
The Mineralogy of Iron Meteorites

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The mineralogy of iron meteorites

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[Plates 1 and 2]

The status of research on the mineralogy of meteorites about 1876 is compared with that of 1976. The ratio of iron meteorites to other meteorites in collections was then, and is now, about 1:2. However, total accumulated mass of iron meteorites exceeds that of stones by a factor 15.

Forty-one well-established minerals in iron meteorites are discussed utilizing twelve tables, presenting the elements, carbides, nitrides, phosphides, sulphides, oxides, phosphates and silicates.

The detailed compositions of kamacite, taenite and troilite are discussed, and the ranges in composition of several other minerals are given. Unequilibrated phases, such as α_2 -iron, and shock-transformation products, such as diamond, are briefly mentioned.

Finally, the products of terrestrial weathering are discussed, and it is shown that lawrencite in iron meteorites is a dubious phenomenon and not a cosmic mineral.

INTRODUCTION

With almost perfect timing the initiation a century ago of the Mineralogical Society was noted by celestial forces: the only iron meteorite reported from England was seen to fall near Rowton, Salop, on 20 April 1876 (Flight 1882). Perhaps it was this remarkable event that inspired the members of the Mineralogical Society to give meteorites a prominent place in its proceedings. Fletcher, Prior, Spencer, Hey, Agrell, Axon and others have all made very important contributions to meteorites, and the *Mineralogical Magazine* has always provided space for their publications.

The present review paper begins with a comparison between the status of the mineralogy of irons in 1876, and that of today, at the centenary of the Society.

METEORITE STUDIES ABOUT 1876

Centres of study were at the university museums in Vienna (Tschermak), Berlin (Rose), and Paris (Daubrée), while secondary centres were located in London (Story-Maskelyne), Amherst and Yale (Shepard), and Harvard (J. Lawrence Smith). Material was not overly plentiful, since only about 800 meteorites had been accumulated in the most important collections (table 1). Of these, two-thirds were stones, one-third irons and stony irons. Tschermak had just improved Rose's classification of the entire meteorite domain (table 2) and the result was so successful that it survived for three generations and is still partly with us, improved and augmented by Tschermak (1885), Brezina (1896), Prior (1920) and Mason (1962).

The terms octahedrite and hexahedrite were introduced by Tschermak; subdivisions according to kamacite bandwidth were called Of, Om and Og, for German fine-, medium- and coarse-

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structured bands. The system was thus purely descriptive, based upon significant structural elements and upon the major minerals.

The science of meteoritics dates from about 1800, with the works by Chladni (1794, 1819), Howard (1802), Biot (1803), Klaproth (1803, 1807) and Schreibers (1813, 1820) as the foundation. Up to 1876 only a score of minerals had been identified and named (tables 3, 4), but of these only about half proved to be valid (table 3). These ten minerals are the most important ones in iron meteorites; they occur as major phases or as conspicuous inclusions. Knowledge of the silicate minerals came in 1876 mainly from studies of Toluca, Netschaev and the two Tucson iron meteorite masses.

TABLE 1. NUMBERS OF METEORITES, INCLUDING PSEUDOMETEORITES,† ABOUT 1876

	stones	stony-irons	irons	total
Mineralogisches Museum, Berlin (Rose 1864)	109	12	60	181
K. & K. Hofmuseum, Vienna (Tschermak 1872)	182	10	91	283
British Museum, London (Story-Maskelyne 1877)	202	12	107	321

† A study of the museum-catalogues indicates that about 10% of the listed meteorites were indeed pseudo-meteorites or duplicates.

TABLE 2. CLASSIFICATION BY ROSE (1864) AND TSCHERMAK (1872)

characteristic minerals	class, name, and author
I anorthite, augite	eukrite (R)
II olivine, bronzite, enstatite	shalkite (R) → amphoterite (T) manegaumite (R) → diogenite (T) howardite (R)
III olivine, bronzite, iron	chondrite (R)
IV silicates and iron, granular	mesosiderite (R)
V silicates and iron, porphyritic	pallasite (R)
VI iron	octahedrite (T) hexahedrite (T) dichte Eisen (T) → ataxite (Brezina)

TABLE 3. MINERALS AND METALLIC PHASES IN METEORITES, 1876

	crystal system	composition	source	meteorite example
kamacite	b.c.c.	Fe-Ni	Reichenbach 1861	Braunau
taenite	f.c.c.	Fe-Ni	Reichenbach 1861	Elbogen
graphite	hexagonal	C	Troost 1840	Magura
schreibersite	tetragonal	(Fe, Ni) ₃ P	Haidinger 1847, 1861	Lenarto
troilite	hexagonal	FeS	Haidinger 1863	Toluca
chromite	cubic	FeCr ₂ O ₄	Rose 1864	Carthage
daubréelite	cubic	FeCr ₃ S ₄	J. L. Smith 1876	Coahuila
olivine	rhombic	(Fe, Mg) ₂ SiO ₄	J. L. Smith 1855	Krasnojarsk
enstatite	rhombic	(Fe, Mg)SiO ₃	Shepard 1854	Vaca Muerta
plagioclase	triclinic	labradorite	Genth 1855	Tucson

The means of identification were mainly wet-chemical analysis, densimetry and goniometry. Because of the limitations of precision and of qualitative analysis (e.g. of nickel), and because some pairs of minerals occur as fine intergrowths (e.g. troilite-metal, cohenite-schreibersite) and ore microscopy was only just starting up with Sorby (1864) and Rose (1864) as pioneers, it is understandable that the list of 'minerals' which later generations discredited is rather long (table 4). Also at that date new mineral names were never scrutinized as they are today,

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wherefore some appeared in print only to become rejected a few years later. Others, such as pyrite, survived among less knowledgeable workers, until Cohen (1894, 1903) in a definitive work showed a number of inconsistencies.

