

THE MINERALOGY AND PETROLOGY OF CHONDRITIC METEORITES

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SUMMARY

Chondritic meteorites are naturally divided into five chemical groups according to their chemical compositions. Within each group a range of textures exists which is generally interpreted as representing progressive degrees of recrystallization, although there are also significant primary variations in each group. In most chondrites the major minerals are olivine, orthopyroxene, albitic feldspar, troilite, calcic pyroxene, and metallic nickel–iron; the compositions of the silicate minerals are usually uniform within recrystallized chondrites and very similar for individual chondrites belonging to a given group. In unrecrystallized chondrites the minerals are usually quite variable in composition and the primary structures of individual chondrules are well preserved. Most of the properties of chondrites can be interpreted in terms of metamorphic equilibration and recrystallization of primary chondritic matter, although alternative hypotheses exist.

INTRODUCTION

Meteorites have been aptly called “the poor man’s space probe” (ANDERS, 1968), since they provide much data about the extraterrestrial environment, particularly cosmic rays in the solar system. In addition, meteorites could also be considered the poor man’s time machine, since meteorites, which are 4.6–4.7 billion years old, are by far the oldest objects in the solar system that can currently be studied in terrestrial laboratories. Thus, it is through the study of meteorites that many of our ideas on the origin of the solar system have developed. Of the several kinds of meteorites, those known as *chondrites* stand the best chance for being samples of the primitive solid matter from which planets formed.

Since chondritic meteorites consist of 70–80% silicate minerals, it is only natural that mineralogic and petrologic studies on chondrites are essential to their thorough understanding. This phase of chondrite research has expanded considerably in the last six years, primarily as a result of the development of electron microprobe analysis as a technique for studying the compositions of fine-grained minerals *in situ*. The point has now been reached where most of the comprehensive studies have been completed, and future research will probably be devoted in large part to detailed studies of individual chondrites. Therefore, it is appropriate

at this time to review the current state of our knowledge of the mineralogy and petrology of chondrites.

In the following presentation I have not attempted to include all references to a particular topic, but have selected those that provide the latest or most complete information. The references that have been cited can, however, serve as guides to further reading for anyone so interested. For good discussions of the relationships of chondrites to other meteorite types, the general works by KRINOV (1960), MASON (1962a, 1967a, b, 1968), WOOD (1963a, 1968a) and ANDERS (1964) are recommended.

CLASSIFICATION

One of the more bewildering aspects of meteoritics to someone newly introduced to the field (and also to many people active in the field) is that of classification. The classification that will be followed in this paper is largely that of VAN SCHMUS and WOOD (1967) in which five recognized chemical *groups* are divided into six petrologic *types*. Thus, any chondrite can be characterized chemically and petrologically by stating its group affinity and petrologic type. This classification, as well as commonly used synonyms for different kinds of chondrites, is presented in Fig.1 and discussed below. Table I gives the number of chondrites of each type; most C group and E group chondrites are included in Table I, but only about half of each of the other groups have been classified. In the following discussion, the historical development of classification has been omitted and the interested reader should refer to VAN SCHMUS and WOOD (1967) as well as to UREY and CRAIG (1953), WIJK (1956), MASON (1962b, 1962c, 1965a, 1966), YAVNELL (1963), AHRENS (1964, 1965), CRAIG (1964), KEIL and FREDRIKSSON (1964), MASON and WIJK (1964), DODD and VAN SCHMUS (1965), FREDRIKSSON et al. (1968), KEIL (1968), and VAN SCHMUS (1969).

	PETROLOGIC TYPE					
	1	2	3	4	5	6
E				Enstatite Chondrites	Type I	Type II
C	Carbonaceous Ch.					
	Type I	Type II	Type III			
H				Bronzite Chondrites		
L					Hypersthene Chond.	
LL			Unequilibrated Ordinary Chondrites			Amphoteric Chond.

Fig.1. A presentation of VAN SCHMUS and WOOD's (1967) classification of chondrites showing the equivalences of commonly used names for various types of chondrites. No chondrites are known from the stippled areas and there is no synonym for C4 chondrites.

TABLE I

ABUNDANCES OF CHONDRITE TYPES¹

Chemical group	Petrologic type					
	1	2	3	4	5	6
E	0	0	1	4	2	6
C	4	16	8	2	0	0
H	0	0	7	35	74	44
L	0	0	9	18	43	152
LL	0	0	4	3	7	21

¹ From VAN SCHMUS and WOOD (1967). The numbers give the number of chondrites that are of a particular type; only classified chondrites (about $\frac{1}{2}$ of total) are tabulated.

TABLE II

SUMMARY OF CHEMICAL GROUP PARAMETERS¹

Group	Fe/Si	Fe°/Fe	Si/Mg	Fa
E	0.83 \pm 0.32	0.80 \pm 0.10	1.27 \pm 0.10	—
C	0.83 \pm 0.08	low	0.95 \pm 0.03	—
H	0.83 \pm 0.08	0.63 \pm 0.07	1.04 \pm 0.03	16-20
L	0.59 \pm 0.05	0.33 \pm 0.07	1.07 \pm 0.03	22-26
LL	0.53 \pm 0.03	0.08 \pm 0.07	1.06 \pm 0.03	27-31

¹ After VAN SCHMUS and WOOD (1967). Ratios are atomic ratios. Fe = total iron; Fe° = metallic iron. Fa = mole % fayalite in olivine for recrystallized chondrites.

Chemical groups

If bulk chemical analyses of individual chondrites are examined in detail (particularly on a volatile-free basis), it can easily be demonstrated that each analysis will fall into one of five distinct groups. Each group is a natural entity and is not dependent upon arbitrarily defined limits. Table II summarizes some of the chemical relationships that are pertinent to recognition of the five groups. Much of the ranges quoted in Table II represents analytical errors, so that actual clustering of parameters is probably much more pronounced than indicated (see later sections). Of the parameters in Table II, total iron content and amount of oxidized iron are the most commonly used (i.e., Fe/Si and Fe°/Fe or mole % Fa, respectively), but the Si/Mg ratio has become increasingly important in distinguishing H, L, and LL group chondrites from C and E group chondrites.

The five groups referred to by VAN SCHMUS and WOOD (1967) as E, C, H, L, and LL groups¹ are also commonly referred to as enstatite (E), carbonaceous (C), bronzite (H), hypersthene (L), and amphoteric (LL) chondrites. Because of their abundances and chemical similarities (apart from total iron content and amount of oxidized iron), chondrites belonging to the H, L, and LL groups are commonly called "ordinary chondrites". In this paper I will use the terms enstatite or E group and carbonaceous or C group interchangeably as convenient, but will use only H, L, and LL group (instead of bronzite, hypersthene, and amphoteric) for the ordinary chondrites. I do this because current definitions of "bronzite" and "hypersthene" (e.g., DEER et al., 1963) place all orthopyroxene of recrystallized ordinary chondrites in the bronzite range. CRAIG (1964), KEIL and FREDRIKSSON (1964), MASON (1965a) and REID and FREDRIKSSON (1967) have recently discussed this problem in detail.

Enstatite chondrites are extremely reduced; all Fe is present as FeS or metallic Fe-Ni, and many other normally lithophilic elements (Mn, Mg, Ca, Cr, K) are present in sulfide phases (MASON, 1966; KEIL, 1968). Si/Mg ratios in enstatite chondrites are greater than those of either ordinary chondrites or carbonaceous chondrites (AHRENS, 1965; MASON, 1966; KEIL, 1968).

In contrast to enstatite chondrites, most of the iron in carbonaceous chondrites is present as Fe^{2+} or Fe^{3+} in silicates, oxides, or sulfides. Carbon and water are major constituents of most carbonaceous chondrites (hence the name) and WIJK (1956) proposed a three-fold subdivision into type I, type II, and type III carbonaceous chondrites as follows: type I, ~ 20% H_2O and 3–5% C; type II, 10–15% H_2O and 1–3% C; type III, ~ 1% H_2O and $\frac{1}{2}$ –1% C. In terms of non-volatile elements, the most diagnostic parameter for classification is the Si/Mg ratio, which is lower for carbonaceous chondrites than for ordinary or enstatite chondrites (UREY, 1961; AHRENS, 1965; AHRENS et al., 1969).

The ordinary chondrite groups are quite similar to one another in their compositions, apart from total iron and nickel and amount of oxidized iron. They all have Si/Mg ratios intermediate to those of carbonaceous or enstatite chondrites. The major distinctions among these three groups are: (a) the total Fe and Ni content; and (b) the amount of the iron that occurs as Fe^{2+} in silicates. As can be seen from Table II, these two parameters have an inverse relationship to one another. There has been much discussion in recent years as to whether LL group chondrites represent a distinct group (KEIL and FREDRIKSSON, 1964; VAN SCHMUS and WOOD, 1967; FREDRIKSSON et al., 1968) or are merely a variety of L group chondrites (MASON, 1965a). However, there is now a sizeable body of data documenting chemical, mineralogical, and physical differences between L and LL group

¹ The use of "H" and "L" was originated by UREY and CRAIG (1953) for high-iron and low-iron chondrites, respectively. KEIL and FREDRIKSSON (1964) proposed the term "LL" for chondrites that are both low in total iron and low in metallic Fe-Ni (~ 1–5%).

chondrites, so that there now seems to be little doubt that each group represents a separate evolutionary sequence. Some of the differences will be referred to in this paper; others may be found in MASON and WIIK (1964), REID and FREDRIKSSON (1967), VAN SCHMUS and WOOD (1967) and FREDRIKSSON et al., (1968).

Petrologic types

Although recognition of the chemical groups described above has been developing for some time, subdivision of groups into six petrologic types is more recent¹ (VAN SCHMUS and WOOD, 1967) and is not yet as widely used as the chemical groups. However, the petrologic subdivisions have been widely used and many recent studies have confirmed their validity, although the original premise that prompted their definition (metamorphic equilibration) is still a subject of considerable discussion.

The petrologic types are unlike the chemical groups in that they are essentially defined subdivisions rather than naturally distinct groupings. Descriptions of the mineralogic and petrologic properties of the six types are covered in later sections of this paper. At this point it is sufficient to state that the six types represent mineral assemblages which more closely approach high-temperature equilibration with increasing type number, and progressive recrystallization for type 4 through type 6. For example, type 1 chondrites contain low-temperature minerals that are not in equilibrium, whereas type 6 chondrites are highly recrystallized and contain mineral assemblages that show a close approach to high-temperature mineral assemblages appropriate to their bulk chemical compositions.

From Fig. 1 and Table I it is apparent that not all types are present in all groups. Thus, the only known chondrites with type 1 or type 2 textures have C group chemical properties, and there are few C group chondrites that show any evidence of recrystallization. It can also be seen that Wiik's types I, II, and III are equivalent to Van Schmus and Wood's C1, C2, and C3 chondrites. ANDERS (1964) referred to the lesser recrystallized enstatite chondrites as type I and the strongly recrystallized enstatite chondrites as type II, and this terminology has been retained by others (MASON, 1966; KEIL, 1968). Keil recognizes type I, intermediate, and type II enstatite chondrites, which are equivalent to (E3 + E4), E5, and E6, respectively, in Van Schmus and Wood's classification.

¹ If one were to consider Van Schmus and Wood's classification purely descriptive, it would be more proper to refer to their subdivisions as *petrographic* types rather than petrologic types. In the view of VAN SCHMUS and WOOD (1967), the types are petrologic in the sense that meteorites of each petrologic type have been subjected to approximately the same degree of recrystallization. Petrographic characteristics have frequently been used in early classification schemes, but in general they have been too superficial in their definition or too many subdivisions have been created. MASON (1962a) presents a good discussion of older classifications.

CHEMISTRY OF CHONDRITES

There are three major aspects of chondrite compositions that are important to their mineralogy and petrology. These are: (a) the degree of homogeneity within a group; (b) the nature of chemical variations that do occur within groups; and (c) the differences between groups. The general compositional differences between groups can be seen from Table III.

