites of the oxidized group (McSWEEN, 1977) contain abun-
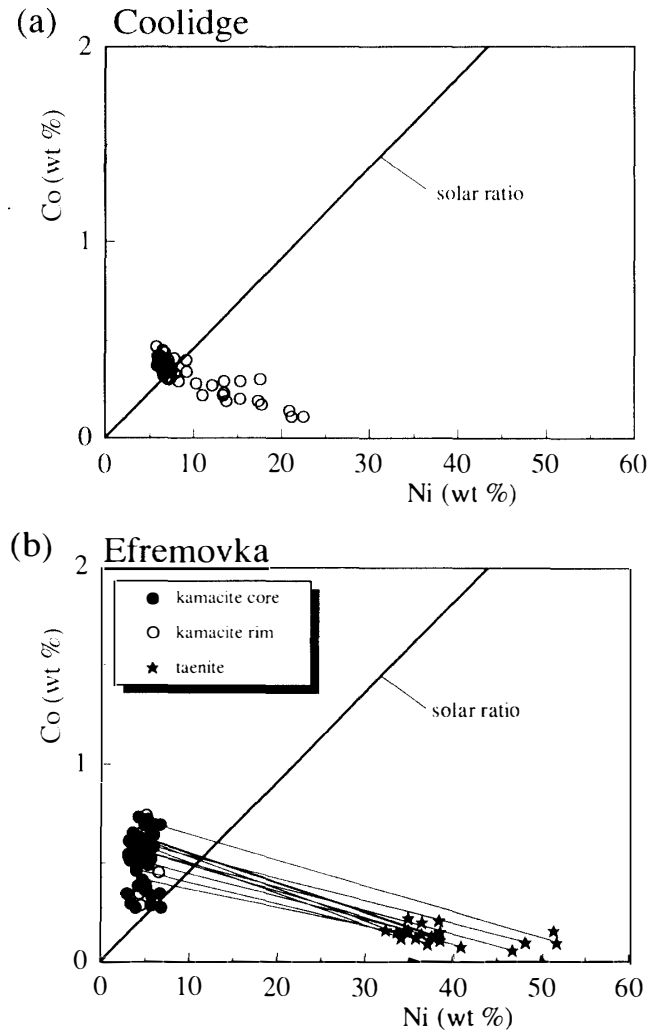
Fig. 16.   *Ni vs Co contents in Fe-Ni metals in Coolidge and Efremovka.*

   (a)   *Fe-Ni metals in Coolidge. Because taenite patches are thin (<5 μm), the distinction between kamacite and taenite is insufficient.*

   (b)   *Fe-Ni metals in Efremovk(a) Because taenite occurs as large grains (<a few tens mm across), the distinction between kamacite and taenite is good. There are tetrataenite or high-Ni taenite in addition to kamacite and taenite. Tie lines indicate coexisting pairs of kamacite and taenite.*

dant Fe-Ni metals with high Ni content and abundant high Ni sulfides. As described in Section 3.5, most of the Fe-Ni metals in Coolidge are kamacite. Sulfide is troilite in Coolidge. Therefore, the Coolidge meteorite resembles the "reduced group" of CV3s on the basis of mineralogy of metal and sulfides.

Ni and Co contents of kamacite in Coolidge are ~7 and ~0.35 wt%, respectively. The compositional ranges of both Ni and Co of kamacite are narrow (~2.5, ~0.2, respectively). Taenite exists as thin exsolution lamellae in kamacite grains (Fig. 8). Due to the small grain size of the taenite exsolution (< 5 μm thick), the distinction between kamacite and taenite compositions is insufficient (Fig. 16a). If the Ni-rich and Co-poor end of the trend on Fig. 16a represents composition of pure

Table 4. Representative compositions of kamacite and tae-
nite in Coolidge.

	Kamacite	Taenite
S	0.00	0.00
Fe	93.01	77.62
Co	0.38	0.14
Ni	6.36	20.85
Total	99.74	98.60

taenite, Ni and Co contents of taenite are $\sim$23 and $\sim$0.1, respectively (Table 4). Figure 16 shows the Ni vs. Co diagrams of metals in Coolidge and Efremovka, one of the "reduced group" of CV3 chondrites for comparison. Kamacites in Efremovka are relatively homogeneous in respect to Ni content (Fig. 16b). Kamacites in Efremovka include relatively coarse (up to $\sim$20 μm across) taenite on their peripheries, as shown in kamacites in the EOCs. Taenite coexisting with kamacite also has relatively homogeneous composition, although some tetrataenites or high Ni taenites are also observed. Figure 16b suggests that kamacite which contained Ni higher than $\sim$8 wt% exsolved taenite. The range of Ni content in kamacite before the exsolution of taenite in Efremovka would be wider than that in Coolidge.

4.6. Others

a) Fe-oxide weathering products

As described in Section 3.5, peripheries of Fe-Ni metal grains are altered to various extents. SEM observation shows that there are at least two kinds of minerals which were probably formed by weathering (Fig. 9). Fe_2O_3 concentrations of them (recalculated from EPMA data) are 55–68 and 78–89 wt%, respectively (Table 5). Table 5 shows compositional ranges of each element in them. Typical limonite contains 70–83 wt% Fe_2O_3 (GOODING, 1981; IKEDA and KOJIMA, 1991). Fe_2O_3 concentrations of goethite (α-FeOOH; the main constituent of limonite) and ferri-hydrite ($5Fe_2O_3 \cdot 9H_2O$: FLEISCHER et al., 1975; $Fe_2O_3 \cdot 2FeOOH \cdot 2.6H_2O$: RUSSELL, 1979) are 89.9 and 83.1, respectively. Therefore, the brighter materials in the BEI photos are probably formed by these phases. On the contrary, Fe_2O_3 concentrations of the darker materials are much lower than these values. Perhaps they contain higher H_2O and CO_2 than the brighter materials.

b) Carbonate and Ca-phosphate

Calcite is observed in some CAIs and in veins. It is not necessarily clear that calcite in CAIs is a weathering product. But calcite in veins is probably a weathering product. The veins are filled by limonite and/or calcite. In two veins, MgO concentration of calcite increases toward their centers (up to $\sim$10 wt%).