TABLE 4. MINERALS WITH AN UNCERTAIN POSITION IN 1876, NOW DISCREDITED

	reported by:	discredited by:	misinterpretation of:
chladnite	Shepard 1846	Kenngott 1862	enstatite
dyslytite	Shepard 1846	Cohen 1894	schreibersite
lamprite	Reichenbach 1861	Haidinger 1861; Rose 1864	cohenite and schreibersite
partschite	Shepard 1853	Cohen 1894	rhabdite
pyrite	Bournon 1802	Proust 1805; Berzelius 1834	troilite
pyrrhotite	Stromeyer 1812; Buddhue 1945	Rammelsberg 1848	troilite
quartz	Joy 1853; Rose 1861	Cohen 1894	contamination
rhabdite	Rose 1864	Tschermak 1872, 1874	schreibersite
shepardite I	Haidinger 1847	Rose 1864	daubréelite
shepardite II	Rose 1864	Cohen 1894	enstatite

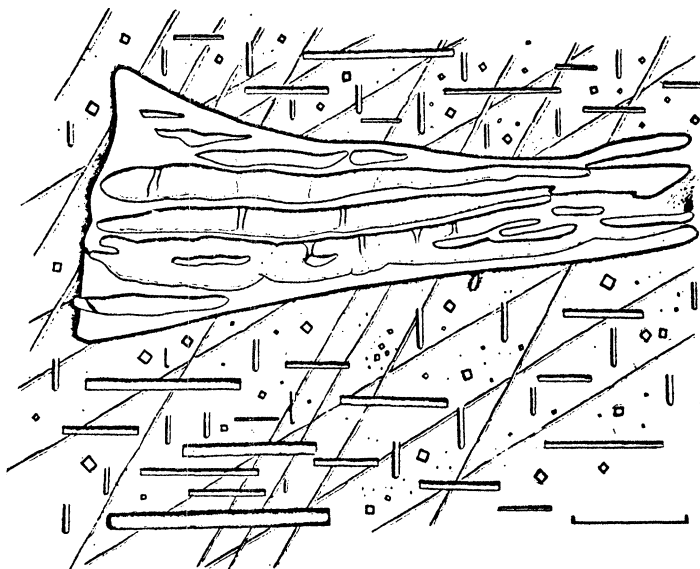


FIGURE 1. The iron meteorite Seeläsgen. One of Rose's original drawings when he first described rhabdite (1864, Plate 1). Above, an open-meshed plesite field; in the kamacite numerous oriented rhabdite prisms and Neumann bands in two directions. 'Hausenblasenabdruck', i.e. gelatine imprint. Scale bar 100 μm .

One of the terms of table 4, 'rhabdite', has survived and is now widely accepted as covering the prismatic and plate-shaped forms of schreibersite, that are particularly common in hexahedrites. Figure 1 is an interesting drawing by Rose who probably was the first to examine gelatine imprints removed from an etched surface and placed under the microscope. It was remarkable how many details he could observe. By applying this technique he anticipated by three generations the electronmicroscopic replica technique.

The proposition by Reichenbach (1861) that the brilliant, hard mineral in Lenarto, Magura, Bohumilitz, etc. should be called lamprite (table 4) was not accepted by Haidinger (1861) and Rose (1864, p. 54). The opposition was partly due to mix-up and imperfect descriptions and partly to personal confrontations, particularly between Haidinger (Vienna) and Reichenbach. In the long run it led to the result that Reichenbach's very important private meteorite

collection, developed at his castle just outside Vienna, was not bequeathed to Vienna, but to *Tübingen* in Württemberg, about 1876 (Berwerth 1913). This initiated meteoritics at Tübingen, now appropriately extended to include von Engelhardt's & Stöffler's work on the Ries Kessel crater problems.

Quartz, although discredited in *iron* meteorites, was later proved to exist as small, accessory grains in some enstatite chondrites (Abee and St. Marks) and eucrites (Jonzac, Peramiho and Stannern).

TABLE 5. CLASSIFICATION OF IRON METEORITES

chemical group	structural class	falls	finds	examples
I	coarse inclusion-rich octahedrites	5	69	Cañon Diablo
I-An	inclusion-rich irons	2	19	Copiapo
IIA	hexahedrites	4	43	Coahuila
IIB	coarsest octahedrites	1	14	Sikhote-Alin
IIC	plestitic octahedrites	0	7	Ballinoo
IID	medium octahedrites (10–11.5% Ni)	2	11	Carbo
IIIA	medium octahedrites (7–8.8% Ni)	4	117	Cape York
IIIB	medium octahedrites (8.6–10.6% Ni)	1	40	Chupaderos
IIIC	fine octahedrites (10.5–13.5% Ni)	1	6	Mungindi
IIID	finest octahedrites	0	5	Tazewell
IIIE	coarse octahedrites	0	7	Kokstad
IVA	fine octahedrites (7.5–10% Ni)	1	39	Gibeon
IVB	ataxites	0	11	Hoba
An	anomalous iron meteorites	4	92	Nedagolla
	unclassified iron meteorites	7	52	Armanty
	total	32	532	

CLASSIFICATION AND STATISTICS, 1976

Let us now move rapidly to our time. The classification of iron meteorites is now founded upon chemical composition rather than structure, an approach which reflects our increasing interest in the problem of how and where meteorites formed. The iron meteorites are subdivided into 13 chemical groups according to their nickel and phosphorus contents, and to their content of trace elements (table 5). While it has been shown by Scott (1972) that there are a number of elements the abundance of which could just as well be used as a basis for a chemical classification system, experience has shown that the initially proposed abundances of the elements nickel and gallium (Goldberg *et al.* 1951), germanium (Lovering *et al.* 1957; Wasson 1967) and iridium (Wasson & Kimberlin 1967) together constitute a very powerful set of parameters for this purpose. The four roman numerals refer to decreasing Ga- and Ge-levels, roughly from 100 parts/ 10^6 Ga (I), through 50 parts/ 10^6 Ga (II) and 20 parts/ 10^6 Ga (III) to less than 1 part/ 10^6 Ga (IV). A number of elements show systematic positive or negative correlations with Ni and Ga (Scott 1972; Wasson 1974; Buchwald 1975*a*).