TABLE III

CHEMICAL ANALYSES OF REPRESENTATIVE CHONDRITES (WT. %).¹

	<i>1</i>	<i>2</i>	<i>3</i>	<i>4</i>	<i>5</i>
SiO ₂	35.26	37.06	39.29	40.96	28.69
MgO	17.48	23.62	24.78	25.70	19.77
FeO	—	9.89	14.96	18.90	21.08
Al ₂ O ₃	1.45	2.09	2.49	2.23	2.19
CaO	0.95	1.75	1.62	1.62	1.92
Na ₂ O	1.01	0.99	0.93	0.84	0.22
K ₂ O	0.11	0.07	0.10	0.12	0.04
Cr ₂ O ₃	0.47	0.54	0.55	0.59	0.44
MnO	0.25	0.28	0.33	0.35	0.21
TiO ₂	0.06	0.15	0.12	0.18	0.09
P ₂ O ₅	0.52	0.34	0.30	0.20	0.32
Fe	24.13	16.21	6.68	1.46	—
Ni	1.83	1.65	1.30	0.99	1.18
Co	0.08	0.10	0.08	0.06	0.06
FeS	14.20	5.21	6.46	6.36	7.67
C	0.43	n.d.	0.18	n.d.	2.78
H ₂ O	1.17	0.39	n.d.	0.13	12.42
Total	99.40	100.34	100.17	100.69	99.08
Total Fe	33.15	27.20	22.42	20.18	21.25
Fe/Si (atomic)	1.01	0.79	0.66	0.53	0.80
Si/Mg (atomic)	1.35	1.05	1.06	1.07	0.97
FeO/Total Fe (atomic)	—	0.27	0.52	0.73	0.77

¹ The analyses are presented in a standard format, and the actual chemical form of some of the elements may be somewhat different, especially for the enstatite chondrite (analysis 1).

The analyses are: *1* = Indarch (Wiik, 1956), a type I enstatite chondrite; *2* = Forest City (MASON and Wiik, 1965), a type 5 H-group chondrite; *3* = Modoc (MASON and Wiik, 1967), a type 6 L-group chondrite; *4* = Näs (MASON and Wiik, 1964), a type 6 LL-group chondrite; *5* = Murray (Wiik, 1956), a type 2 carbonaceous chondrite.

A comprehensive discussion of analytic problems and techniques relative to determining elemental abundances in chondrites is beyond the scope of this paper. It can be safely stated, however, that many wet chemical analyses of chondrites

in the literature are not accurate (or, perhaps, precise) enough to allow detailed examination of minor chemical variations for major (Fe, Mg, Si) or minor (Ca, Al, Na, K, Mn, Cr, Ti, P, Ni, Co) elemental abundances within or between chondrite groups. Certain variations in these elements have been noted, however, that are outside reasonable limits of error.

EDWARDS (1955), EDWARDS and UREY (1955), KAISER and ZÄHRINGER (1965) and GOPALAN and WETHERILL (1969) have demonstrated that potassium abundances in LL group chondrites vary from about 200 p.p.m. to about 1,100 p.p.m. from chondrite to chondrite; the abundance also varies within some chondrites. This is in marked contrast to potassium abundances in H and L group chondrites, which are relatively uniform at about 800–850 p.p.m.

The total iron content of enstatite chondrites varies more or less continuously from a high of 35.0% to a low of 20.7%. The variation is such that type I (unrecrystallized) enstatite chondrites have the higher iron contents, while type II (recrystallized) enstatite chondrites have the lower iron contents (MASON, 1966; KEIL, 1968). In addition, the abundances of Ni and S, and the Si/Mg ratios in enstatite chondrites decrease from type I to type II (KEIL, 1968).

Variations in carbon and water content among carbonaceous chondrites were mentioned above pertaining to classification. In addition, MOORE and LEWIS (1967) have shown that carbon varies from about $\frac{1}{2}$ –1% in type 3 ordinary chondrites to less than 0.1% in type 6 ordinary chondrites.

Recently, AHRENS et al. (1969) have studied major and minor element abundances in chondrites using X-ray fluorescence analysis and report that Si/Mg, Si/Al, Si/Ca, Si/Ti, and Al/Ca ratios from one ordinary chondrite to another (in the same group) vary only about 1–2%, which is near the limit of precision. They have also confirmed the Si/Mg variation among enstatite chondrites. Thus, except for a few cases, it appears that the major and minor elemental compositions of individual chondrites are amazingly uniform, both within chondrites (on a scale of centimeters) and among individual chondrites of a given group. Such homogeneity of compositions is a fundamental characteristic of chondrites.

The abundances of trace elements vary significantly within E, H, L and LL groups. These variations are chiefly in the abundances of the so-called "volatile trace elements" (e.g., Hg, Pb, Bi, Cd, In, Cl) and the heavy primordial rare gases (Ar, Kr, and Xe). Recent papers on these variations include those of LARIMER and ANDERS (1967), MARTI (1967), OTTING and ZÄHRINGER (1967), ANDERS (1968), HEYMANN and MAZOR (1968), TANDON and WASSON (1968) and ZÄHRINGER (1968). In brief, available data yield two important relationships:

(a) The abundances of these elements vary over 2–3 orders of magnitude (per element) such that type 3 (unrecrystallized) chondrites contain larger abundances and type 6 (well-recrystallized) chondrites contain lower abundances;

(b) The ratios of these elements relative to one another vary only slightly, in contrast to large variations in absolute abundance.

In addition, abundances vary among C1, C2, and C3 chondrites in a manner that indicates these elements may be concentrated in low-temperature (matrix) components and depleted in the high-temperature component (chondrules) of these chondrites (ANDERS, 1968).

Prior's "Rules"

PRIOR (1916) was one of the first to recognize that chondrites could be classified according to the amount of FeO and metallic iron in them, and his subdivisions were the fore-runners of the present E, H, L, and LL groups. Prior noted "... in meteoritic stones generally, the poorer they are in nickel-iron, the richer that iron is in nickel, and the richer in iron are the magnesium silicates" (PRIOR, 1916, p.44). Prior did not recognize the differences in total iron, however, primarily due to the many poor analyses then available. As a result of his observations, Prior postulated that all chondrites came from a common source and differed only in their degree of oxidation. Thus, he envisioned a continuum from the reduced enstatite chondrites to the oxidized LL group chondrites.

UREY and CRAIG (1953), MASON (1962a, b), CRAIG (1964), and KEIL and FREDRIKSSON (1964) have subsequently shown that chondrites do not represent a continuum, and since they are clustered into five distinct groups, they are accumulates from five more or less distinct environments (CRAIG, 1964). None-the-less, many authors have still attempted to show that an oxidation sequence of the type postulated by Prior does exist, at least within groups (MASON, 1965a; MUELLER and OLSEN, 1967).

As our knowledge of chondrite compositions has improved, the compositional clusters have become smaller and smaller. Thus, since it has now been demonstrated that Si/Mg, Si/Ca, Si/Al, etc. ratios are constant to 1–2 % within ordinary chondrite groups, any increase in Fe^{2+} must be accompanied by an increase in the abundance and fayalite content of the olivine. From the known compositions of olivine and their ranges (KEIL and FREDRIKSSON, 1964; FREDRIKSSON et al., 1968) it can be demonstrated that the ranges in FeO and metallic Fe values reported in chondrite chemical analyses far exceed those permitted by norm calculations. Thus, any attempt to evaluate Prior's rules within groups using wet chemical analyses for FeO and metallic Fe content are highly questionable, and other, more precise, ways of evaluating in-group variations must be sought.

The above comments do not lessen the importance of the qualitative aspects of Prior's rules: Why are H group chondrites, with the most total iron and the most metallic iron, more reduced than LL group chondrites, which contain the least total iron and the least metallic iron? This, in turn, is only part of the larger problem: What is the origin of the compositional hiatuses, in general, among the different groups? To date there is no good answer to either question.

Organic compounds

Not included in the above discussions is the question of organic compounds in chondrites, a topic of much controversy in recent years, the principal argument being whether the compounds are the result of biologic or non-biologic synthesis. Although this problem is not fully resolved at present, it appears that those advocating non-biologic synthesis are probably nearer the truth. Recent papers by NAGY (1966), UREY (1966), HAYES (1967), NOONER and ORO (1967), OLSON et al. (1967), HAYATSU et al. (1968) and STUDIER et al. (1968) are good sources for further reading on this very provocative topic.

AGES

Several types of ages have been determined on chondrites, most of which are relevant to mineralogic and petrologic aspects of chondrite origin and evolution.

Absolute ages

Rb-Sr age studies on chondrites (GOPALAN and WETHERILL, 1968, 1969; KAUSHAL and WETHERILL, 1969; KEMPE and MÜLLER, 1969) indicate that they formed as solid bodies 4.5–4.7 billion years ago. Although some differences exist in ages from different chondrite groups (and different laboratories), the ages generally agree within experimental error; from absolute age determinations there is no good evidence at present for one chondrite group being older than another.

Formation interval ages

Recent studies on the decay of now-extinct ^{129}I to ^{129}Xe (cf., HOHENBERG et al., 1967a; ZÄHRINGER, 1968) and the decay of now extinct ^{244}Pu (HOHENBERG et al., 1967b; RAO et al., 1968) indicate that chondrites began to retain xenon 60–150 million years after isolation of the primitive solar nebula from galactic nucleosynthesis. Furthermore, the data of HOHENBERG et al. (1967a) currently indicate that this time interval was the same (\pm 2 million years) for all groups and all petrologic types, but it is still uncertain whether the event dated is that of chondrule formation or of cooling after recrystallization (WOOD, 1968b).

Gas retention ages

K-Ar ages and Th, U-He ages of chondrites are in the range 4.0–4.5 billion years for most H, LL, and E group chondrites (ZÄHRINGER, 1968). For L group chondrites, however, most K-Ar ages range from 4.0 billion years to 0.5 billion years, and HEYMANN (1967) has recently shown that L group chondrites were involved in a cosmic collision about 500 million years ago. Thus, the age spectrum for L group chondrites represents varying degrees of outgassing of radiogenic ^{40}Ar during that event. Gas retention ages in many C group chondrites are low, but this is probably because of their very fine-grained nature.

Cosmic ray exposure ages

Chondrites show a spectrum of cosmic ray exposure ages (that is, those ages reflecting the time a given chondrite existed in space as a small body prior to capture by the earth) ranging from less than 1 million years up to about 70 million years, with a pronounced mode at about 20 million years. Such ages are important in determining the source within the solar system from which chondrites are derived, but this problem is currently not well resolved (WÄNKE, 1968; WETHERILL, 1968, 1969; ANDERS and MELLICK, 1969). At present, however, some sort of asteroidal (particularly Mars- or earth-crossing asteroids) or cometary source seems the best possibility, although the moon and Mars have also been suggested (UREY, 1967, 1968; WÄNKE, 1968).

PRIMARY PROPERTIES OF CHONDRITES

In this paper, primary properties of chondrites are defined as those that are a consequence of their pre-formational history and the mechanical aspects of the formation process. Such properties can best be studied in chondrites that have undergone only very slight chemical or physical alteration during or subsequent to their formation.¹ Such chondrites are those classified as types 1, 2, and 3 by VAN SCHMUS and WOOD (1967) and commonly called "unequilibrated chondrites" or "primitive" chondrites.

Primary textures and structures

From Fig. 2 and 3 it is apparent that chondrites are aggregates of individual rounded particles, *chondrules*, packed in, or bound together by, a variable quantity of *matrix*.

Textures of types 1, 2, and 3 chondrites vary considerably. Type I chondrites (Wiik's type I carbonaceous chondrites; WIIK, 1956) do not contain chondrules, but consist wholly of fine-grained ($< 1 \mu$) oxides, sulfides, hydrated layer-lattice silicates, and other minerals (DUFRESNE and ANDERS, 1962; MASON, 1962c; BOSTROM and FREDRIKSSON, 1966; KERRIDGE, 1969); they are considered chondrites primarily because their non-volatile chemical compositions are essentially the same as those of other C group chondrites.