Ca-phosphates are observed in CAIs, Fe-Ni metal grains, and chondrules. Their FeO, MgO, and Na_2O concentrations are 0–2.53, 0–3.10, and 0.36–2.49, respectively. These data suggest that there are both whitlockite and apatite in this meteorite.

Table 5. Compositional ranges of Fe-oxide weathering products.

	Bright	Dark
SiO_2	0–2.92	0.02–7.95
TiO_2	n.d.	0–0.06
Al_2O_3	0–0.09	0–1.20
Cr_2O_3	n.d.	n.d.
Fe_2O_3[a]	78.16–89.25	55.50–66.40
NiO	2.99–7.66	1.87–10.52
MnO	n.d.	n.d.
MgO	0–0.15	0–0.26
CaO	0–0.43	0.02–1.04
Na_2O	n.d.	0–0.16
K_2O	n.d.	n.d.
P_2O_5	0.10–0.78	0.07–2.07
SO_3	0.07–1.21	0.07–2.94
Cl	n.d.	n.d.
Total[b]	87.2–91.24	59.55–76.09

[a]: All the iron is assumed to occur as Fe^{3+}.

[b]: Differences from 100 are interpreted as weights of H_2O and CO_2.

5. Bulk Compositions of AOIs, CAIs, Chondrules, and Matrices in Coolidge and Efremovka

Bulk compositions of AOIs, CAIs, chondrules, and matrices of CV chondrites including Coolidge and Efremovka were investigated by McSWEEN (1977). In this paper, three kinds of diagrams are also used in addition to the $FeO–(CaO + Al_2O_3)$–MgO diagrams.

It can be seen in a $FeO–(CaO + Al_2O_3)$–MgO diagram (Fig. 17a) that chondrules and AOIs in Coolidge have equilibrated compositions and CAIs are rich in FeO. This result is consistent with that in McSWEEN (1977). He also noted that higher FeO contents in CAIs might result from unidentified Cr-rich spinels in them. It is confirmed that higher FeO contents and higher $Fe/(Fe + Mg)$ ratios of CAIs (Figs. 17a and b) result from chromites with higher $Fe/(Fe + Mg)$ ratios as described in Sections 3.3 and 4.4.

In a Si–Mg–Fe diagram (Fig. 17b), most of the chondrules in Coolidge are plotted between the olivine and low-Ca pyroxene solid solution lines. They are normal ferromagnesian chondrules. It corresponds that olivine in them has equilibrated compositions. However, some chondrules are plotted between low-Ca pyroxene and Ca-rich pyroxene solid solution lines in the diagram. They correspond to plagioclase pyroxene chondrules which were described in Section 3.1. Figure 17 c shows that such plagioclase pyroxene chondrules contain less Al than CAIs. In Fig. 17a, the points plotted nearest to the $CaO + Al_2O_3$ apex among chondrules correspond to the plagioclase pyroxene chondrules.

In an Al–Na–Ca diagram (Fig. 17d), chondrules are plotted near the point where $Ca/(Ca + Al)$ ratio is 2:1 on the edge between Ca and Al apices. As described

Coolidge
(a)

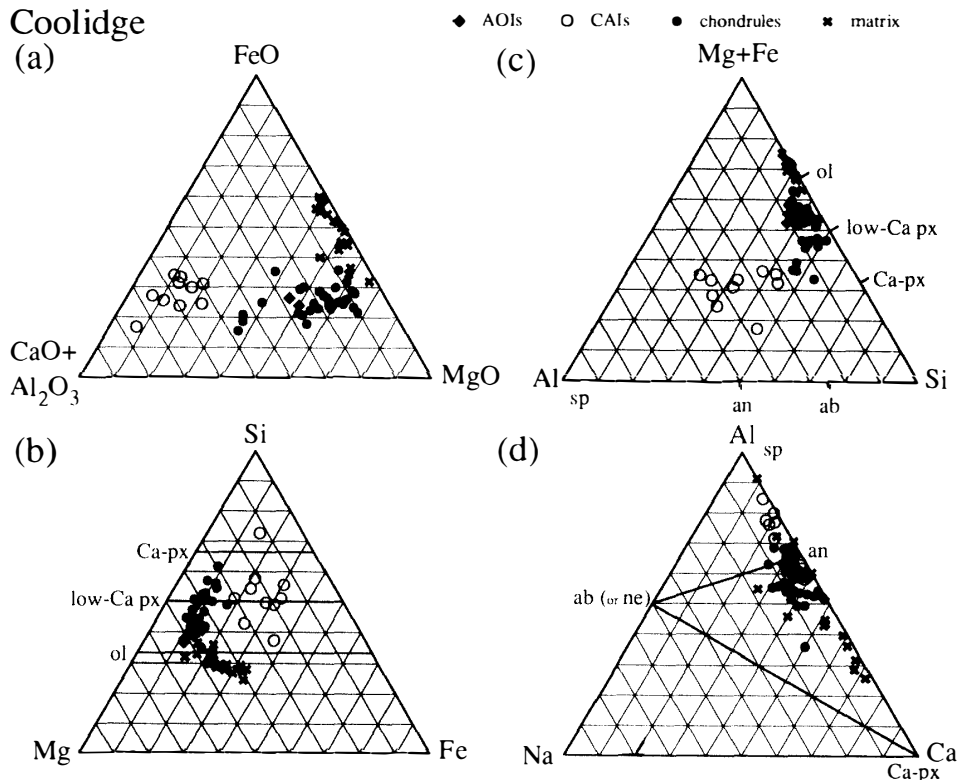◆ AOIs   O CAIs   ● chondrules   ✱ matrix

Fig. 17. Bulk compositions of AOIs, CAIs, chondrules, and matrix in Coolidge.

in Sections 3.1 and 4.3, chondrules contain anorthite-rich plagioclase or glass rich in CaO and Al_2O_3. This fact is consistent with Fig. 17d. It is interesting that Na is highly depleted in not only CAIs and AOIs but also in chondrules and in matrix, in Coolidge (Fig. 17d).