The classification scheme based upon chemical principles has not made obsolete the previous schemes based upon structure. On the whole, it turns out that there is a very good correlation between the two approaches, particularly if, in addition to the macrostructure, one also observes and utilizes the microstructure. It is therefore proposed (Buchwald 1975*a*) that any iron meteorite should be introduced and classified with a 'fingerprint', consisting of a structural specification, such as 'medium octahedrite with a kamacite bandwidth of (0.9 ± 0.1) mm', plus a bulk chemical analysis, reporting at least Ni and P, but also, if possible, Ga, Ge and Ir. Since these three elements usually are present on the parts per million level

only, it is recommended that the analytical work on their determination be carried out within a few experienced laboratories in order to save valuable material for future investigations.

The frequency of distribution of nickel content in selected analyses of iron meteorites is shown in figure 2. No iron meteorite contains less than about 5 % Ni and only nine contain more than 20 %. The rather confusing outline of the figure – resembling the Manhattan skyline – is analysed in figure 3. It is obvious that the ragged appearance is due to the histogram being a sum curve of several independent populations, each of which displays a smoother distribution. The sum histogram, figure 2, is therefore in itself of little meaning.

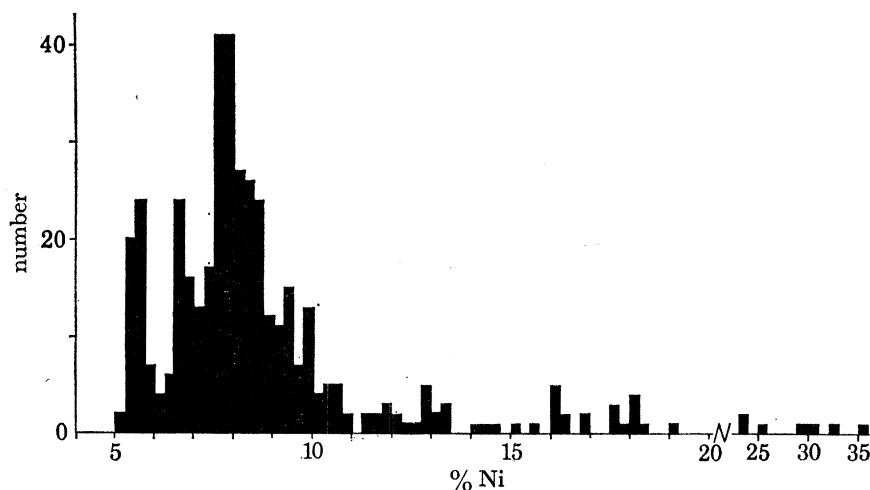


FIGURE 2. A histogram of nickel content in iron meteorites, plotted at 0.25 % Ni intervals. Outside the histogram lie Dermbach (42%) and Oktibbeha County (61%). This type of diagram was first used by Yavnel (1958), but it is here refined and updated using selected analyses. (From Buchwald 1975*a*.)

The nickel-poor members of each group are the most common, both in terms of number and cumulative weight. Thus in group I, the huge meteorites Magura, Campo del Cielo, Morasko, Youndegin, Cranbourne and Cañon Diablo all have less than 7.1 % Ni. In group IIIA–IIIB, the crater-producing Wabar, Henbury and Boxhole meteorites and the monsters Morito, Willamette and Cape York all have less than 7.9 % Ni. In group IVA, the only really large meteorite, Gibeon, with 7.9 % Ni, is also situated near the lower limit of the group. It appears, from this histogram and from various other lines of evidence, that the ‘average’ nickel content in iron meteorites is close to 7.9 %.

Iron meteorites are rare. New ones arrive and are recognized with an average frequency of one every seventh year. The newest that we know of is the 25 kg Portuguese meteorite Juro-menha, which fell at 17h55 on the 14 November 1968. This is chemically an average meteorite, transitional between group IIIA and IIIB, but structurally it is unique, showing secondary phase transformations that have altered a pre-existing medium Widmanstätten structure into a complex and partly unexplained fine microstructure, figure 4, plate 1.

The information in table 6 has been broken down into 20-year periods. If we look at numbers, the activity was at a maximum about the time of the First World War; but if we look at the accumulated weight, the Sikhote-Alin fall in 1947 totally dominates all other meteorites. If anything, these facts serve to indicate how imperfect our statistical background is; unfortunately it is not likely to improve very much, even if significant programmes have been

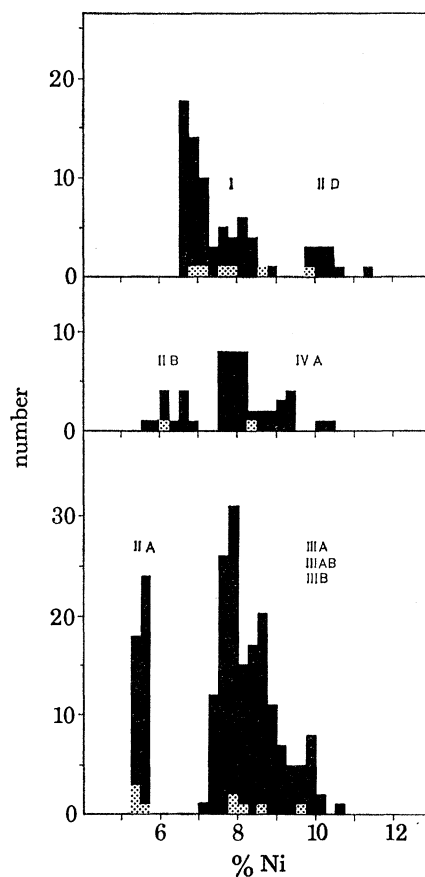


FIGURE 3. Analysis of figure 2. The sharp peak at 5.5% Ni is caused by the hexahedrites of group II A; the peak at 6.8% Ni by the coarse octahedrites of group I; and the peak at 8.0% Ni by the combined effects of group III A and IV A. Grey squares indicate observed falls. (From Buchwald 1975*a*.)