Type 2 (Wiik's type II carbonaceous) chondrites consist of chondrules imbedded in abundant fine-grained matrix. Fig. 2 illustrates the C2 chondrite Murray (FREDRIKSSON and KEIL, 1964), which, however, is by no means representative of all C2 chondrites; many contain chondrules that are much larger (averaging up to 2 mm diameter) and contain somewhat less matrix.

¹ "Formation" refers to the formation of chondrites as rocks and not to formation of their individual components.

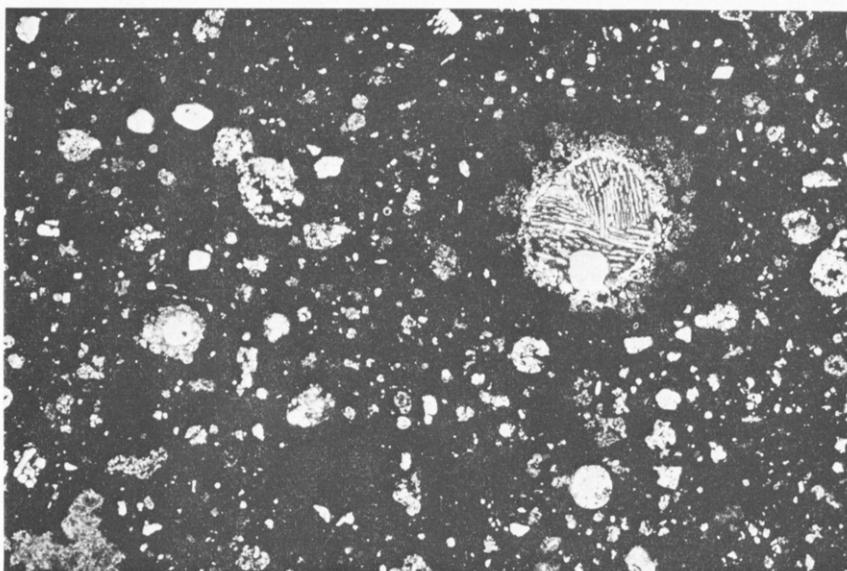


Fig.2. A thin-section view of the type C3 chondrite Murray showing chondrules (light) scattered throughout the matrix (dark). Plane polarized light; field of view 6 mm wide.

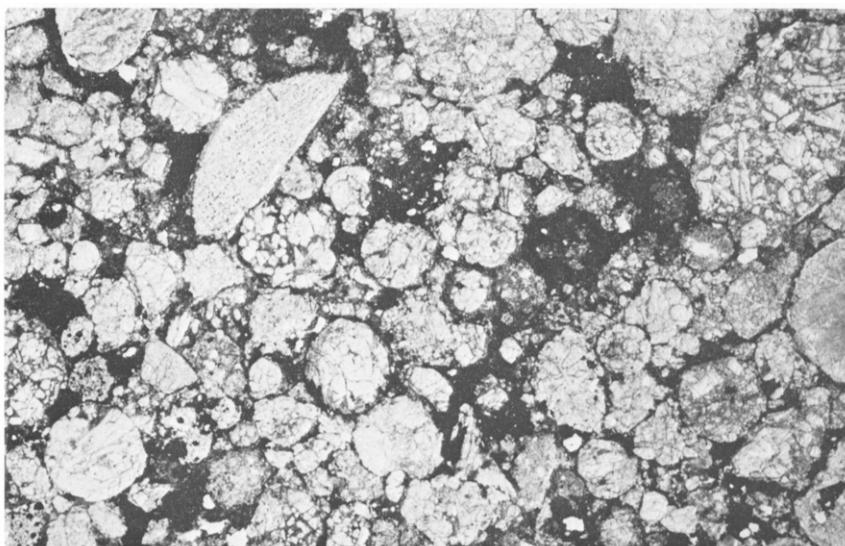


Fig.3. A thin-section view of the type L3 chondrite Hallingeberg. Note the close-packed association of the chondrules. Plane polarized light; field of view 6 mm wide.

Type 1 and 2 chondrites occur only in the C group (Table I), but type 3 chondrites occur in all five groups. Type 3 ordinary chondrites generally consist of close-packed chondrules and chondrule fragments with minor amounts of

matrix (Fig.3; DODD et al., 1967). However, C3 chondrites display two distinct textures; some consist of close-packed aggregates of small (0.1–0.4 mm diam.) chondrules with little matrix (much like type 3 ordinary chondrites) while others consist of larger (0.5–2.0 mm diam.) chondrules with substantially greater amounts of matrix (VAN SCHMUS, 1969). Textures of type 1 enstatite chondrites can also be described as close-packed aggregates of chondrules (cf., MASON, 1966, fig.1) but the presence of large quantities of opaque minerals generally obscures the relationships.

In general, chondrites appear to be an unsorted, unstructured agglomerate. However, recent studies by DODD (1965) indicate that there are definite foliations and occasional weak lineations in chondrites (as determined from elongation directions of chondrules). Dodd attributes such structure to deposition of chondrules in a gravitational field, possibly with accompanying lateral movement of the chondrules.

Many chondrites are not simply aggregates of chondrules, chondrule fragments, and matrix, but also contain angular to rounded fragments of pre-existing chondritic or other crystalline material (BINNS, 1967a, 1968; VAN SCHMUS, 1967; DODD, 1968; FREDRIKSSON, 1969). Such a chondrite (Castalia) is shown in Fig.4, and chondrites like this are commonly referred to as "polymict breccias" (WAHL, 1952) or "xenolithic" chondrites (BINNS, 1968). The fragments in Castalia range from chondritic fragments with type 2 textures through those with type 6 properties and also include fragments that may not have been derived from chondritic

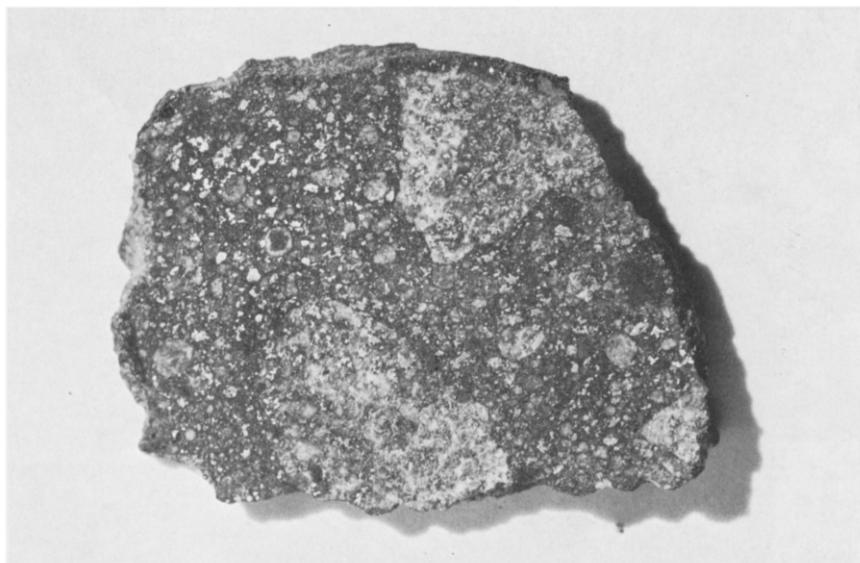


Fig.4. A cut surface on a small (2.0 cm across) fragment of the xenolithic chondrite Castalia. The angular fragments (light gray) are recrystallized (types 4 to 6) chondritic material while the host (dark) is unrecrystallized (type 3).

sources. In keeping with the concept of primary properties used in this paper, such features as polymict structure are considered primary for the Castalia chondrite (and others like it) only; clearly such features are secondary, or later, in so far as the sources of individual fragments are concerned.

Chondrules

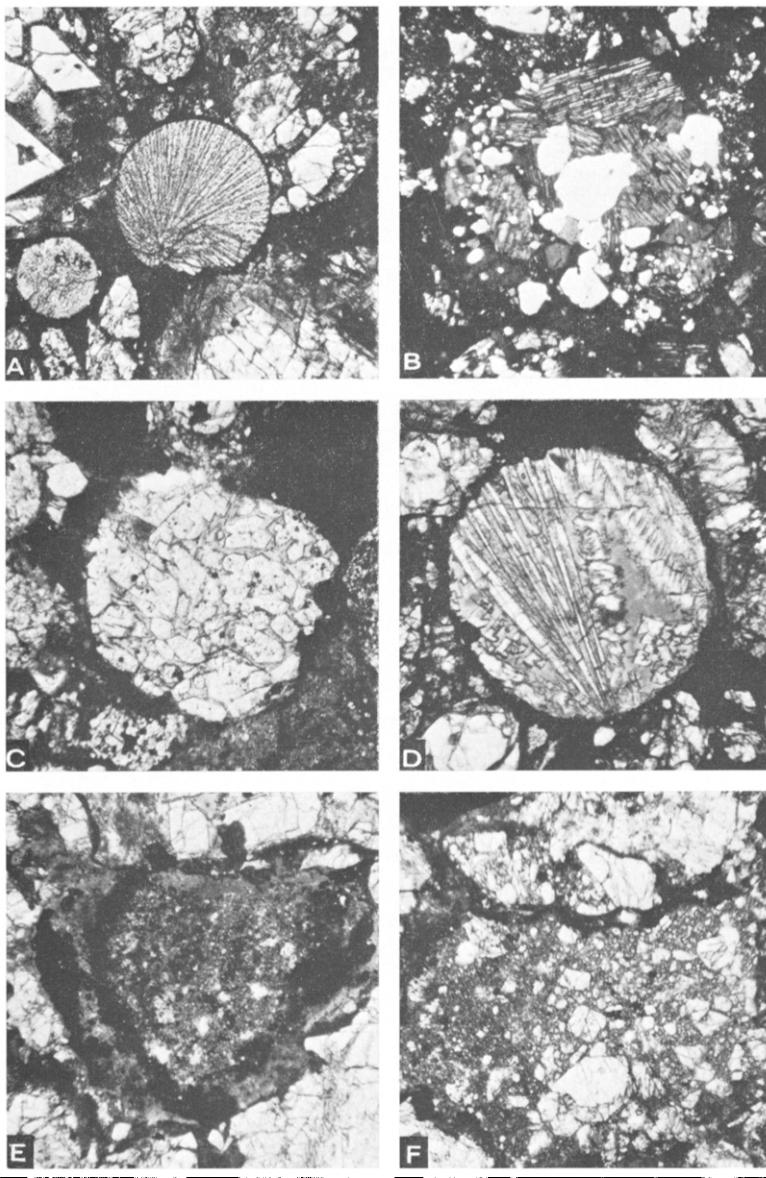
Chondrules are the principal components of chondrites and occur in a variety of shapes, sizes, internal textures and mineralogies. Generally speaking, however, they are rounded to irregular ovoids, 0.1–5 mm in diameter, and contain olivine, low-Ca (Fe, Mg) pyroxene, diopsidic to augitic pyroxene, troilite (FeS), and metallic Fe–Ni (kamacite, 5–6% Ni, and taenite, > 20% Ni) in a glassy to microcrystalline interstitial groundmass. The internal characteristics of most chondrules can be related to one of several recurring types. In addition to the following descriptions there are numerous recent papers that describe individual chondrules in detail (KEIL et al., 1964; FREDRIKSSON and REID, 1965; DODD et al., 1966; KURAT, 1967a,b, 1969; VAN SCHMUS, 1967, 1969; DODD, 1968; BINNS, 1968; FREDRIKSSON, 1969; MICHEL-LEVY, 1969) as well as several recent and older works that contain numerous illustrations and descriptions (TSCHERMAK, 1885; MERRILL, 1920, 1921; KRINOV, 1960; MASON, 1962a; WOOD, 1963a).