Because Efremovka is a CV3 chondrite, the four constituent components are not equilibrated. CAIs, AOIs, chondrules, and matrix occupy different regions on the $FeO-(CaO+Al_2O_3)-MgO$ diagram (Fig. 18a). This result is consistent with that in McSWEEN (1977). In this figure, two ferroan chondrules (type II chondrules) are plotted among the analyses of the matrix. The similar tendency is shown in the other diagrams (Figs. 18b, c and d). McSWEEN (1977) already noted that the few porphyritic chondrules which had the similar composition to the matrices were type II chondrules. Figure 18 shows that there is a compositional gap between type I and II chondrules in Efremovka.

Most of Fe-Ni metals in type I chondrules include Cr in various amounts (< 1.2 wt%). However, type II chondrules contain euhedral chromite crystals. This difference clearly indicates that type II chondrules were formed under more oxidized conditions than type I chondrules. There are more CAIs in Efremovka which contain higher amounts of Al than that in CAIs in Coolidge (Fig. 18b and c). This result is probably related to the high Cr_2O_3 contents in CAIs in Coolidge.

In an Al-Na-Ca diagram (Fig. 18d), most of the chondrules are plotted near the point where the Ca/(Ca + Al) ratio is 2 : 1 on the edge between Ca and Al apices. This suggests that the chondrules contain anorthite-rich plagioclase or glass rich in
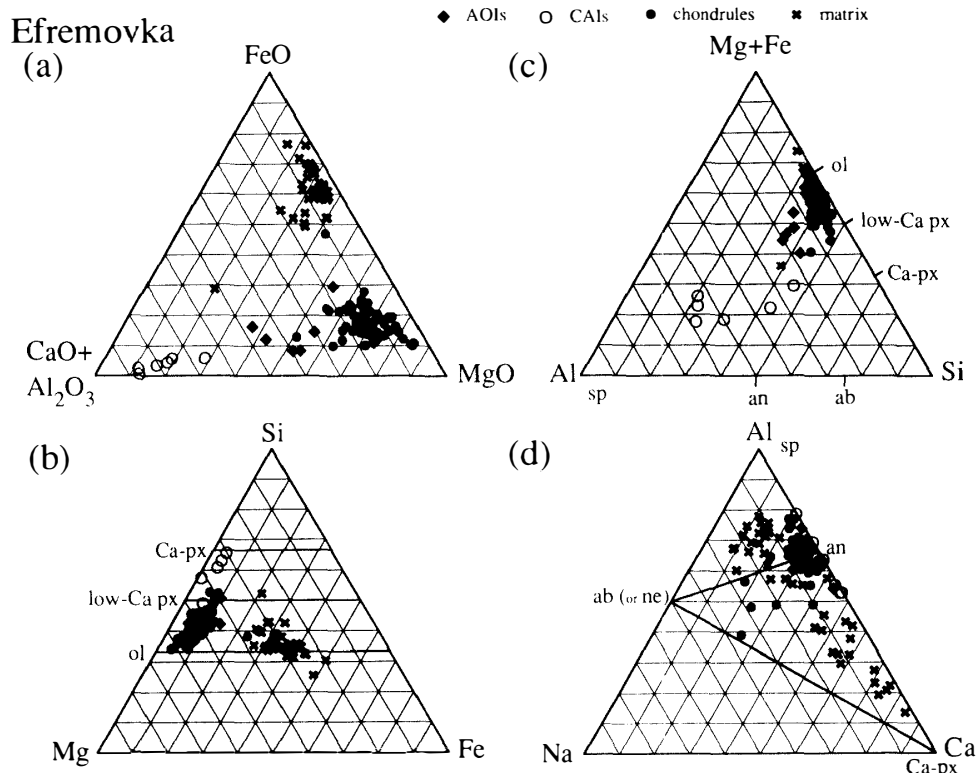
Fig. 18. Bulk compositions of AOIs, CAIs, chondrules, and matrix in Efremovka.

CaO and Al_2O_3. It is interesting that Na is highly depleted in not only CAIs and AOIs but also in chondrules in Efremovka, as in Coolidge (Figs. 17d and 18d). The matrix in Efremovka has low Na content, but Na content in the matrix are higher than that in type I chondrules, AOIs, and CAIs (Fig. 18d). SEM observation shows that chondrules in Efremovka as well as AOIs and CAIs seem not to have experienced pre-accretionary alteration, which is well recognized in Allende (e.g. HASHIMOTO and GROSSMAN, 1987; MCGUIRE and HASHIMOTO, 1989; KIMURA and IKEDA, 1992). Primary compositions of these constituent components are probably well preserved in this meteorite.

6. Discussion

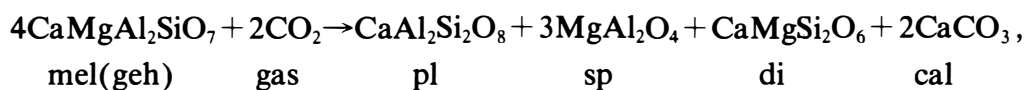### 6.1. Plagioclase and glass in chondrules

The existence of acicular to prismatic plagioclases in chondrules is characteristic for Coolidge. The morphology of these plagioclases indicate that they were crystallized from chondrule melts. Chondrules which contain weakly devitrified brown glass (microcrystalline recrystallized) coexist with the plagioclase-bearing chondrules in this meteorite. The compositional range of glass overlaps with that of plagioclase in chondrules, although there are more chondrule glasses which are more sodic than plagioclase in chondrules (Fig. 13). These data suggest that the different degree of devitrification among chondrules does not necessarily result from the difference in composition of chondrule glass. There is a possibility that the

cooling rates after crystallization of olivine and pyroxene varied among chondrules. Weakly devitrified brown glass in chondrules which is microcrystalline recrystallized were not coasely recrystallized during thermal metamorphism of Coolidge due to the short duration of metamorphism. This interpretation is consistent with the existence of low-Ca pyroxenes with remarkable $Mg/(Mg+Fe)$ zoning.