DESCRIPTION OF PLATE 1

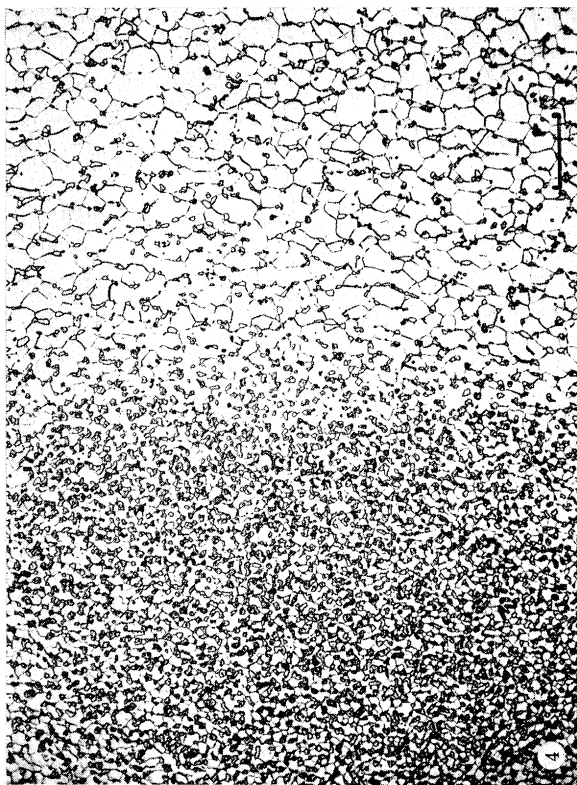
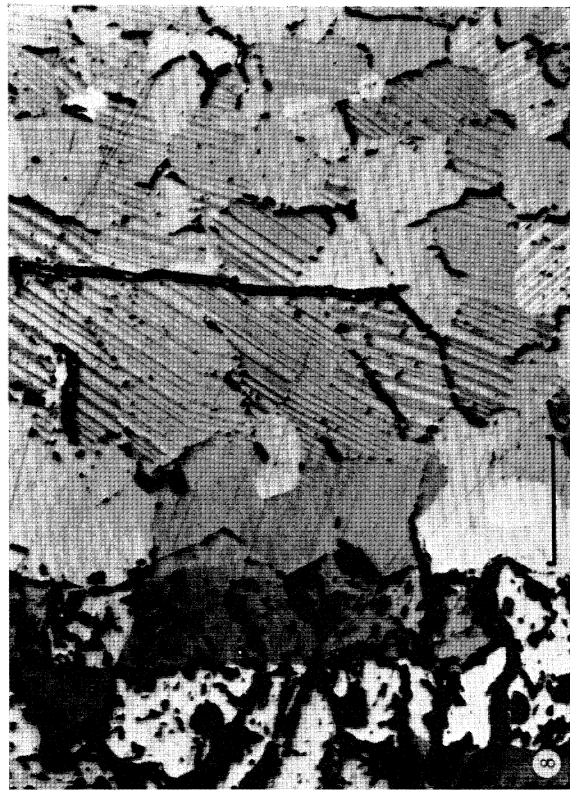
FIGURE 4. Section through the iron meteorite Juromenha, which fell 14 November 1968 (U.S.N.M. No. 3534).

The meteorite apparently represents an extreme case of a cosmically reheated structure. What was previously kamacite lamellae (right) and taenite-plessite regions (left) were entirely decomposed and spheroidized. The particles in the kamacite grain boundaries are of two kinds: angular phosphides and irregular, unequilibrated high-nickel phases. Polished and etched. Scale bar 100 μm .

FIGURE 6. The iron meteorite Dungannon (U.S.N.M. No. 644). Softly rounded cohenite crystals once occupied most of the section. A thorough, late annealing decomposed the cohenite to granular kamacite (ferrite with about 1.5% Ni) and lamellar graphite (black). Simultaneously the kamacite bands recrystallized and the plessite field was annealed (large black wedge). Etched. Scale bar 400 μm . (From Buchwald 1975*a*.)

FIGURE 7. The iron meteorite Cañon Diablo. (Inst. of Meteoritics, Albuquerque, New Mexico). A section through one of the severely distorted fragments from the vicinity of Meteor Crater, Arizona. Original cohenite (centre, white) is partly decomposed. The characteristic reaction zones contain more nickel and less carbon as one moves away from the cohenite crystal. Etched. Scale bar 100 μm . (From Buchwald 1975*a*.)

FIGURE 8. An artificially altered troilite-daubrélite nodule from the iron meteorite Yanhuitlan (U.S.N.M. No. 459B). Reheating to about 700–800 °C by Mexican workmen altered the kamacite to distorted α_2 (not shown), while troilite (left) and daubrélite (right) recrystallized and, in addition, decomposed into a lamellar aggregate. Polished. Crossed polars. Scale bar 100 μm .



FIGURES 4 AND 6-8. For description see opposite.