Barred olivine chondrules occur in chondrites of the C, H, L, and LL groups, but are not found in E group chondrites because these latter are not SiO₂ deficient. The barred olivine chondrule in Fig. 2 consists of a skeletal crystal of forsteritic olivine with interstitial microcrystalline material. In many barred olivine chondrules the skeletal structure is crystallographically a single crystal (monosomatic), while in others several crystals may be present (polysomatic). The outer rim of these chondrules consists of olivine that is crystallographically continuous with interior olivine and contains scattered to abundant inclusions of metal–sulfide droplets. The interstitial material in barred olivine chondrules is either isotropic silica-poor soda-alumina-silica glass (Table IV, analyses 2 and 3) or microcrystalline material.

Radiating pyroxene chondrules also occur commonly in chondrites (Plate IA). These chondrules consist of radiating laths of low-Ca (Mg, Fe) pyroxene with or without minor amounts of glassy or microcrystalline material between the laths, have bulk compositions apparently close to that of low-Ca Mg, Fe pyroxene, and range in size from less than 0.1 mm to over 5 mm diameter. Individual laths of pyroxene range from less than 1 micron to over 10 microns in width.

Porphyritic chondrules are probably the most common type, although several variants of “porphyritic” chondrules can be recognized. They consist of subhedral to euhedral crystals of olivine and pyroxene in glassy to microcrystalline interstitial material, but the proportions of olivine and pyroxene vary considerably so that absence of one or the other is not unusual (Plate IB, C). The pyroxene commonly is clino-enstatite to clino-hypersthene that exhibits excellent polysynthetic twinning when viewed in thin section using crossed polarized light (Plate IB). The clinopyr-

PLATE I



Thin-section views of six chondrules. Field of view approximately 0.6 mm in all cases.

- A. A radiating pyroxene chondrule.
- B. A porphyritic chondrule containing polysynthetically twinned clinobronzite (striped) and olivine (white to gray). Crossed polarized light.
- C. A porphyritic chondrule of subhedral to euhedral olivine crystals (light gray) in a clear, isotropic glass (darker gray).
- D. A glassy chondrule of skeletal olivine crystals (light) in clear, isotropic glass (darker gray).
- E. An "agglomeratic" chondrule. The dark concentric bands are caused by greater concentrations of sulfide and metallic mineral grains. The texture appears accretionary rather than igneous.
- F. A lithic chondrule or "lithic fragment". Note the very irregular outline. The crystals are predominately olivine in a microcrystalline groundmass.

oxene is sometimes mantled by high-Ca (diopsidic) pyroxene a few microns thick, but microprobe traces across interiors of such crystals show there are no Ca-rich lamellae involved in the polysynthetic twinning; the twinning is purely a structural feature. Crystals of olivine poikilitically enclosed in pyroxene are also common. In most porphyritic chondrules in type 3 chondrites, olivine and pyroxene are zoned, having more Fe-rich outer portions. Other minerals that occur in porphyritic chondrules are troilite and the metallic Fe-Ni minerals kamacite and taenite.

Glassy chondrules. A few chondrules can be found in thin sections of un-equilibrated ordinary chondrites that consist largely (sometimes almost wholly) of isotropic glass (Plate 1D). This glass is usually a pale, slightly purplish pink in color. Recent studies (FREDRIKSSON and REID, 1965; KURAT, 1967a; VAN SCHMUS, 1967) show that this glass is rich in Na-Al-Si but deficient in Fe-Mg-Ca (Table IV). The Al/Si ratios of the glasses vary considerably, but Na/Al is reasonably constant (near unity) so that normative compositions range from nepheline plus albite to quartz plus albite. FUCHS (1968) has reported the presence of nepheline X-ray diffraction lines from some such glasses. In addition to glass, these chondrules also contain skeletal, feathery, radiating, or euhedral crystals of olivine, low-Ca pyroxene, diopsidic to augitic pyroxene, or spinel, plus metal-sulfide droplet-like inclusions. FREDRIKSSON and REID (1965) and KURAT (1967a) have described such chondrules in detail.

TABLE IV

ANALYSES OF GLASSES FOUND IN CHONDRULES (%)

Component	Analysis number (and reference) ¹						
	1(a)	2(b)	3(b)	4(b)	5(c)	6(c)	7(c)
SiO ₂	64.0	50.4	63.0	72.3	54.3	57.2	55.1
Al ₂ O ₃	15.7	25.7	20.9	14.4	20.1	24.5	26.7
Na ₂ O	8.0	15.7	10.5	7.6	11.9	13.1	14.8
FeO	2.7	1.8	1.0	0.9	6.3	1.2	0.8
CaO	1.7	0.0	0.1	0.2	1.5	0.1	0.1
MgO	5.4	0.0	0.1	1.0	6.2	0.1	0.2
Other ²	2.8	n.d.	n.d.	n.d.	0.6	3.7	3.0

¹ References are:

a = FREDRIKSSON and REID (1965); b = VAN SCHMUS (1967); c = KURAT (1967a).

² Mainly K₂O, plus some Cr₂O₃, MnO, and TiO₂; n.d. = not determined.

Agglomeratic chondrules are a type of chondrule that have internal textures that suggest that they are mechanical mixtures of individual small crystal fragments of silicates, oxides, sulfides, and metal rather than a crystalline assemblage that is the result of crystallization from a melt. Commonly these chondrules have

a pronounced concentric structure such that opaque minerals are more concentrated toward the outer portions of the chondrule (Plate IE), consistent with the interpretation of these chondrules as mechanical aggregates.

Lithic chondrules or lithic fragments are terms commonly applied to chondrule-sized bodies that have internal textures and mineralogies that suggest that they are fragments of larger crystalline masses. This is supported by the fact that such chondrules often have irregular or angular outlines (Plate IF). Whether such objects should properly be called chondrules is a semantic problem and of no concern in this paper. These chondrules commonly differ from other chondrules in that the olivines and pyroxenes in them are uniform in composition, and the compositions are frequently similar to those found in recrystallized chondrites (VAN SCHMUS, 1967; KURAT, 1967a; DODD, 1968; FREDRIKSSON, 1969).

The chondrules described above are generally representative of unequilibrated ordinary chondrites, but also apply to many chondrules found in unrecrystallized E group and C group chondrites. However, there are some notable differences. Many chondrules in C group chondrites contain irregular to spherical particles of sulfide (mainly FeS) or metallic Fe-Ni profusely distributed throughout, giving them a "spongy" appearance. Many others contain calcic plagioclase as a primary mineral, although crystalline feldspar is extremely rare in type 3 ordinary or E group chondrites. In porphyritic chondrules of type I enstatite chondrites, cristobalite or tridymite commonly occurs as a primary mineral accompanying clino-enstatite (MASON, 1966; GRANT, 1968; KEIL, 1968).

Not all chondrules were incorporated into chondrites in unaltered condition. Commonly chondrules were broken prior to, or during, accretion, so that chondrule fragments are abundant in chondrites. Furthermore, some chondrules show evidence of plastic deformation, suggesting that they were involved in mutual collisions while still hot. Chondrules have also been found that show evidence of chemical alteration prior to accretion. DODD et al. (1966) described altered chondrules from the Mezö-Madaras chondrite which contain secondary fayalite co-existing with clino-enstatite, and FREDRIKSSON (1969) describes apparently altered chondrules from the Sharps chondrite.

Matrix

The chemistry and mineralogies of the fine-grained parts of type 1 and type 2 carbonaceous chondrites have been studied in detail by DU FRESNE and ANDERS (1962), MASON (1962c), BOSTROM and FREDRIKSSON (1966), NAGY (1966) and KERRIDGE (1969). The chemical composition of the matrix component is similar to that for the bulk chondrite (a necessity for type 1 chondrites!); minerals such as serpentine or chlorite, epsomite, magnetite, troilite, and bloedite have been identified, although much of the matrix appears to be amorphous. The chemical and mineralogical composition of the matrix (other than chondrule fragments) of type 3 chondrites is poorly known at present. This is because its very fine-

grained nature and its largely interstitial occurrence preclude detailed study. Studies indicate that its chemical composition is probably close to that of the bulk chondrite, although it may contain significantly greater amounts of "volatile" minor and trace elements than the chondrite as a whole. Mineralogically, magnetite (WOOD, 1962) may be present as well as some hydrous silicates of the serpentine or chlorite families, analogous to matrix in types 1 and 2 chondrites.

Primary mineralogies

Since the whole is made up of its parts, the primary mineralogies of unrecrystallized chondrites are the cumulative properties of individual chondrules plus matrix.

For unequilibrated ordinary chondrites the major primary minerals are olivine, low-Ca (Mg-Fe) pyroxene, troilite, and the Fe-Ni minerals kamacite and taenite. DODD et al. (1967) have extensively studied the properties of olivine and pyroxene in these meteorites. Olivine is generally of non-uniform composition, both with respect to single crystal compositions, crystals in individual chondrules, and the chondrite as a whole. Individual crystals are commonly zoned with Mg-rich cores and more Fe-rich rims, as would be expected from non-equilibrium crystallization of olivine from a melt. Average compositions of individual olivine crystals range from pure forsterite up to about 90 mole % fayalite; Mn is usually present at about the 1 mole % level. It is not unusual to find adjacent chondrules having grossly dissimilar mineral compositions, such as forsteritic olivine in one chondrule and fayalitic olivine in the other. Fig.5 shows the results of a random survey of olivine and pyroxene compositions from the Bishunpur chondrite. Histograms for type 3 (or unequilibrated) ordinary chondrites range from ones

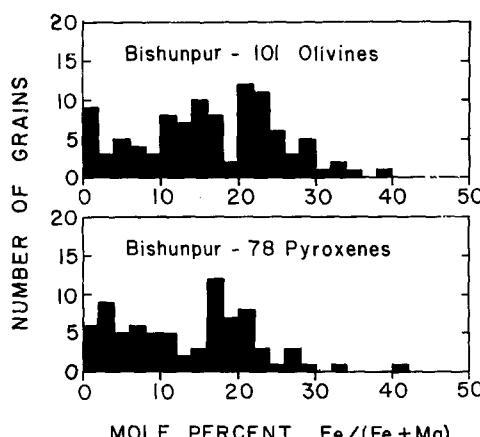


Fig.5. Histograms showing the compositions of individual randomly selected olivine and pyroxene grains in the type L3 chondrite Bishunpur (after DODD et al., 1967). Bishunpur has a texture much like that of Hallingeberg (Fig.3).

like that in Fig.6 through ones with very pronounced modes and much smaller ranges in composition. Thus there is a spectrum in the degree to which an individual type 3 chondrite represents a non-homogenous or non-equilibrium assemblage. The data for pyroxene obtained by DODD et al. (1967) are similar to that obtained for olivine, except that pyroxene is not as homogenous as olivine for those chondrites showing trends toward homogeneity.

TABLE V

MINERAL COMPOSITION DATA FOR L3 AND L4 CHONDRITES¹
(From DODD et al., 1967)

Chondrites	Olivine			Pyroxene			Type
	m.d. (%)	Fe	Fa	m.d. (%)	Fe	Fs	
Krymka	45	11.5	15.5	52	6.4	12.0	L3
Bishunpur	39	12.3	16.6	56	7.0	13.1	L3
Hallingeberg	38	14.4	19.8	57	5.7	10.6	L3
Mezö-Madaras	28	15.7	21.7	44	7.5	14.1	L3
Manych	26	17.8	24.9	40	7.6	14.3	L3
Khohar	18	17.2	24.0	36	8.2	15.5	L3
Carraweena	6.0	17.8	24.9	20	10.4	19.9	L3
Ioka	5.8	17.5	24.5	31	8.1	17.3	L3
Barratta	4.2	17.0	23.7	23	8.8	16.7	L3, 4
Goodland	2.6	17.9	25.1	14	10.0	19.1	L4
Cynthiana	xx	18.2	25.6	6.1	10.8	20.7	L4
Tennasilm	xx	16.9	23.5	6.9	10.1	19.3	L4
Bjurböle	xx	18.6	26.2	xx	10.8	20.7	L4

¹ m.d. (%) = percent mean deviation for an analysis; approximately 100 olivine grains and 50–100 pyroxene grains were analysed per sample. Fe = average iron content in weight percent. Fa = mole percent fayalite corresponding to \overline{Fe} for olivine; Fs = mole percent ferrosilite corresponding to Fe for pyroxene. Petrologic types from VAN SCHMUS and WOOD (1967). "xx" means samples were of uniform composition within analytical precision.