6.2. Metamorphism of Coolidge: compositional changes of minerals by thermal metamorphism and metamorphic conditions

As described in Sections 3 and 4, Coolidge experienced thermal metamorphism. However, textural and bulk compositional changes vary among the constituent components. Textures and bulk compositions (except for compositions of olivine) of chondrules and matrix show that they were not affected severely by thermal metamorphism, because glass in some chondrules is only microcrystalline recrystallized and the fine-grained nature of the matrix is well preserved. On the other hand, $Mg/(Mg+Fe)$ ratios in chondrule olivines (0.86, average) are similar to those in matrix olivines (0.8–0.85; these values are estimated from the intersection between the magnesian extension of the matrix trend in Fig. 7a and the line which indicates olivine solid solution). As described in Section 3.4, EPMA analysis suggests that the matrix contains abundant olivine (Fig. 7). Due to the small grain sizes of matrix olivine, the homogenization of olivines between chondrules and matrix occurred. Although bulk compositions of CAIs significantly changed during metamorphism (e.g. Fig. 17a), the textures of them changed to a lesser extent.

In CAIs, fine-grained materials fill the interstices of large spinel, plagioclase, and Ca-rich pyroxene crystals. The constituent minerals are also spinel, plagioclase, and Ca-rich pyroxene. However, their occurrences and their sub-μm grain sizes suggest that the fine-grained materials were formed by breakdown of primary mineral(s) in the CAIs. One of the plausible primary phases of the fine grained materials is melilite which is commonly observed in type A and B1 CAIs (e.g. MACPHERSON et al., 1988). The introduction of Na and Fe into the inclusions and chondrules is remarkable in Allende (e.g. HASHIMOTO and GROSSMAN, 1987; McGUIRE and HASHIMOTO, 1989; KIMURA and IKEDA, 1992). However, textures and compositions of CAIs and AOIs in Coolidge (for example, low Na_2O contents of them; McSWEEN, 1977) suggest that introduction of Na from the ambient circumstances did not take place extensively. Because we can see only the final products after thermal metamorphism and because I could not find melilite relics, it is difficult to specify the reactions which form the fine-grained materials in the CAIs. The following reaction is one of the plausible reactions to form the fine-grained materials in CAIs during thermal metamorphism:

$$4CaMgAl_2SiO_7 + 2CO_2 \rightarrow CaAl_2Si_2O_8 + 3MgAl_2O_4 + CaMgSi_2O_6 + 2CaCO_3,$$

   mel(geh)   gas   pl   sp   di   cal

where it is assumed that CO_2 gas was supplied by the reduction of ferromagnesian silicates by carbon. Supply of CO_2 gas during metamorphism is plausible because the carbon content of Coolidge is 0.19 wt%, which is much lower than that of the CV

3 chondrites (VAN SCHMUS, 1969). After or during the formation of these phases, exchange of Mg and Fe among olivines, pyroxene, and spinels, and that of Al and Cr among spinels probably occurred as a consequence of thermal metamorphism. This reaction suggests that bulk compositional change of CAIs, especially the increase of Cr_2O_3 concentration, might be enhanced by the formation of secondary spinel in them. As described in Section 4.6, calcite is observed in CAIs. Perhaps there is calcite which was formed during the above reaction.

As described in Section 4.4, spinels occur in CAIs, AOIs, chondrules (not associated with Fe-Ni metal and/or troilite), and Fe-Ni metal and/or troilite grains. They have different compositions according to their occurrences. The relationships among major and minor elements in spinels (Figs. 14 and 15) probably depend on both their primary compositions and compositional changes during metamorphism.

Chromites associated with Fe-Ni metal and/or troilite grains were reported by KIMURA (1984). He noted that such chromites in UOCs are especially ferroan and chromian (Mg/(Mg + Fe): 0.02–0.06; Y_{Cr}: >0.97) among spinel group minerals observed in a meteorite. Their occurrences suggest that they were formed by oxidation of Fe-Ni metals including Cr. Because the occurrences of chromian spinels in Coolidge are identical to those in the UOCs, the chromites in this meteorite were probably formed by the oxidation of Cr-bearing Fe-Ni metals, which are observed in CV3 chondrites. KIMURA (1984) also stated that chromites associated with Fe-Ni metal and/or troilite in equilibrated ordinary chondrites (EOCs) are less chromian and more magnesian than those in the UOCs with the same occurrences. Chromian spinels in this meteorite are much less chromian and much more magnesian than those in the EOCs. Most of them are plotted around Cr/(Cr + Al) ≈0.67 and Mg/(Mg + Fe) ≈0.46 (Fig. 14). On the other hand, those values in chromites in the H6 chondrites are around ∼0.87 and ∼0.15 (e.g. BUNCH et al., 1967; KRACHER and KURAT, 1980). Aluminous spinels are much more abundant in Coolidge than in the EOCs, and the modal ratio of aluminous spinels to chromites in this meteorite is much higher than that in the EOCs. Therefore, when the exchange of Al and Cr cations among spinels occurred, chromian spinels associated with Fe-Ni metal and/or troilite probably became more aluminous and magnesian during thermal metamorphism than those in the EOCs.

As described in Section 4.4, there are negative correlations of TiO_2 and V_2O_3 with Al_2O_3. On the other hand, there are positive correlations of ZnO with Al_2O_3, and of MnO with FeO. These correlations are not shown in spinels in inclusions in CV3 chondrites. For example, spinels in Allende CV3 chondrite (KORNACKI and WOOD, 1985) in spinel-rich inclusions have positive correlation between FeO and ZnO contents. This relationship is the opposite of that of Coolidge, because Al_2O_3 contents in spinels in Coolidge decreases as FeO content increases.