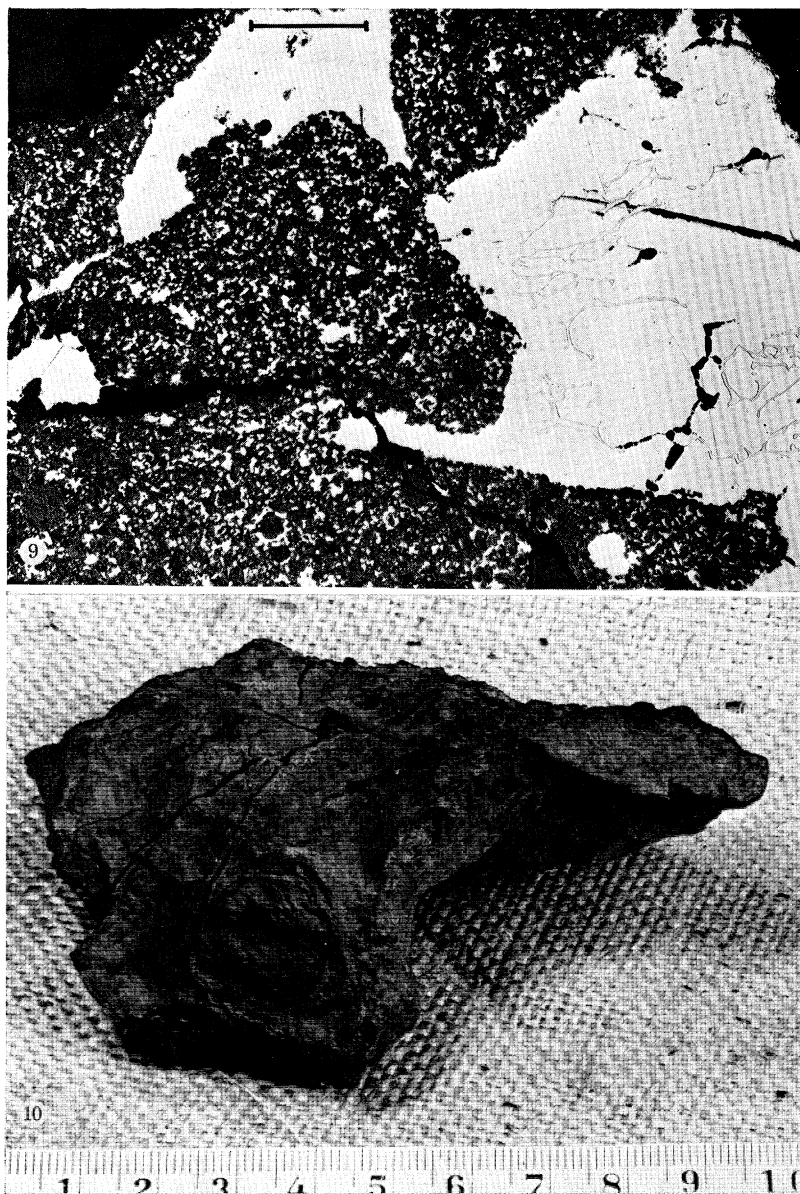


FIGURE 9. Netschaevo, Russia (Brit. Mus. No. 33, 953). An anomalous medium octahedrite with angular silicate inclusions. The inclusions are chondritic and display nickel-iron, olivine and enstatite as major, and plagioclase and chromite as minor minerals. The bulk metal has 8.46% Ni and Ga-Ge-Ir contents that place Netschaevo outside the chemical groups. Scale bar 3 mm.

FIGURE 10. A corroded surface fragment (390 g) of the medium octahedrite Cape York (Agpalilik), Greenland (Copenhagen No. 1967, 410). While the upper part of the 20-ton mass was well preserved, such parts that had been in contact with the soil were corroded. The average density of this sample, when taken in 1963, was 5.0, but on removal to my office in Copenhagen it rapidly disintegrated and was after 12 months nothing but a loose heap of scale and dust. Scale in centimetres.

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initiated. Such programmes as the Prairie Network in the United States and similar ones in Czechoslovakia and Canada (McCrosky & Ceplecha 1969), aimed at monitoring the sky over long periods with automatic cameras, will no doubt record whatever material happens to pass in front of the cameras, and a concerted search-effort may retrieve the fallen material. Basically, however, these programmes were initiated to provide much needed quantitative data about the physics of the fall and from these, recent orbital parameters may be derived. More hope of recovery of new meteorites must be attached to the ever-increasing general interest in the sky and to the improved exchange of information on all levels. Improved information-exchange with large and densely populated areas such as China, India and Indonesia might be most rewarding.

TABLE 6. THE 32 WITNESSED FALLS OF IRON METEORITES

	names	recovered mass/kg
1740–1859	Hraschina, Charlotte, Braunau	93
1860–1879	Nedagolla, Rowton	8
1880–1899	Mazapil, Hassi-Jekna, Cabin Creek, Quesa, Magnesia	69
1900–1919	N'Goureyma, Okano, Avče, N'Kandhla, Annaheim, Treysa, Boguslavka, Garhi Yasin, Norfolk, Rembang	406
1920–1939	Pitts, Samelia, Udei Station, Repeev Khutor, Bahjoi, Nyaung	133
1940–1959	Soroti, Sikhote-Alin, Yardymly	23,000
1960–1976	Bogou, Muzzaffarpur, Juromenha	35

TABLE 7. DISCREDITED FALLS OF IRON METEORITES (BUCHWALD 1975 *a*)

Elbogen, Gundaring, Helt Township,
Kirkland, Majorca, Mariaville,
Para de Minas, Patos de Minas, Puerta de Arauco,
Tandil, Victoria West and Winburg.

At present, the total number of known and preserved meteorites is about 1750. About two-thirds of these are stone meteorites. On a mass basis, however, the iron meteorites (460 tonne) outweigh all other categories (total: 30 t) by a factor of fifteen. If, in addition, it is taken into consideration that the crater-producing bodies, which we so far have been able to identify, are all iron meteorites (Buchwald 1975 *b*), it will be understood that the flux of meteoric debris penetrating to Earth is predominantly of metallic composition.

The ratio 1:2 of known iron to stone meteorites is also reflected in the large collections in the Smithsonian Institution (Washington), Arizona State University (Tempe), American Museum of Natural History (New York), Field Museum (Chicago), British Museum (London), Humboldt-Universität (Berlin), Musée d'Histoire Naturelle (Paris) and Geological Museum (Copenhagen). Roughly one-third of all specimens are iron meteorites and the largest specimens on display are all irons, notably the Cape York masses of 30.9 t (New York) and 20.1 t (Copenhagen).

In some collections and in textbooks on meteorites a number of irons are still erroneously treated as genuine meteorites or as actually observed falls. However, a close examination of problematic material listed in table 7 has revealed that these meteorites are much too corroded to be accepted as truly observed falls. Furthermore, Kirkland was a deliberate fraud and Tandil has turned out to be a piece of iron ore!

Having thus sorted the material and selected the genuine undamaged meteorites, let us turn to the individual minerals. In tables 8–20 the currently known metallic phases, minerals and problematic occurrences are presented. For the most important ones, such as kamacite, taenite and troilite, information on the detailed composition is presented while the remaining ones are only cursorily described. For further information the reader is referred to recent works by Bunch *et al.* (1970, 1972), Mason (1971, 1972*a*), Ramdohr (1973, 1975), Wasson (1974) and Buchwald (1975*a*).