Table V presents the data of DODD et al. (1967) for L group chondrites and shows the continuous spectrum in degree of heterogeneity from chondrites like Bishunpur (Fig.6) to those having uniform olivine and pyroxene compositions. Although most homogenization takes place within the type 3 subdivision, it is accompanied by only slight recrystallization. VAN SCHMUS and WOOD (1967) defined the break between type 3 and type 4 chondrites primarily on the presence of optically isotropic glass in chondrules (chondrites containing at least some glass are type 3; others are type 4, or higher), and this transition occurs at about 4.0% m.d. for olivine. FREDRIKSSON and KEIL (1964), WOOD (1967a) and VAN SCHMUS (1969) have examined olivine and pyroxene compositions in C2 and C3 chondrites. Distributions of olivine compositions in C3 chondrites are similar

to that in Bishunpur, but those for C2 chondrites have their major modes at less than 4 mole percent fayalite with scattered values ranging up to about 90 mole % fayalite (Fig.6).

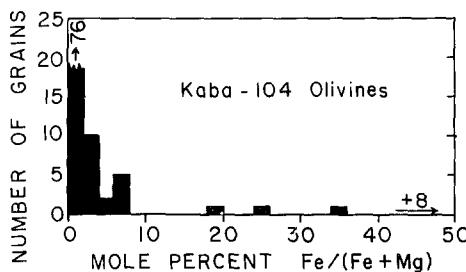


Fig.6. A histogram showing the compositions of randomly selected olivine grains in the type C2 chondrite Kaba. Note the pronounced abundance of grains with less than 4 mole % $\text{Fe}/(\text{Fe} + \text{Mg})$.

Much pyroxene in type 2 and type 3 chondrites is clino-enstatite to clino-hypersthene, and the remainder of the pyroxene is orthopyroxene, but the relative proportions of ortho- to clinopyroxene and the compositional ranges for the two structural polymorphs have not been studied in detail. POLLACK (1968) has studied the structure of chondritic pyroxene single crystals and reports that clino-enstatite to clino-hypersthene, ordered orthopyroxene, and disordered orthopyroxene (stacking disorder) are all found in type 3 chondrites; those consisting of clino-pyroxene with some disordered orthopyroxene resemble ones produced synthetically by "slow quenching" of a melt (POLLACK, 1968).

WOOD (1967a, 1967b) has studied the metallic and sulfide minerals in chondrites. His data for type 3 chondrites show that kamacite, taenite, and troilite are found and approach equilibrium relative to one another. That is, the sulfide is Ni-free FeS, as expected in the Fe-Ni-S system (KULLERUD, 1963), and the kamacite and taenite show compositional zoning of the type predicted by non-equilibrium cooling in the low-temperature portion of the Fe-Ni system (GOLDSTEIN and OGILVIE, 1966). On the basis of the taenite compositions, Wood (1967b) estimated that type 3 chondrites cooled at the rate of 0.2–2°C per million years through the temperature range 550–450°C. In type 2 chondrites, Wood found that metal and sulfide do not begin to approach equilibrium; troilite, pentlandite (20–30% Ni), kamacite, and, rarely, taenite are found. The morphology of the metal-sulfide particles (when they are not enclosed in chondrules as round droplets) can be described as "globular" and the particles do not have simple outlines.

BUNCH et al. (1967) have studied chromite compositions in chondrites and report that chromite in type 3 chondrites is commonly variable in composition, both between grains and within grains.

Since individual chondrule compositions vary considerably (WALTER, 1969), chondrules may occasionally include unusual minerals (either from the standpoint of terrestrial or chondritic mineralogy). Thus, DODD et al. (1965) discovered the mineral merrihueite $[(\text{K}, \text{Na})_2(\text{Fe}, \text{Mg})_5\text{Si}_{12}\text{O}_{30}]$ in a few isolated chondrules of the Mezö-Madaras chondrite, MICHEL-LEVY (1969) has found minerals such as spinel and melilite in C3 chondrites, BINNS (1967b) and CHRISTOPHE-MICHEL-LEVY et al. (1965) have reported cristobalite from chondrites, and the author has observed a chondrule in the Goodland chondrite that is silicon-free (based on microprobe analysis) and has a bulk composition appropriate to a chrome-bearing spinel.

In unrecrystallized enstatite chondrites the abundant primary minerals are clino-enstatite, troilite, and kamacite with minor amounts of cristobalite or tridymite, oldhamite (CaS), and daubreelite (FeCr_2S_4). In addition many other accessory minerals are found that do not occur in other chondrites and are not found in the earth's crust, primarily because of the extremely reduced nature of enstatite chondrites. A complete list will not be given here, but can be found, with references, in KEIL (1968); among the more interesting minerals are niningerite $[(\text{Mg}, \text{Fe})\text{S}]$, djerfisherite $[\text{K}_3(\text{Na}, \text{Cu})(\text{Fe}, \text{Ni})_{12}\text{S}_{14}]$, and roedderite $[(\text{Na}, \text{K})_2\text{Mg}_5\text{Si}_{12}\text{O}_{30}]$.

SECONDARY PROPERTIES OF CHONDRITES

Secondary properties of chondrites are regarded in this paper as those that are a consequence of processes acting on chondrites during or subsequent to the accretion of individual chondrules to form chondritic masses or "rocks". Such properties fall into several broad categories: (a) physical changes in internal textures of chondrites due to thermal recrystallization; (b) changes in mineral compositions; (c) changes in bulk chemical compositions; (d) physical changes resulting from impact metamorphism or brecciation; and (e) space "weathering" and terrestrial weathering. All but weathering are covered below.

The following paragraphs cover much material that is strongly contested by some workers in the field of meteoritics and reflect my own opinions and prejudices. I have deferred discussion of alternative hypotheses to the section of the paper dealing with "Origin and evolution of chondrites".

*Recrystallization*¹

If one examines a random selection of chondrite thin sections, two observations can readily be made:

¹ I will follow the lead of KEIL (1968) and regard "recrystallization" to mean "formation of new mineral grains (or new mineral boundaries—author) in a rock while in a solid state" (HOWELL, 1960).

(1) Many chondrites do not contain well-preserved chondrules, chondrule textures, or chondrule boundaries.

(2) There is a continuous spectrum in degree of recrystallization so that all gradations exist between primitive, unrecrystallized chondrites (i.e., type 3) and those with xenomorphic or xenoblastic textures in which only remnants or ghosts of chondrules may be found (Fig.7).

VAN SCHMUS and WOOD (1967) examined this spectrum in detail and

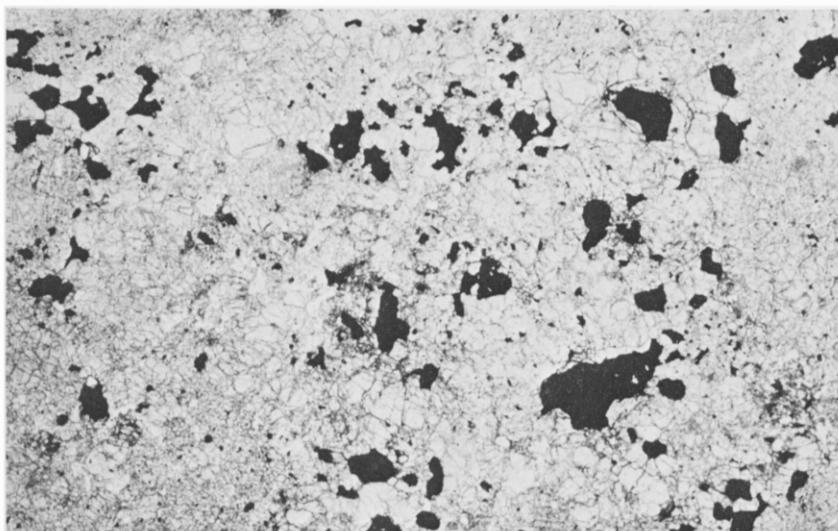


Fig.7. A thin-section view of the type L6 chondrite Kyushu showing recrystallized texture. Compare with Fig.3. The black objects are metal and sulfide grains; the rest is predominately silicate. Plane polarized light; field of view 6 mm wide.

proposed that ordinary chondrites be divided into four types: types 3 (discussed above), 4, 5, and 6. The last three types all show petrographic evidence of recrystallization, with type 4 being the least recrystallized and type 6 being the most strongly recrystallized.

In type 4 chondrites, chondrules are well preserved, both with respect to their internal textures and their boundaries. In thin section, recrystallization is readily apparent: the interchondrule matrix is gray and fine-grained (< 1 micron) rather than opaque and optically unresolvable as in type 3 chondrites, and the intercrystalline material within chondrules never occurs as an isotropic glass, as is commonly the case in type 3 chondrites, but is always a microcrystalline to fibrous aggregate that probably consists of olivine, pyroxenes, and feldspar (REID and FREDRIKSSON, 1967). In addition, the edges and faces of many crystals show feathery reaction intergrowths with adjacent matrix. The opaque minerals, troilite, kamacite, and taenite still have the irregular, globule-like structures that are found in type 3 chondrites.

Type 5 chondrites are more recrystallized and chondrule boundaries show considerable evidence of recrystallization and reaction with surrounding matrix and adjacent chondrules. The intercrystalline material in chondrules is somewhat coarser grained, and X-ray diffraction patterns of type 5 chondrites may show lines corresponding to oligoclase and diopside (MASON, 1965b). Metallic minerals and troilite commonly have irregular shapes which indicate that they have been somewhat mobilized; they are interstitial to the more refractory silicates and have lost their original globule-like morphology.

Type 6 chondrites are the most recrystallized chondrites. They have recrystallized to the point that all potential minerals can readily be recognized in thin section (Fig. 7). Individual minerals, including opaque minerals, are essentially xenomorphic, except for chromite, which is commonly subhedral to euhedral (RAMDOHR, 1967). Chondrule structures can still be found in most type 6 chondrites, particularly with the aid of crossed polarized light, but in some cases (e.g., the Shaw chondrite, FREDRIKSSON and MASON, 1967) it is difficult to find any relic chondrule structure.

Recrystallization trends in enstatite chondrites have recently been discussed by MASON (1966), and much more extensively by KEIL (1968). In general, there is a range in degree of recrystallization in these chondrites, much as in ordinary chondrites. However, because of the small number of these chondrites (Table I), the continuous nature of the phenomenon (if it exists) is not well established.

There are only two C group chondrites, at present, that are considered recrystallized. One of these, Coolidge, shows only slight evidence of recrystallization, and the other, Karoonda, is moderately recrystallized, but is so fine grained that the texture and internal structures are obscure. If there has been extensive recrystallization of C group chondrites, such chondrites have not been captured by (or recognized on) the earth.

Although the above discussion refers primarily to the work of VAN SCHMUS and WOOD (1967), it must be emphasized that their work is only a recent attempt at interpreting recrystallization textures. Recrystallization in chondrites has been recognized since the beginning of relevant petrographic observations (SORBY, 1877). Selected references to this topic are MERRILL (1921), WOOD (1962), MASON (1965b, 1966), DODD et al. (1967), REID and FREDRIKSSON (1967), KEIL (1968) and DODD (1969).