It is interesting that similar correlations to those in Coolidge are reported in spinels from EOCs. Strong positive correlations of MnO, V_2O_3, and TiO_2 with Mg/ (Mg + Fe) ratio are reported in spinels in ordinary chondrites (KRACHER and KURAT, 1980). YABUKI et al. (1983) noted a positive correlation of ZnO and Al_2O_3 and a high ZnO content (up to 2.3 wt%) in spinels in Arapahoe (L5). Considering
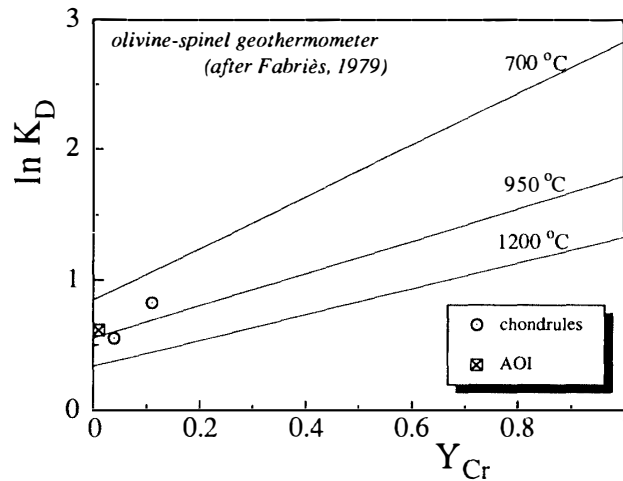
Fig. 19.   A plot of Y_{Cr} [$=Cr/(Cr+Al)$] vs. Fe-Mg distribution coefficient between olivine and
spinel. Isotherms are after FABRIÈS (1979). This geothermometry was applied for
the nearest pairs of olivine and spinel.

a strong positive correlation of Al_2O_3 and $Mg/(Mg+Fe)$ ratio in spinels in Coolidge, the correlations shown in spinels in Coolidge are consistent with those in EOCs. Because the positive correlation is observed in equilibrated chondrites, the correlations shown in spinels in Coolidge were probably established during thermal metamorphism.

ZnO content in more aluminous spinels in Coolidge is ~ 5 times lower than those in Arapahoe (L5) (YABUKI et al., 1983). However, the bulk composition of Coolidge shows similar Ni-normalized Zn abundance as bulk compositions of L chondrites (KALLEMEYN, 1987; WASSON and KALLEMEYN, 1988). This difference perhaps resulted from a higher abundance of aluminous spinels in Coolidge than that in L chondrites.

Interdiffusion of Al and Cr in spinel is probably slower than that between Fe and Mg in spinels. However, chromian spinels associated with Fe-Ni metal contain fairly large amounts of Al_2O_3 as discussed above. Therefore, maximum metamorphic temperatures must have been high enough to exchange Al and Cr cations among spinels. Olivine-spinel geothermometry suggests that Coolidge experienced high temperatures around $\sim 900°C$ (Fig. 19). Because $Cr/(Cr+Al)$ ratios of spinel are various, this temperature is an apparent one by local equilibrium. The Coolidge meteorite was probably heated to temperatures as high as $\sim 900°C$. However, this meteorite was not kept around $\sim 900°C$ for a long time, because of the presence of low-Ca clinopyroxene in this meteorite and Fe-Mg compositional zoning in the pyroxene. If this meteorite had been kept around $\sim 900°C$ for a long time and had cooled slowly, low-Ca pyroxene would have transformed into orthopyroxene and the composition of spinel would be homogenized irrespective of the difference in occurrences and glass in chondrules would be coarsely recrystallized. Even if the duration of thermal metamorphism of this meteorite is short, it is remarkable that the peak metamorphic temperature of this meteorite is much higher than those of

CK and ordinary chondrites belonging to petrologic type 4 (*e.g.* 720°C for Mara-linga CK4: GEIGER and BISCHOFF, 1991; 600–700°C for ordinary chondrites: McSWEEN *et al.*, 1988). Thermal metamorphism of this meteorite perhaps resulted from shock heating.

Although Mg/(Mg + Fe) ratios of pyroxenes varies in Coolidge, some minor elements in them probably changed during metamorphism. As described in Section 4.2, the TiO_2 and Na_2O contents of Ca-rich pyroxene in chondrules are positively correlated with Al_2O_3 contents (Fig. 12). Positive correlation among TiO_2 and Al_2O_3 wt% of Ca-rich pyroxenes is also observed in CV3, CO3, and CM2 chondrites (NOGUCHI, 1989). On the contrary, most of the Ca-rich pyroxenes in CV3, CO3, and CM2 chondrites include almost no Na_2O (<0.07 wt%; NOGUCHI, 1989). Ca-rich pyroxenes in unequilibrated ordinary chondrites (UOCs) include Na_2O. Their averages are from ~0.2 to ~0.3. However, their Na_2O contents are nega-tively correlated with Al_2O_3 contents (NOGUCHI, 1989). In contrast, UOCs of petrologic subtype >3.7 and some L4 chondrites involve Ca-rich pyroxenes with positive correlation between Al_2O_3 and Na_2O contents (NOGUCHI, unpublished data). Therefore, this positive correlation between Al_2O_3 and Na_2O contents in Ca-rich pyroxenes probably resulted from thermal metamorphism.

It is thought that the compositions and textures of Fe-Ni metals reflect a thermal history below ~500°C. As described in Sections 3.5 and 4.5, grain sizes of taenites in kamacites in Coolidge are smaller than those in the Efremovka CV3 chondrite. The ratios of the Co content in kamacite to the Co content in taenite in Coolidge and Efremovka are 3.27 and 5.17, respectively. Equilibration temperatures estimated by these ratios (after AFIATTALAB and WASSON, 1980) are ~540 and ~450°C, respectively. These estimated temperatures reflect the blocking tem-peratures at which Co stopped diffusing between kamacite and taenite. Because Efremovka shows the highest shock stage among the CV3 chondrites (SCOTT *et al.*, 1992), it is thought that this meteorite was reheated by shock to temperatures above ~450°C and then cooled slowly to around ~450°C. On the other hand, the higher estimated temperatures in Coolidge probably result from faster cooling around 500°C. In the other members of the "reduced group" of CV3 chondrites, the range of Ni contents in kamacite tends to be wider than that in Efremovka (see Fig. 3 in McSWEEN, 1977). These data suggest that kamacites in them probably do not exsolve coarse grained taenite, because either they did not experience a reheating event higher than 400–500°C or they experienced a short reheating event.