TABLE 8. MINERALS IN IRON METEORITES. I: ELEMENTS AND CARBIDES

	composition	crystal system	source	meteorite example
kamacite	Fe–Ni, Ni < 7.5 %	b.c.c. α -iron	Reichenbach 1861	Cape York
taenite*	Fe–Ni, Ni > 30 %	f.c.c. γ -iron	Reichenbach 1861	Twin City
graphite	C	hexagonal	Troost 1840	Mundrabilla
copper	Cu	cubic	Nininger 1941	Morito
cohenite	(Fe, Ni) ₃ C	rhombic	Weinschenk 1889	Smithville
haxonite*	(Fe, Ni) ₂₃ C ₆	cubic	Scott 1971	Rhine Villa

Note: An asterisk * in this and the following tables indicates that the mineral is unknown as a terrestrial mineral. Kamacite and cohenite do occur on Earth, but are extremely rare, having been reported only from Ovifak, Greenland, Bühl, Germany, and a few other places.

ELEMENTS AND CARBIDES

Kamacite, taenite and transformation products of these

The major minerals are the two metallic phases, kamacite and taenite, table 8. When annealed and equilibrated, they are soft iron–nickel alloys with respectively 5.4–7.5 % Ni (Vickers hardness number (100 g) 145–160) and more than 30 % Ni (150–165 V.h.n.). Kamacite may occur as foot-sized crystals in hexahedrites such as Coahuila, or as millimetre-sized crystals in recrystallized hexahedrites and octahedrites such as Bingera and Joel's Iron, or in octahedrites as extended platelets or ribbons, the Widmanstätten bands proper, whose thicknesses are the basis for the structural classification.

In table 9 an attempt has been made to tabulate the more important elements in solid solution in kamacite and taenite. The information is scattered and of varying quality, and it is the author's opinion that more work is needed in order to establish partition coefficients between the two major phases, and between these and troilite and silicates. The present information indicates that P, Si and Co are enriched in the kamacite phase, while C, Cr, Cu, Ga and Ge are enriched in the taenite phase, which is in qualitative agreement with laboratory results on ferritic–austenitic steels. However, quantitative data would be appreciated.

The detailed composition, size, substructure and hardness of the kamacite carry intrinsic important information as to the initial cooling-rate and subsequent reheating and/or shock events. For a thorough discussion of these aspects the reader is referred to Buchwald (1975*a*, vol. 1). Here, it must suffice to mention that the kamacite and taenite, being ductile metallic phases, will often occur in cold-worked, recovered or recrystallized forms.

The kamacite may, in addition, occur as unequilibrated phases, produced by a martensitic transformation mechanism, table 18. The α_2 -iron is a hatched kamacite resulting from the diffusionless transformation of high-temperature austenite (with 5–25 % Ni) to metastable ferrite of the same composition. This phase is a normal constituent of the heat-affected surface

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TABLE 9. APPROXIMATE COMPOSITION OF KAMACITE AND TAENITE IN ORDINARY AND ANOMALOUS IRON METEORITES

phase	meteorite	class	group	Bulk Ni		Ni %	P →	C ←	Si	Cr	parts/10 ⁶				
				←	→						Co	Cu	Ga	Ge	
kamacite	ordinary hexahedrite	H	IIA	5.6	5.4	5.4	0.07	100	< 30	40	4300	140	60	180	
kamacite	ordinary octahedrite	Og	I	7.1	6.9	44	0.08	200	< 30	4	6000	150	70	275	
taenite								1500							
kamacite	ordinary octahedrite	Om	IIIA	7.9	7.3	40	0.08	200	< 20	30	6000	100	15	30	
taenite								1000							
kamacite	anomalous	Opl	Butler	15.7	5.8	43	< 0.02		< 23		17000	100	60	1700	
taenite	anomalous		Horse Creek	6.3	4.0	4.0	0.05		25000		6000	300	90	4000	
kamacite	anomalous	Santa Catharina		35.3	35.3	35.3	0.15			5	6000	1000	5	10	

Note: Since all elements mentioned may show a wide range and steep concentration gradients, especially in taenite, the above data should be regarded only as approximate indications.

The papers by Nichiporuk (1958), Goldstein (1966, 1967), Smales *et al.* (1967), Reed (1965, 1969), Wai & Wasson (1969), Wai (1974) and Wasson (1974) should be consulted for further information.

zone of uncorroded iron meteorites; the rapid frictional heating followed by renewed rapid cooling during atmospheric flight is sufficient to create a shallow, 2–4 mm thick, surface zone. The zone has a characteristic hardness of 190–220 V.h.n., somewhat dependent on the exact chemical composition of the parent kamacite and on the grain size. Figure 5 shows a typical hardness traverse across the heat-affected zone towards the interior, which was unaffected by the brief atmospheric flight.

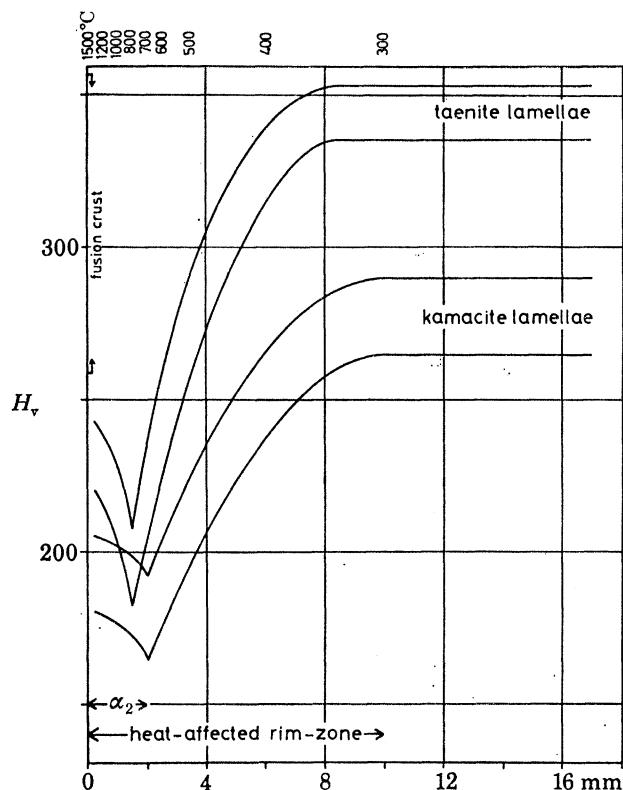


FIGURE 5. Hardness bands for typical kamacite and taenite lamellae in the iron meteorite Cape York. The heat-affected zone is 10 mm deep, but optically visible changes are only present in the exterior 2 mm. The temperatures above, in °C, indicate the approximate maximum temperature reached during the atmospheric flight. (From Buchwald 1975*a*.)