Secondary mineralogies

The mineralogies of chondrites ranging from type 3 (unrecrystallized) through type 6 (strongly recrystallized) show a steady transition from grossly non-equilibrium assemblages that characterize type 3 chondrites to assemblages that appear to closely approach those that would be expected from equilibration during high-temperature recrystallization. Table VI lists the minerals found in well-recrystallized chondrites of the E, H, L, and LL groups; the abundances given are largely based

TABLE VI

APPROXIMATE ABUNDANCES OF PRINCIPAL MINERALS IN RECRYSTALLIZED CHONDRITES

Mineral	Nominal composition	Weight percent abundance			
		E group	H group	L group	LL group
Olivine	(Mg,Fe) ₂ SiO ₄	—	33-37	45-49	56-60
Orthopyroxene	(Mg,Fe)SiO ₃	50-60	23-27	21-25	14-18
Calcic pyroxene	Ca(Mg,Fe)Si ₂ O ₆	—	4-5	4-5	4-5
Feldspar	Or ₁₋₆ An ₁₀₋₁₅ Ab ₈₁₋₈₈	5-10	9-10	9-10	9-10
Tridymite or quartz	SiO ₂	1	—	—	—
Troilite	FeS	5-10	5-6	5-6	5-6
Oldhamite	CaS	0.5	—	—	—
Ferroan alabandite	(Mn,Fe)S	0.4	—	—	—
Chromite	FeCr ₂ O ₄	—	0.5	0.5	0.5
Daubreelite	FeCr ₂ S ₄	0.6	—	—	—
Chlorapatite	Ca ₅ (PO ₄) ₃ · Cl }	—	—	—	—
Whitlockite	Ca ₃ (PO ₄) ₂ }	—	0.6	0.6	0.6
Schreibersite	(Fe,Ni)P	0.6	—	—	—
Kamacite	(Fe,Ni) (low-Ni)	15-25	15-17	6-8	1-2
Taenite	(Fe,Ni) (high-Ni)	trace	2-3	2-3	2-4

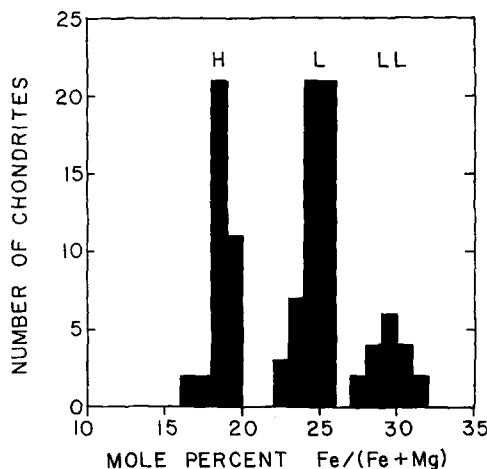


Fig. 8. Histogram showing compositions of olivine in individual chondrites. The three peaks correspond to H, L, and LL group chondrites as indicated. Note the complete gaps between groups. Data from KEIL and FREDRIKSSON (1964) and FREDRIKSSON et al. (1968).

on the author's and published norm calculations, using the actual compositions of the minerals in question, and the ranges in abundances account for uncertainties

in average compositions of chondrite groups as well as possible variations in individual chondrite bulk compositions, although the latter may be quite small.

Olivine is the most abundant single mineral in chondrites. MASON (1963) and KEIL and FREDRIKSSON (1964) have demonstrated that not only is olivine of uniform composition within most ordinary chondrites, but that olivine compositions for individual chondrites fall into a tri-modal distribution (Fig. 8) corresponding to their group affinities. As a result, olivine composition is one of the easiest ways to classify a chondrite (Table II). Microprobe analyses (KEIL and FREDRIKSSON, 1964; FREDRIKSSON et al., 1968) yield values of 16–20 mole % fayalite for the H group, 22–25 mole % for the L group, and 27–31 mole % for the LL group. Chondritic olivine also contains about 1 mole % Mn_2SiO_4 , but unlike terrestrial olivine contains very little Ni (< 100 p.p.m., FREDRIKSSON and REID, 1967). To date no systematic analytical studies have been done to determine whether there is any systematic variation in olivine composition with petrologic type, but even if there is a trend, it probably does not range over more than 1–2 mole % fayalite and may be masked by chondrite to chondrite variations.

Low-Ca pyroxene. In types 5 and 6 chondrites this pyroxene is orthorhombic, but in type 4 chondrites much of it is monoclinic and polysynthetically twinned. The ratios of Fe/(Fe + Mg) in low-Ca pyroxenes also show the tri-modal distribution characteristic of olivine, but available data do not indicate any significant variation in Fe/(Fe + Mg) ratio with type. VAN SCHMUS and KOFFMAN (1967, and unpublished data) have analysed pyroxene from type 6 chondrites of all three ordinary chondrite groups and, in addition to Mg, Fe, and Ca (Table VII), found

TABLE VII

SOME AVERAGE MINERAL COMPOSITIONS IN RECRYSTALLIZED (TYPE 6) CHONDRIES¹

Gr.	Olivine ²		Orthopyroxene ²			Ca-Pyroxene ²			Feldspar ³			Chromite ⁴		
	Fo	Fa	En	Fs	Wo	En	Fs	Wo	Or	An	Ab	Fe	Mg	Mn
E	absent		98.4	0.3	1.3	absent			4.0	15.0	81.0	absent		
H	80.7	19.3	81.6	16.8	1.6	48.8	6.0	45.2	5.8	12.3	81.9	83.1	14.0	2.9
L	74.8	25.2	77.3	20.9	1.8	47.7	7.6	44.7	5.6	10.2	84.2	87.1	10.6	2.3
LL	68.7	31.3	72.6	25.2	2.2	46.5	9.6	43.9	3.6	10.5	85.9	90.5	8.6	1.9

¹ Values in mole percents, normalized to 100% for tabulated values.² Van Schmus and Koffman (unpublished data) except for E group orthopyroxene (KEIL, 1968). Fo = forsterite, Fa = fayalite, En = enstatite, Fs = ferrosilite, and Wo = wollastonite. Olivine and orthopyroxene also contain about 0.4 wt. % Mn; Ca-pyroxene also contains about 0.5 wt. % Cr, 0.5 wt. % Na, 0.2 wt. % Al, and 0.2 wt. % Mn.³ VAN SCHMUS and RIBBE (1968) except for E group chondrites (KEIL, 1968). Or = orthoclase, An = anorthite, and Ab = albite. Or content for LL group chondrites is variable (0.9 to 5.2 mole %).⁴ BUNCH et al. (1967). Values cited are for M^{2+} cation in mole percents; M^{3+} cations are Cr (dominant), Al, Fe, Ti, and V.

about 1 mole % MnSiO_3 and traces of Na, Al, and Cr. POLLACK (1968) reported that (stacking) disordered orthopyroxene as well as ordered orthopyroxene is found in recrystallized chondrites. DUNDON and WALTER (1967) have recently applied Mossbauer spectroscopy to chondritic orthopyroxene and report that all Fe^{2+} appears to be ordered into the M_2 site (in type 3 chondrites Fe^{2+} is commonly disordered, occupying both M_1 and M_2 sites). To date no systematic studies have been made on the abundances of minor elements in ordinary chondrite pyroxene, particularly with reference to compositional changes that correlate with the petrologic types.

Calcic pyroxene (diopside) is readily observable and analyzable in type 6 chondrites, where it occurs as anhedral crystals $< 50 \mu$ in diameter. It also exists in type 4 and 5 chondrites, but is only found in the microcrystalline material interstitial to olivine and orthopyroxene crystals in partly recrystallized chondrules. VAN SCHMUS and KOFFMAN (1967, and unpublished data) have analyzed diopside from type 6 chondrites of the H, L, and LL groups (Table VII); in addition to Fe, Mg, and Ca, they found Cr (~ 0.5 wt. %), Na (~ 0.5 wt. %), Al (~ 0.3 wt. %) and Mn (~ 0.2 wt. %). The Na, Al, and Cr can be represented by the hypothetical composition $\text{Na}(\text{Al}_{0.5}\text{Cr}_{0.5})\text{Si}_2\text{O}_6$ and would account for about 4 mole % of the total diopside. As with olivine and orthopyroxene, the compositions of diopside are tri-modal, correlating with H, L, or LL groups.

Feldspar is an important constituent of chondrites, potentially making up about 10 wt. % of the meteorite. However, as noted by MASON (1965b) optically recognizable feldspar only occurs in more strongly recrystallized chondrites. VAN SCHMUS and WOOD (1967) used this relationship in defining their petrologic types (above), so that abundantly observable feldspar is confined to type 6 chondrites. VAN SCHMUS and RIBBE (1968) have recently studied chondritic feldspar by microprobe analysis (Table VII) and X-ray powder diffraction. Values of mole % An and Or for feldspar from individual chondrites within a group vary by only about $\pm \frac{1}{2}$ mole %, except for the Or content of LL group feldspar, so that feldspars from H, L, and E group chondrites are compositionally distinct from one another. Feldspar from LL group chondrites has the same An content as that from the L group, but the Or content varies by about a factor of 6 in direct response to the variability in potassium abundances in these chondrites. X-ray diffraction data and refined cell-parameters for the feldspar from type 6 chondrites yield high-intermediate to high structural states.

Maskelynite is a glass of feldspar composition commonly found in chondrites. Compositionally and texturally it is indistinguishable from crystalline feldspar (BINNS, 1967c; VAN SCHMUS and RIBBE, 1968) and all data to date indicate that it is the result of shock induced alteration of pre-existing crystalline feldspar.

Chromite in chondrites has been described in detail by BUNCH et al. (1967) and RAMDOHR (1967). The divalent ion compositions of ordinary chondritic chromites are shown in Table VII. The trivalent ions are approximately (in mole

fractions): Cr(0.78), Al(0.11), Fe(0.06), Ti(0.04), and V(0.01). Chromite compositions show systematic variations with chemical group, and the $\text{FeO}/(\text{MgO} + \text{FeO})$ ratios yield the most distinct variations: they are trimodal, analogous to olivine and pyroxene compositions. In addition, BUNCH et al. (1967) have demonstrated that chromite compositions vary with petrologic type, most noticeably with respect to TiO_2 and Al_2O_3 , which increase about three-fold in concentration from type 3 to type 6, and Cr_2O_3 , which decreases by about 4–5 wt. %.

Phosphates. There are two phosphate minerals usually found in type 6 ordinary chondrites. One is chlorapatite, $\text{Ca}_5(\text{PO}_4)_3(\text{Cl})$, with minor ($\sim 1\%$) substitution of Fe and Na for Ca and about 10% substitution of F by Cl (FUCHS, 1962, 1969; VAN SCHMUS and RIBBE, 1969). The second mineral is whitlockite¹, approximately $\text{Ca}_3(\text{PO}_4)_2$ (FRONDEL, 1941), although KEPPLER (1965) recently proposed $\text{Ca}_9(\text{PO}_4)_6\text{MgH}(\text{PO}_4)$ as the composition of whitlockite. Within current analytical uncertainty, chondritic whitlockite analyses (FUCHS, 1962, 1969; VAN SCHMUS and RIBBE, 1969) are consistent with either interpretation for the formula, with some Fe, Mg, and Na substituting for Ca (or Ca and Mg). No dependence on chemical group has been noted to date for phosphate compositions, in contrast to most other minerals. Total phosphates amount to about 0.6 wt. % for ordinary chondrites, but the relative proportion of the two minerals varies considerably (FUCHS, 1962, 1969).

Troilite is one of the major minerals in chondrites. Although no systematic study has been made, available data indicate that troilite in chondrites is stoichiometric FeS (KEIL, 1968, reports the possibility of a 1% deficiency in Fe, but the effect is within his analytical error).