6.3. Similarities and differences between Coolidge and CV3 chondrites belonging to the "reduced group"

Type I chondrules in Efremovka are plotted near the point where the Ca/(Ca + Al) ratio is 2:1 on the edge between Ca and Al apices as well as chondrules in Coolidge in Al–Na–Ca diagrams (Figs. 17d and 18d). These diagrams suggest that ferromagnesian chondrules in Coolidge primarily had similar bulk compositions to type I chondrules in Efremovka. Not only bulk compositions but also textures of chondrules in Coolidge are similar to those in Efremovka. The Al–Na–Ca diagrams

also suggest that AOIs and CAIs in Coolidge had similar primary compositions to those in Efremovka. Therefore, the compositions of AOIs, CAIs and most of the chondrules in Coolidge and Efremovka are characterized by the depletion of Na (and K). The opaque mineralogy of Coolidge is also similar to that of Efremovka, as shown in Section 4.5.

However, the bulk composition of the matrix in Coolidge is different from that in Efremovka. Figure 17d shows that the matrix in Coolidge contains a component which is represented by anorthite-rich plagioclase and Ca-rich pyroxene. In contrast, Fig. 18d shows that the matrix in Efremovka include the component represented by plagioclase with higher amounts of Na (or nepheline) and Ca-rich pyroxene. Therefore, the matrix in Coolidge is depleted in Na. It is not yet clear at present that the enrichment of Na in the matrix in Efremovka is common to the other "reduced group" of CV3 chondrites. However, the depletions of Na in chondrules, AOIs, CAIs, and the matrix in Coolidge is clearly the cause for their depletion in the bulk composition of Coolidge as reported by KALLEMEYN (1987). He also gave the depletion of K as well as Na. It is implausible that the depletion took place by metamorphism or terrestrial weathering which occurred in Coolidge. If Na and K abundances of the matrix had been established during metamorphism, Na and K would have been redistributed to the other constituent components in this meteorite. If Na and K had been depleted by terrestrial weathering, abundant anorthite-rich plagioclase and chondrule glass rich in Ca and Al would have suffered alteration. Therefore, the depletions of Na and K were probably established before the metamorphism of Coolidge. It is thought that Coolidge is similar to the "reduced group" of CV3 chondrites, but was formed by agglomeration of materials depleted in Na and K.

6.4. Comparison of Coolidge and CK chondrites

Most of the CK chondrites except for Watson 002 (GEIGER et al., 1993) are equilibrated. They were formed under oxidizing conditions (e.g. GEIGER and BISCHOFF, 1989, 1991; NOGUCHI, 1993). In this section, mineralogical and petrologic differences between Coolidge and CK chondrites are discussed.

GEIGER and BISCHOFF (1989) noted some obvious mineralogical and petrologic differences between Coolidge and the CK chondrites. The most obvious differences are mineral species of opaque minerals and Ni content in olivine. CK chondrites contain abundant magnetite with small amounts of Ni-bearing sulfides, instead of kamacite, taenite and troilite which are abundant in Coolidge. Olivines in CK chondrites contain significant amounts of NiO (0.4–0.6 wt%; e.g. GEIGER and BISCHOFF, 1991; NOGUCHI, 1993). These features are related to the oxidizing conditions which CK chondrites experienced.

In Coolidge, spinel group minerals occur in CAIs, AOIs, chondrules, and Fe-Ni metal and/or troilite grains. In CK chondrites, spinels occur in CAIs, chondrules, and isolated opaque mineral inclusions which are composed mainly of magnetite and high-Ni sulfides. Although spinels exist in various occurrences in both meteorites, spinels in Coolidge have much wider compositional variations than those in CK

chondrites. The $Cr/(Cr+Al)$ and $Mg/(Mg+Fe)$ ratios in the spinels in Coolidge vary from 0.01 to 0.7, and from 0.45 to 0.77, respectively. On the other hand, spinels in CK chondrites are hercynite-rich spinels and $Cr/(Cr+Al)$ ratios do not change significantly (GEIGER and BISCHOFF, 1991; KELLER et al., 1992; NOGUCHI, 1993). Most remarkable compositional differences are shown between chromian spinels associated with Fe-Ni metal grains and hercynite-rich spinels associated with opaque mineral inclusions. These compositional differences are probably related to the differences of coexisting spinel-group minerals. Because CK chondrites contain abundant magnetite which contains considerable amounts of Cr, Al, Mg, and Ti, the distributions of Cr and Al were quite different between spinels in Coolidge and those in CK chondrites.

Bulk composition of Coolidge is different from bulk compositions of CK chondrites (KALLEMEYN, 1987; KALLEMEYN et al., 1991). Na and K are depleted in Coolidge and not depleted in CK chondrites, when these elements are compared to the average of CV chondrites. The existence of Fe-Ni metals and the absence of magnetite in Coolidge indicate that it experienced reducing conditions after agglomeration. Therefore, it is clear that primary compositions of the constituent components of Coolidge were different from those of CK chondrites and that Coolidge experienced more reduced conditions before and after agglomeration of its parent body than CK chondrites.

Although bulk compositions of all the constituent components of Coolidge are depleted in Na and K, some chondrules and CAIs contain relatively sodic plagioclase (Fig. 13). The compositional range of plagioclase in Coolidge is as large as the ranges in CK chondrites (e.g. SCOTT and TAYLOR, 1985; GEIGER and BISCHOFF, 1991; NOGUCHI, 1993). However, the occurrences of plagioclase are different between Coolidge and CK chondrites. For example, coarse sodic plagioclase crystals ($<$about $100\,\mu m$ across) are observed even in the matrix in the Karoonda CK4 chondrite (NOGUCHI, 1993), although Karoonda is less equilibrated than Coolidge (SCOTT and TAYLOR, 1985). Sodic plagioclases in CK matrices might be condensates, because they can condense from a protosolar nebula under highly oxidized conditions (WOOD and HASHIMOTO, 1993). EPMA analysis suggests that the matrix in Coolidge includes an anorthite-rich component, but large plagioclase crystals were not observed by SEM. Plagioclases in Coolidge do not show reverse zoning which is often observed in CK chondrites (NOGUCHI, 1993).

7. Conclusions

1) The Coolidge meteorite has some similarities to the "reduced group" of CV3 chondrites. (1) Abundant Fe-Ni metal and sulfide are kamacite, taenite, and troilite. (2) AOIs, CAIs, and chondrules are depleted in Na and K. (3) The main constituent mineral of matrix is olivine.