The α_2 transformation products also occur in the exterior layers of stones and stony irons whenever the metallic parts were heated above 750–800 °C; in these meteorites the continuity and presence as a *zone* is, of course, less conspicuous.

If a pressure of about 130 kbar is reached or exceeded, kamacite will transform to a stable, hexagonal close-packed ϵ -iron of high density (Smith 1958; Maringer & Manning 1962; Takahashi & Bassett 1964). In experiments conducted at static pressures, the density of the α -phase increases gradually from 7.86 g/cm³ at 1 atm (10⁵ Pa) to 8.46 g/cm³ at 130 000 atm (13 GPa), then jumps abruptly to 8.81 g/cm³ at the phase transition. Upon pressure release, the ϵ -phase reverts to α (table 18). Rapid $\epsilon \rightarrow \alpha$ transformation results in a distorted α -structure that appears as a densely-hatched martensite upon a polished and etched section. Its microhardness is higher, 300–350 V.h.n., than that of any other kamacite modification. For the sake of brevity and convenience it is often called ϵ -iron, although it is *now* a distorted

α -phase, resulting from the transformation of ϵ -iron proper. Circumstantial evidence and numerous laboratory experiments indicate that the hatched ϵ -iron of iron meteorites was produced when transient shock-waves of high intensity, above 130 kbar (13 GPa), triggered the transformation $\alpha \rightarrow \epsilon \rightarrow$ distorted α . The shock-hatched phase is unequilibrated and will, upon annealing, recover and recrystallize. Iron meteorites do, in fact, present a variety of structures which may be interpreted as the result of various combinations of peak shock-pressures and annealing time and temperature.

Carbon modifications

A survey of all iron meteorites shows that carbon is a common element (0.5–2 % by mass) in group I and some anomalous meteorites, is on a lower level in groups IIIC–D–E and IIA–B (0.05–0.5 %), and is below about 100 p.p.m. in the remaining groups. In accordance with this, graphite and carbides, table 8, are usually present only in groups I, IIA–B, IIIC–D–E and some anomalous irons, while the remaining meteorites only contain carbon in solid solution in the metallic phases, particularly in taenite.

Diamond and lonsdaleite, table 18, are restricted to Cañon Diablo, a group I meteorite which was almost entirely obliterated when it impacted Arizona in a cratering event (Nininger 1956). The impact was of such high intensity that many of the surviving meteorite fragments display a variety of phase transformations, including partial melting of kamacite, troilite, cohenite and schreibersite. Pre-existing graphite under these conditions gave rise to the formation of minute carbonados, which upon closer examination proved to be fine-grained mixtures of (black) diamond and lonsdaleite.

The most common occurrence of graphite is as a constituent of the complex inch-sized nodules of group I. The nodules consist of troilite, graphite and silicates in varying proportions and textures, and they are themselves usually rimmed with schreibersite and cohenite, exsolved from the surrounding metal. Several such inclusions were recently examined by Deines & Wickman (1975) to establish the ^{12}C – ^{13}C ratio of meteoritic minerals.

Triangular or arrowhead development of graphite occurs in a few iron meteorites, such as Kendall County, Morrill and Mundrabilla. Basket weave morphology is particularly well-developed in Kendall County. Cliftonite, i.e. graphite crystals that, in section, display cubic, cubo-octahedral and pseudohexagonal outlines, is common in many group I irons, such as Youdegin, Magura, Seligman, Odessa and Cañon Diablo. The occurrences described in this paragraph are probably best interpreted as the result of carbon precipitation from solid solution in the metal. Although it is often maintained that the distinct crystallographic outlines are due to pseudomorphy, in the author's opinion neither the parent mineral nor the process has been identified. It is more likely that the fine details are governed by lattice forces that seek to accommodate the graphite precipitates with a minimum of lattice strain in the kamacite (and taenite) parent crystals.

Very often the carbon atoms have segregated to preexisting non-metallic inclusions. Thus in Pitts, Four Corners and Persimmon Creek, silicate inclusions have acted as heterogeneous nuclei and are now rimmed by 10–100 μm thick graphite precipitates.

The above-mentioned occurrences are of a primary nature, apparently produced during the primary cooling of the meteorite on its parent body. Graphite does, however, also occur as a late decomposition product of cohenite and haxonite, carbides which probably formed late in the primary cooling period. Although thermodynamically metastable, they have

survived in a number of meteorites and are perhaps over 10^9 years old. In meteorites that have suffered shock and reheating, however, all stages of decomposition may be observed, according to the reaction $(\text{Fe, Ni})_3\text{C} \rightarrow 3\text{Fe}(\text{Ni}) + \text{C}$, figure 6, plate 1. It is noteworthy that the reaction leads to a kamacite that is unusually low in nickel (0.5–2%), because the parent cohenite was low in nickel. The precipitated graphite is lamellar and it may often be observed that decomposition was initiated on a crack, either inside cohenite or along a cohenite–schreibersite interface. Thus here, as elsewhere in meteoritics, nucleation problems appear to have been decisive for the process.

Copper

Native copper is apparently present in minute quantities in many stony meteorites (Ramdohr 1973) and iron meteorites. It has been observed in Campo del Cielo, Cape York, Charcas, Coahuila, Henbury, Landes, Morito, Odessa, Sacramento Mountains, Thule and Toluca. In Cape York it occurs throughout some troilite nodules, concentrated along their rims or associated with chromite and/or phosphates. Sometimes it forms minute specks in chromite. The copper grains are relatively pure, however, with 1–2% iron and nickel in solid solution (Kracher *et al.* 1975); they are bright red on a polished section but tarnish easily. Native copper has not been reported within the metal proper, but the kamacite phase often contains 100–200 parts/ 10^6 Cu in solid solution.