Kamacite and *taenite* are the mineralogic names for $\alpha\text{-Fe}$, Ni and $\gamma\text{-Fe}$, Ni respectively, and are the stable low-temperature phases in the Fe–Ni system (GOLDSTEIN and OGILVIE, 1966). Kamacite is body-centered cubic and in chondrites contains 5–6% Ni. Taenite is face-centered cubic and contains 20–60% Ni; the range in Ni content of taenite is due to compositional zoning of this mineral, as predicted from non-equilibrium cooling in the Fe–Ni system. WOOD (1967b) has extensively studied the metallic minerals and concludes that the zoning and grain morphologies are due to in situ recrystallization and cooling processes, in contrast to UREY and MAYEDA (1959) who considered the metal particles to have been introduced as fragmentary particles. However, there is little doubt now that the in situ model is correct. Wood also concluded, on the basis of the taenite compo-

¹ SHANNON and LARSEN (1925) referred to this mineral as "merrillite" and determined its composition to be $3\text{CaO} \cdot \text{Na}_2\text{O} \cdot \text{P}_2\text{O}_5$. All subsequent studies have failed to find any such composition (naturally or experimentally) and FUCHS (1962) concluded that merrillite and whitlockite are the same mineral, Shannon and Larsen's analysis being in error. Resolution of the correct name for this mineral may soon be made by the International Mineralogical Association (merrillite has precedence; whitlockite the best description).

sitions and zoning, that recrystallized (type 6) chondrites cooled through the temperature range 550–450°C at 2–10°/million years.

Enstatite chondrites. KEIL (1968) has recently published the results of an exhaustive study on the mineralogy of enstatite chondrites, detailing the compositions of the minerals as well as compositional and mineralogical variations that parallel recrystallization. Albite (or albite glass) occurs in type I enstatite chondrites, whereas oligoclase feldspar ($An_{15}Or_4Ab_{81}$) is found in type II enstatite chondrites; sinoite (Si_2N_2O ; ANDERSON et al., 1964) is only found in type II enstatite chondrites; niningerite (Mg, FeS) is found in type I enstatite chondrites but ferroan alabandite (Mn, FeS) is found in type II; kamacite of type I contains about 3% Si, but kamacite in type II enstatite chondrites contains only about 1% Si.

Petrologic relationships

One of the most important relationships in recrystallized ordinary chondrites is the composition of coexisting olivine, orthopyroxene, and calcic pyroxene. Most of the interest in this mineral assemblage in recent years has been because of the possibility of using distribution coefficients for the Fe^{2+} – Mg^{2+} partitions between coexisting phases as geothermometers (RAMBERG and DEVORE, 1951; KRETZ, 1963) and of using the mole percent Fe^{2+} in these minerals to estimate the oxygen partial pressures that prevailed during their formation (MUELLER, 1964).

MUELLER (1964) and CRAIG (1964) attempted to use the compositions of coexisting olivine and orthopyroxene to estimate the temperatures at which these minerals formed in ordinary chondrites, and both workers obtained values of the order of 1,000–1,200°C. However, the laboratory data available to them were of insufficient quality for the interpretations they were making. Recent studies by NAFZIGER and MUAN (1967), LARIMER (1968a), and MEDARIS (1969) have demonstrated that the partition of Fe^{2+} and Mg^{2+} between coexisting olivine and orthopyroxene is not sensitive enough to temperature changes to be used as a geothermometer over the interval 900–1,300°C, and furthermore, if the distribution coefficient (K_d) is defined (RAMBERG and DEVORE, 1951) as:

$$Kd_1 = \frac{X_{Ol}^{Fe} \cdot X_{Opx}^{Mg}}{X_{Ol}^{Mg} \cdot X_{Opx}^{Fe}} \quad (1)$$

where X_j^i is the mole fraction ($Fe^{2+} + Mg^{2+} = 1.0$) of component i in mineral j , then this value is dependent on the compositions of the minerals, in contrast to the situation that should occur if the partition were ideal as defined. GROVER and ORVILLE (1969) have recently discussed the partition of Fe^{2+} and Mg^{2+} between olivine and orthopyroxene in terms of single-site (olivine) and multi-site (pyroxene) phases and concluded that geothermometry based on the Fe/Mg ratios of coexisting olivine and orthopyroxene is not feasible by their model either. Thus, the olivine–orthopyroxene geothermometer cannot be used for chondritic meteorites.

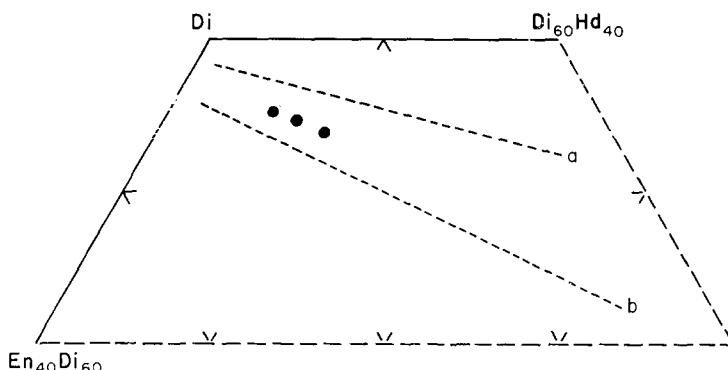


Fig.9. The diopside-rich portion of the pyroxene quadrilateral. The three black circles represent diopsidic pyroxene from ordinary chondrites (Table VII). The dashed lines indicate the compositions of metamorphic (granulite facies) pyroxenes (*a*) and igneous pyroxenes (*b*) (BINNS, 1965). Data are for calcic pyroxenes that coexist with orthopyroxenes.

BINNS (1967a) and VAN SCHMUS and KOFFMAN (1967) have used compositions of coexisting orthopyroxene and calcic pyroxene to estimate equilibration temperatures in chondrites. BINNS (1967a) plotted the composition of the calcic pyroxenes in the pyroxene quadrilateral and showed that they plotted intermediate to terrestrial igneous and metamorphic (granulite facies) pyroxenes (Fig.9). BINNS (1967a) estimated that the corresponding temperature was < 900°C.

VAN SCHMUS and KOFFMAN (1967) based their conclusions on the work of KRETZ (1963), who demonstrated that for terrestrial rocks:

$$Kd_2 = \frac{X_{\text{Ca}-\text{Px}}^{\text{Fe}} \cdot X_{\text{Opx}}^{\text{Mg}}}{X_{\text{Ca}-\text{Px}}^{\text{Mg}} \cdot X_{\text{Opx}}^{\text{Fe}}} \quad (2)$$

and the distribution coefficient is independent of composition and apparently significantly temperature dependent. VAN SCHMUS and KOFFMAN (1967) analysed the compositions of coexisting orthopyroxene and calcic pyroxene in ordinary chondrites and found Kd_2 is independent of the chondrite group (i.e., pyroxene composition), having a value of 0.61. This value for Kd_2 is intermediate to those obtained by KRETZ (1963) for igneous rocks (0.73) and metamorphic rocks (0.54). A direct correlation of Kd_2 with temperature is not possible at present, since laboratory data are lacking, but using KRETZ's (1963) estimate of Kd_2 vs. temperature, VAN SCHMUS and KOFFMAN (1967) obtained a temperature of about 800–850°C for the equilibration temperature of the system. As noted by VAN SCHMUS and KOFFMAN, the temperature of 800–850°C is not necessarily the maximum temperature attained during recrystallization, but probably represents the lowest temperature at which Fe²⁺ and Mg²⁺ ions possess sufficient mobility to effect equilibration among phases.

GROVER and ORVILLE (1969) also discussed the Fe²⁺ and Mg²⁺ partition between coexisting orthopyroxene and calcic pyroxene in terms of their multi-site

model. According to their model the data of VAN SCHMUS and KOFFMAN (1967) indicate a temperature of 600°C for the formation of the observed Fe/Mg partition (correspondingly, terrestrial metamorphic pyroxene equilibration temperatures are lowered to about 400°C). Thus, it appears at present that the bulk of the data favor the interpretation that the Fe/Mg partition among coexisting silicates in recrystallized chondrites was established at subsolidus temperatures (600–900°C).

MUELLER (1964), on the basis of thermodynamic calculations, estimated that the oxygen partial pressure needed to form olivine with ~ 0.22 mole % fayalite was ~ 10^{-16} atm. (1,000°C) to ~ 10^{-20} atm. (800°C). Recent experimental studies by NAFZIGER and MUAN (1967) and LARIMER (1968a) came to essentially the same conclusions. LARIMER (1968a) gives the following expression for the partial pressure of oxygen when olivine coexists with metallic Fe:

$$\frac{1}{2} \log p_{\text{O}_2} = \log N(\text{FeSi}_{0.5}\text{O}_2) - 14,540/T + 3.6 \quad (3)$$

where $N(\text{FeSi}_{0.5}\text{O}_2)$ is the mole % fayalite in the olivine and T is in °K. LARIMER (1968b) recently discussed the stability of minerals in enstatite chondrites and concluded that enstatite chondrites formed under extremely reducing conditions (perhaps more reducing than a nebula of solar composition).

The clino-enstatite to clino-hypersthene found in chondrules of types 2, 3, and 4 chondrites is apparently a natural consequence of crystallization and rapid cooling from a high-temperature melt. At high temperature the pyroxenes crystallize out with the proto-enstatite structure and invert upon rapid cooling to polysynthetically twinned clino-enstatite to clino-hypersthene (BOYD and ENGLAND, 1965). The absence of clino-bronzite in type 5 or 6 chondrites could be either the result of annealing by slow cooling or of reheating after agglomeration of chondrules. Unfortunately, the phase relationships between orthopyroxene and low-Ca clinopyroxene are poorly known at present (BROWN, 1968; MUÑOZ, 1968), so that any petrologic conclusions based on this transition must await further work.

The presence of tridymite in enstatite chondrites indicates that these chondrites came from a low pressure environment (MASON, 1966; GRANT, 1968; KEIL, 1968).

Secondary chemical characteristics

The chemical variations that occur among the petrologic types (e.g., the rare gases and volatile trace elements) have been outlined above. There is considerable uncertainty about a secondary (metamorphic) origin for these variations (cf., LARIMER and ANDERS, 1967); further comment on this topic is given later in this paper.

Secondary structures

The secondary structures of interest are those due to disruptive forces acting on chondritic matter. These fall into three main periods of chondritic history:

(a) brecciation processes on the surface of chondrite parent bodies prior to their disruption; (b) disruptive processes acting on parent bodies; and (c) cosmic collisions while meteorites are orbiting in space. The results of such processes are manifested in chondrites in several ways.

Veining occurs to some degree in most chondrites. The stones are cut by narrow, dark veins of glassy material that probably represents fused chondrite.

Brecciation occurs in many chondrites. However, there are at least three distinct types of brecciation, which differ in degree and may also grade into one another.

One type of brecciation is that described above under "Primary properties", where fragments of recrystallized chondritic material are enclosed in a primitive, unrecrystallized host (BINNS, 1967a, 1968; VAN SCHMUS, 1967; FREDRIKSSON, 1969). These chondrites are being recognized in ever greater numbers, so that their formation must be an integral part of any model of chondrite formation. Not all "xenolithic" chondrites have remained unaltered since their formation; at least some appear to have undergone mild recrystallization subsequent to their formation (the author has observed xenolithic fragments in the slightly recrystallized (type 4) chondrites Cynthiana and Tennasilm). The so-called gas-rich, light-dark-structured chondrites such as Pantar and Breitscheid may be "xenolithic" chondrites of this type (SUÈSS et al., 1964; R. A. BINNS, personal communication, 1968).