2) There are also some differences between them. The differences can be divided into primary and secondary ones. Secondary differences are related to thermal metamorphism which Coolidge experienced. The differences which are

thought to be primary are the following: (1) The matrix as well as AOIs, CAIs, and chondrules in Coolidge is depleted in Na, although the matrix in Efremovka is not so depleted in Na. The depletions of Na abundances in bulk compositions of Coolidge relative to the average of CV3 chondrites (KALLEMEYN, 1987) correspond to this difference. (2) There are chondrules which are composed of plagioclase, pyroxene, and silica mineral. (3) The abundance of POP chondrules including acicular to prismatic plagioclase crystals are higher in Coolidge than CV3s. The differences formed by thermal metamorphism are the following. (1) Bulk compositions of CAIs are rich in FeO and Cr_2O_3. (2) Compositional variations of spinel in Coolidge are very wide. $Cr/(Cr+Al)$ ratio varies from 0.01 to 0.7, and $Mg/(Mg+Fe)$ ratio varies from 0.45 to 0.77. (3) Correlations between some elements in spinel and those in pyroxenes are not shown in the "reduced group" of CV3 chondrites but are shown in EOCs.

    3) Olivine-spinel geothermometry suggest that Coolidge was heated to the temperatures as high as $\sim900°C$. The duration of thermal metamorphism was short enough to preserve many petrographic features such as Fe-Mg compositional zoning in low-Ca pyroxene. The distribution of Co between kamacite and taenite suggests fast cooling around $\sim500°C$.

    4) Although Coolidge and most of the CK chondrites are equilibrated chondrites, there are differences in mineralogy and petrology. (1) As described above, olivine is the main constituent mineral of the matrix in Coolidge. The matrix minerals of the Karoonda CK4 chondrite are olivine and sodic plagioclase. This difference results from Na and K depletions in the Coolidge matrix. The Coolidge matrix is finer than the matrices in CKs. (2) Compositions of plagioclase in Coolidge are as various as those in the CK chondrites. However, plagioclase does not occur in the matrix in Coolidge. Plagioclase in CK chondrites often shows reverse zoning up to 40 mol% An. In Coolidge, compositional variation within a single plagioclase grain is much smaller than that in the CK chondrites. (3) Spinel group minerals exist in various occurrences in Coolidge and CK chondrites. However, their compositional variations are much wider in Coolidge than those in CK chondrites. Although spinels in Coolidge change both $Mg/(Mg+Fe)$ and $Cr/(Cr+Al)$ ratios, those in CK chondrites change mainly $Mg/(Mg+Fe)$ ratios.

Acknowledgments

References

Afiattalab, F. and Wasson, J. T. (1980) Composition of the metal phases in ordinary chondrites: implications regarding classification and metamorphism. Geochim. Cosmochim. Acta, **44**, 431–446.

Bence, A. E. and Ablee, A. L. (1968) Empirical correction factors for the electron microanalysis of silicates and oxides. J. Geol., **76**, 382–403.

Bischoff, A. and Keil, K. (1983a) Ca-Al-rich chondrules and inclusions in ordinary chondrites. Nature, **303**, 588–592.

Bischoff, A. and Keil, K. (1983b) Ca-Al-rich chondrules and inclusions in ordinary chondrites: Evidence for a related genesis of ordinary and carbonaceous chondrites. Lunar and Planetary Science XIV. Houston, Lunar Planet. Inst., 47–48.

Bischoff, A. and Keil, K. (1983c): Catalog of Al-rich chondrules, inclusions and fragments in ordinary chondrites. Spec. Publ. No. 22, Univ. of New Mexico, Institute of Meteoritics, 1–33.

Bischoff, A. and Keil, K. (1984): Al-rich objects in ordinary chondrites: Related origin of carbonaceous and ordinary chondrites and their constituents. Geochim. Cosmochim. Acta, **48**, 693–709.

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

Dodd, R. T. (1981): Meteorites. Cambridge, Cambridge Univ. Press, 368 p.

Fleischer, M., Chao, G. Y. and Kato, A. (1975): New mineral names. Am. Mineral., **60**, 485–486.

Gasparick, T. (1984): Two-pyroxene thermobarometry with new experimental data in the system CaO-MgO-Al$_2$O$_3$-SiO$_2$. Contrib. Mineral. Petrol., **87**, 87–97.

Geiger, T. and Bischoff, A. (1989): Mineralogy of metamorphosed carbonaceous chondrites. Meteoritics, **24**, 269–270.

Geiger, T. and Bischoff, A. (1991): The CK chondrites-Conditions of parent body metamorphism. Meteoritics, **26**, 337.

Geiger, T., Spettel, B., Clayton, R. N., Mayeda, T. K. and Bischoff, A. (1993): Watson 002–The first CK/type 3 chondrite. Meteoritics, **28**, 352.

Gooding, J. L. (1981): Mineralogical aspects of terrestrial weathering effects in chondrites from Allan Hills, Antarctica. Proc. Lunar Planet. Sci., 12B, 1105–1122.

Grossman, J. N., Rubin, A. E., Nagahara, H. and King, E. A. (1988): Properties of chondrules. Chondrites and the Early Solar System, ed. by J. F. Kerridge and M. S. Matthews. Tucson, Univ. Arizona Press, 436–461.

Hashimoto, A. and Grossman, L. (1987): Alteration of Al-rich inclusions inside amoeboid olivine aggregates in the Allende meteorite. Geochim. Cosmochim. Acta, **51**, 1685–1704.

Ikeda, Y. and Kojima, H. (1991): Terrestrial alteration of Fe-Ni metals in antarctic ordinary chondrites and the relationship to their terrestrial ages. Proc. NIPR Symp. Antarct. Meteorites, **4**, 307–318.

Johnson, C. A. and Prinz, M. (1991): Chromite and olivine in type II chondrules in carbonaceous and ordinary chondrites: Implications for thermal histories and group differences. Geochim. Cosmochim. Acta, **55**, 891–904.

Kallemeyn, G. W. (1987): Compositional comparisons of metamorphosed carbonaceous chondrites. Mem. Natl Inst. Polar Res., Spec. Issue, **46**, 151–161.