Most LL group chondrites are brecciated (Fig.10) but the host material

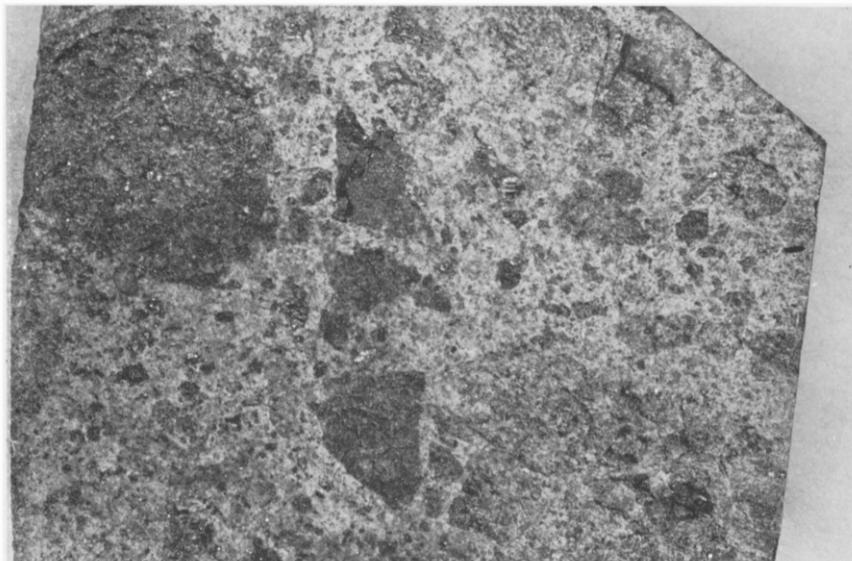


Fig.10. A cut surface on a fragment (3.0 cm across) of the Jelica chondrite, showing the brecciation typical of LL group chondrites. The dark fragments are type 6 (recrystallized) chondritic material; the lighter colored host is fine-grained granular silicate, probably finely brecciated chondritic material.

surrounding the various crystalline fragments ranges from unequilibrated material (St. Mesmin) to gray, granulated crystalline silicates that have uniform olivine compositions. In some cases both host and fragments may have uniform olivine compositions, but their values differ by 1–2 mole % Fe_2SiO_4 (FREDRIKSSON et al., 1968).

The third type of brecciated chondrites is primarily an extension of veining. In such chondrites angular fragments are surrounded by a black, glassy matrix which has apparently been derived in situ by localized fusion. Gradations exist between veined chondrites and those in which, at least locally, glassy matrix may predominate.

An important variety of L group chondrite is the "black hypersthene chondrites". These are black, rather than gray, in hand-specimen, and have apparently been altered by shock metamorphism during cosmic collisions. HEYMANN (1967) has studied them in detail and demonstrated that they were involved in a major outgassing event, probably a major cosmic collision, about 500 million years ago. Black chondrites are not restricted to the L group, however; Rose City, for example is a black H group chondrite. To date, no systematic relationship has been found for other than L group black chondrites; they may only represent isolated instances of major collisions.

In thin section several other secondary structural features can be observed. The most common is fracturing of the silicates, presumably due to shock. In addition, CARTER et al. (1968) have described deformation structures (undulatory extinction, kink banding, mosaic structure) in olivine and pyroxene grains in many chondrites and concluded that they were formed by shock waves with pressure pulses ranging up to 1 megabar. BINNS (1967c) and VAN SCHMUS and RIBBE (1968) have recently shown that the degree to which maskelynite is present correlates well with other indicators of shock alteration (deformation in other minerals, veining, low K–Ar ages).

DISCUSSION

Metamorphism of chondrites

From the standpoint of mineralogy and petrology, one of the major areas of controversy is that of the role of "metamorphism" in the origin and evolution of chondrites. This topic is important because one's concept of the primary characteristics of chondrules and chondrites is strongly dependent on the degree to which one attributes the properties of recrystallized chondrites to metamorphic alteration. DODD (1969) has extensively reviewed the overall problem of metamorphism. Only a brief discussion of the problem is presented here.

Part of the problem is that of definition. To some workers, to say that a chondrite has been metamorphosed implies that it has undergone a second stage of alteration, completely separate from any changes that may have occurred during

the accretion process. To others it implies any changes in chondrules or chondrites that occurred during *or* subsequent to the accretion process. In this paper I have referred to metamorphism in two ways. "Autometamorphism" refers to any physical, mineralogic, chemical, or textural changes that occurred *during* the accretion of chondrules and matrix material into chondritic deposits. Since chondrites contain very little water, the principal agent of "autometamorphism" would have been heat, and *autometamorphism* would have ceased upon cooling of the deposits. "Secondary metamorphism" refers to any physical, chemical, mineralogic, or textural changes that occurred as a result of heating (or reheating) *after* accretion; it is implied that chondrites were cool prior to any "secondary metamorphism". Just "metamorphism" will refer to the effects of either "auto-metamorphism" or "secondary metamorphism" (or both).

With the above terms in mind, then it is apparent that type 6 ordinary chondrites have been "metamorphosed". The major question is: what did present type 6 chondrites look like (mineralogically, petrologically, and chemically) prior to metamorphism? In the case of autometamorphism, this question refers to the characteristics of individual chondrules before accretion; in the case of secondary metamorphism it refers to the characteristics of accreted chondrules in unaltered chondrites (assuming no prior autometamorphism).

The keys to the above question are type 3 and type 4 ordinary chondrites, and two main chains of thought have developed. One theory supposes that all ordinary chondrites originally had the characteristics now found in the most unequilibrated ordinary chondrites, and the characteristics of the more homogenous type 3 ordinary chondrites and of types 4, 5, and 6 ordinary chondrites are due to "metamorphism" (DODD and VAN SCHMUS, 1965; DODD et al., 1967; VAN SCHMUS and WOOD, 1967; LARIMER, 1968a; DODD, 1969).

Adherents to the metamorphic hypothesis must accept that Fe^{2+} - Mg^{2+} ionic diffusion can be very effective at temperatures significantly lower than those required to cause appreciable textural recrystallization (in an anhydrous system), since many type 4 chondrites have homogenous (or nearly so) olivine and pyroxene compositions, yet show evidence of only slight textural recrystallization (DODD et al., 1967; VAN SCHMUS and WOOD, 1967; DODD 1969). However, many workers do not accept such a sequence of recrystallization versus changes in mineral compositions and believe instead that significant changes in mineral compositions occurred only after recrystallization was well along. Thus, most chondrules originally formed with olivine and pyroxene compositions that were uniform from chondrule to chondrule as well as within individual chondrules and the few chondrites that are "unequilibrated" are considered to have been the result of non-equilibrium perturbations of the chondrule-forming process (FREDRIKSSON, 1963, 1969; KEIL and FREDRIKSSON, 1964; KURAT, 1967b; REID and FREDRIKSSON, 1967).

For comprehensive discussions of the arguments on both sides of this controversy, the interested reader should refer to papers by FREDRIKSSON (1963, 1969); DODD and VAN SCHMUS (1965); DODD et al. (1967); KURAT (1967a, b, 1969); REID and FREDRIKSSON (1967); VAN SCHMUS and WOOD (1967); LARIMER (1968a); DODD (1969).

As postulated by DODD and VAN SCHMUS (1965), DODD et al. (1967), VAN SCHMUS and WOOD (1967), and ZÄHRINGER (1966), variations in volatile trace element abundances among types 3, 4, 5 and 6 chondrites were also a consequence of metamorphism. However, there is now much evidence that such variations may have been the result of primary variations; that is, at the time of their accretion, the parental materials of type 3, 4, 5 and 6 chondrites had systematically differing amounts of volatile elements (MARTI, 1967; LARIMER and ANDERS, 1967; TANDON and WASSON, 1968). The fact that major element chemical variations exist between type I and type II enstatite chondrites (particularly total Fe and the Si/Mg ratio) strongly support this concept (KEIL, 1968). LARIMER and ANDERS (1967) have extensively developed this alternative in the context of chondrule production and accretion in a cooling solar nebula.

Origin and evolution of chondrites

The following paragraphs briefly outline two of many possible models for the origin and evolution of chondrites. Each model is largely based on ideas already in the literature, but is modified to a greater or lesser degree to suit the author as a reasonable alternative. Neither model is free from serious problems, and only future research can narrow possible interpretations.

Two assumptions have been made in outlining the models presented below. First, it is assumed that each chemical group represents a separate evolutionary environment for chondrites although similar processes prevailed. Therefore, the models below can apply to any group by suitably specifying overall compositions (including Po_2). Second, it is assumed that Fe^{2+} - Mg^{2+} ionic diffusion was able to effectively homogenize olivine and pyroxene compositions before extensive recrystallization occurred. I believe, at present, that this is a reasonable assumption, particularly since the petrologic system under consideration was essentially anhydrous. This assumption removes the need for postulating chondrule forming mechanisms that resulted in chondrules with completely homogenous olivine and pyroxene compositions.

Model 1. In this model it is assumed that chondrules were produced directly from the solar nebula. Although models have been proposed for direct condensation of chondrules from the solar nebula (e.g., WOOD, 1963b) it is now generally conceded that it was not possible to directly condense molten droplets, particularly ones containing significant amounts of ferrous iron, from the solar nebula. The most attractive mechanism at present is that of WHIPPLE (1966) in which he proposed

that lightning-like discharges occurred in the solar nebula, and such discharges were able to fuse nebular dust into molten droplets. As these droplets cooled, the typical textures and mineralogies of chondrule resulted.

Following (or accompanying) chondrule production, chondrules plus variable amounts of dust accreted to form one or more parent bodies (per group). Such parent bodies were initially chondritic throughout, and the outer portions may have contained fewer chondrules and more dust (matrix). The chondritic deposits in interior portions of these parent bodies underwent varying degrees of metamorphism, and the degree of recrystallization increased with depth. A stage may have been reached at which segregation of metal and silicate resulted in achondritic and metallic meteorites. The heat that caused the metamorphism could have been generated during accretion in some way, resulting in "autometamorphism", or the bodies could have initially been cool and been heated in their interiors as a result of the decay of short-lived radio-nuclides (FISH et al., 1960).

Eventually the parent body cooled, and at some later time was broken up.

Model 2. In this model chondrules were formed on the surface of a pre-existing body (or bodies). Actual production of molten droplets could have been the result of a cosmic impact (chondrules = splash drops) or as the result of explosive release of a molten silicate magma (FREDRIKSSON, 1963, 1969; FREDRIKSSON and RINGWOOD, 1963; RINGWOOD, 1966; REID and FREDRIKSSON, 1967; UREY, 1967). In either case, chondrules and other debris accumulated on the surface of the body as local deposits. Thus, each of the five groups could have come from a single parent body.

Parts of the deposits were metamorphosed, probably through "autometamorphism" as a result of residual heat remaining from the formation process, but upper portions of the deposits remained cool and unrecrystallized. At some later time, break-up of the parent body occurred.

General comments. The following comments apply in a general sense to both models.

First, the mineralogies and textures of types 4, 5, and 6 chondrites were a direct consequence of whatever metamorphism occurred. Mineralogically and texturally, chondrules originally resembled those now found in the unequilibrated chondrites (provided, of course, that differences among groups did exist—for example, enstatite chondrites versus H group chondrites).

Second, the chemical variations that exist *within* groups (major, minor, or trace elements) very likely are a consequence of the accretion process itself, that is, the material that was accreting to form chondrites was also undergoing changes in bulk chemical compositions as the accretion process proceeded (e.g., LARIMER and ANDERS, 1967; KEIL, 1968). However, it is unlikely that chondrites remained totally closed systems during metamorphism, so that at least part of the chemical

variations within groups could be due to losses (or gains) of particular elements.

Third, the numerous instances of recrystallized chondritic fragments enclosed within more primitive chondritic host material ("xenolithic chondrites") require that catastrophic forces (either cosmic impacts or volcanic explosions) were able to eject recrystallized material into areas where chondrule deposits were still unindurated and unrecrystallized (I do not believe the chondritic fragments as such existed in space prior to a second cycle of accretion, but such could have been the case).

Finally, the other features in chondrites such as veining, blackening, cataclastic brecciation, and deformation of minerals can easily be accounted for by collisional processes prior to the capture of the chondrites by the earth.

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