Kallemeyn, G. W., Rubin, A. E. and Wasson, J. T. (1991): The compositional classification of chondrites: V. The Karoonda (CK) group of carbonaceous chondrites. Geochim. Cosmochim. Acta, **55**, 881–892.

Keller, L. P., Clark, J, C., Lewis, C. F. and Moore, C. B. (1992): Maralinga, a metamorphosed carbonaceous chondrite found in Australia. Meteoritics, **27**, 87–91.

Kimura, M. (1984): Spinel group minerals in unequilibrated ordinary chondrites. Proc. 17th ISAS

Lunar Planet. Symp., 37–38.

KIMURA, M. and IKEDA, Y. (1992): Pre-accretionary alteration of chondrules in Allende meteorite. Papers Presented to the 17th Symposium on the Antarctic Meteorites, 19–21 August 1992. Tokyo, Natl Inst. Polar Res., 31–33.

KORNACKI, A. S. and WOOD, J. A. (1985): Mineral chemistry and origin of spinel-rich inclusions in the Allende CV3 chondrite. Geochim. Cosmochim. Acta, 49, 1219–1237.

KORNACKI, A. S., COHEN, R. E. and WOOD, J. A. (1983): Petrography and classification of refractory inclusions in the Allende and Mokoia CV3 chondrites. Mem. Natl Inst. Polar Res., Spec. Issue, 30, 36–59.

KRACHER, A. and KURAT, G. (1980): Ordinary chondrites: The spinel puzzle. Meteoritics, 15, 319–320.

MACPHERSON, G. J., WARK, D. A. and ARMSTRONG, J. T. (1988): Primitive material surviving in chondrites: Refractory inclusions. Chondrites and the Early Solar System, ed. by J. F. KERRIDGE and M. S. MATTHEWS. Tucson, Univ. Arizona Press, 746–807.

MASON, B. (1963): Olivine composition in chondrites. Geochim Cosmochim. Acta, 27, 1011–1023.

MCGUIRE, A. V. and HASHIMOTO, A. (1989): Origin of zoned fine-grained inclusions in the Allende meteorite. Geochim. Cosmochim. Acta, 53, 1123–1133.

MCSWEEN, H. Y. (1977): Petrographic variations among carbonaceous chondrites of the Vigarano type. Geochim. Cosmochim. Acta, 44, 1777–1790.

MCSWEEN, H. Y., SEARS, D. W. G. and DODD, R. T. (1988): Thermal metamorphism. Chondrites and the Early Solar System, ed. by J. F. KERRIDGE and M. S. MATTHEWS. Tucson, Univ. Arizona Press, 102–113.

NOGUCHI, T. (1989): Texture and chemical composition of pyroxenes in chondrules in carbonaceous and unequilibrated ordinary chondrites. Proc. NIPR Symp. Antarct. Meteorites, 2, 169–199.

NOGUCHI, T. (1993): Petrology and mineralogy of CK chondrites: Implications for the metamorphism of the CK chondrite parent body. Proc. NIPR Symp. Antarct. Meteorites, 6, 204–233.

PALME, H., LARIMER, J. W. and LIPSCHUTZ, M. E. (1988): Moderately volatile elements. Chondrites and the Early Solar System, ed. by J. F. KERRIDGE and M. S. MATTHEWS. Tucson, Univ. Arizona Press, 436–461.

RUSSELL, J. D. (1979): Infrared spectroscopy of ferrihydrite: Evidence for the presence of structural hydroxyl groups. Clay Mineral., 14, 109–113.

SCOTT, E. R. D. and TAYLOR, G. J. (1985): Petrology of types 4-6 carbonaceous chondrites. Proc. Lunar Planet. Sci. Conf., 15th, Pt. 1, C699-C709 (J. Geophys. Res., 90 suppl.).

SCOTT, E. R. D., KEIL, K. and STÖFFLER, D. (1992): Shock metamorphism of carbonaceous chondrites. Geochim. Cosmochim. Acta, 56, 4281–4293.

TESHIMA, J. and WASSERBURG, G. J. (1985): Texture, metamorphism and origin of type A CAI's. Lunar and Planetary Science XVI. Houston, Lunar Planet. Inst., 855–856.

TSUCHIYAMA, A., FUJITA, T. and MORIMOTO, N. (1988): Fe-Mg heterogeneity in the low-Ca pyroxenes during metamorphism of the ordinary chondrites. Proc. NIPR Symp. Antarct. Meteorites, 1, 173–184.

VAN SCHMUS, W. R. (1969): Mineralogy, petrology, and classification of types 3 and 4 carbonaceous chondrites. Meteorite Research, ed. by P. M. MILLMAN. Dordrecht, D. Reidel, 480–491.

VAN SCHMUS, W. R. and WOOD, J. A. (1967): A chemical-petrologic classification for the chondritic meteorites. Geochim. Cosmochim. Acta, 31, 737–765.

WARK, D. A. (1987): Plagioclase-rich inclusions in carbonaceous chondrite meteorites: Liquid condensates? Geochim. Cosmochim. Acta, 51, 221–242.

WASSON, J. T. and KALLEMEYN, G. W. (1988): Composition of chondrites. Philos. Trans. R. Soc. London, A325, 535–544.

WATANABE, S. and MORIMOTO, N. (1987): A subsilicic aluminian pyroxene pair in a fragment of Y-82038 (H3). Papers Presented to the 12th Symposium on the Antarctic Meteorites, 8–10 June 1987. Tokyo, Natl Inst. Polar Res., 44–46.

WEISBERG, M. K., PRINZ, M., KOJIMA, H., YANAI, K., CLAYTON, R. N. and MAYEDA, T. K. (1991):

The Carlisle Lakes-type chondrites: A new grouplet with high $\Delta^{17}O$ and evidence for nebular oxidation. Geochim. Cosmochim. Acta, **55**, 2657–2669.

Wood, J. A. and Hashimoto, A. (1993): Mineral equilibrium in fractionated nebular systems. Geochim. Cosmochim. Acta, **57**, 2377–2388.

Yabuki, H., El Goresy, A. and Ramdhor, P. (1983): A petrologic microprobe survey of coexisting olivines, pyroxenes, and spinels in L- and LL-chondrites. Meteoritics, **18**, 426–428.