Cohenite and haxonite

Cohenite is the orthorhombic cementite of the metallurgist, distinguished, however, by its significant content of nickel and cobalt, of the order of 0.7–2.3% and 0.02–0.3% respectively. Although phosphorus is sometimes reported as a substituent in the formula, it is probably safe to say that P is below 0.02 mass%, any excess above this being due to contamination by schreibersite.

Cohenite occurs as conspicuous, millimetre-sized crystals in virtually all iron meteorites of group I. Its hardness is about 1100 V.h.n. and it is quite ductile. Since, however, it is irregularly distributed, even large sections of a particular group I iron, such as Magura or Cranbourne, may appear devoid of cohenite. The mineral is also present, but in much smaller quantities, in group IIA–B and in some anomalous meteorites, e.g. Santa Rosa and Chihuahua City. Haxonite is particularly common in group IIIC–D and IIIE; being situated as ‘carbide roses’ inside some plessite fields, it is inconspicuous and often best observed on a polished, unetched slice; here it stands in relief due to its high hardness of about 800 V.h.n.

The carbides are some of the latest minerals to have formed in iron meteorites; they usually precipitated from solid solution after all the major structural elements had grown to their final shape and composition. Therefore, cohenite occurs as rims around troilite, silicate, and, particularly, schreibersite crystals, or it is located along the centre line of kamacite bands. Haxonite apparently precipitated during a slightly later period, either inside kamacite bands, as in Tazewell, or as complex intergrowths with spheroidized plessite, forming ‘carbide roses’ of 0.1–1 mm size. At the relatively low temperatures when haxonite formed, 500–400 °C(?), diffusion of nickel away from the site of precipitation was rather restricted; therefore haxonite will usually have 2–3 times as much nickel in solid solution as cohenite.

As noted above, shock-fracturing plus annealing will decompose the carbides, figure 6; or intense reheating alone can do it, as may be observed in the heat-affected surface zone. A special case is the rapid, but incomplete decomposition that takes place in iron fragments from cratering impacts, such as Cañon Diablo and Morasko. Here the cohenite may be rimmed

by characteristic dark-etching zones of reaction products, figure 7, plate 1, indicating that the fragment for a few minutes reached peak temperatures of 800–900 °C. Thus the secondary transformations of carbides can support the evidence from metallic phases and from graphite and troilite, so that we can reconstruct a cratering-event in some detail.

It has been maintained that cohenite was restricted to meteorites with bulk nickel compositions between 6 and 8 %. Recent work (Buchwald 1975*a*) has, however, demonstrated that both cohenite and haxonite are present in a wide variety of irons, ranging from hexahedrites with 5.4 % Ni (e.g. Negrillos), through octahedrites with 6–9 % Ni (e.g. Ainsworth, Toluca) to anomalous polycrystalline irons of very high bulk nickel content (Pitts 12.9 %; San Cristobal 25.6 %).

TABLE 10. MINERALS IN IRON METEORITES. II: NITRIDES, PHOSPHIDES AND SULPHIDES

	composition	crystal system	source	meteorite example
carlsbergite*	CrN	cubic	Buchwald & Scott 1971	Kenton County
schreibersite*	(Fe, Ni) ₃ P	tetragonal	Haidinger 1847, 1861	Tombigbee
alabandite	(Mn, Fe, Cr)S	cubic	Bunch <i>et al.</i> 1970	Mundrabilla
brezinaite*	Cr ₃ S ₄	monoclinic	Bunch & Fuchs 1969 <i>a</i>	Tucson
chalcopyrite†	CuFeS ₂	tetragonal	Ramdohr 1973, p. 43	Delegate
daubréelite*	FeCr ₂ S ₄	cubic	J. Lawrence Smith 1876	Coahuila
djerfisherite	K ₃ CuFe ₁₂ S ₁₄ Cl	cubic	El Goresy <i>et al.</i> 1969, 1971	Cape York
mackinawite†	FeS	tetragonal	Ramdohr 1973, p. 34	Delegate
sphalerite	ZnS	cubic	El Goresy 1965	Cañon Diablo
troilite*	FeS	hexagonal	Haidinger 1863	Cape York

† Minute and rare occurrences and not always of undisputed primary origin.

NITRIDES, PHOSPHIDES AND SULPHIDES

Carlsbergite, CrN, is a cubic mineral (table 10) which is quite common in group IIIA and IIA, but since it mainly occurs as minute rods and platelets in the kamacite phase it is easily overlooked – or mistaken for rhabdite. Carlsbergite is much more ductile than rhabdite: in a distorted metal matrix carlsbergite platelets will bend and twist, while rhabdites will break. Carlsbergite is also more stable than rhabdite against brief reheating: while rhabdites will decompose and finally melt above 1100–1200 °C, changes in carlsbergite are difficult to detect at the corresponding thermal treatment. Carlsbergite and rhabdite often occur together, evidently having coprecipitated from kamacite solid solution.

Schreibersite (table 10) is a very important mineral. It is not quite as hard (V.h.n. \approx 900) as cohenite and much more brittle. The pure end members, Fe₃P and Ni₃P, are not known, most occurrences showing from 10–15 mass % Ni (large crystals in Bellsbank, Cañon Diablo, São Juliao and Sikhote-Alin) to about 50 % (small grains in α - α and α - γ grain boundaries in numerous octahedrites). Cobalt is also present, but usually on a much lower level of 0.15–0.5 %. Preliminary analyses indicate that some copper (300–500 parts/10⁶) and zirconium (500–1000 parts/10⁶) are also present. For a penetrating discussion and analysis of schreibersite occurrences, see Clarke (1976).

Due to its brittle nature, schreibersite usually became severely fragmented and brecciated when the parent body or meteorite was exposed to shock-events. Once landed on Earth, terrestrial ground water therefore had easy access to the deep interior. Many iron meteorites are, far from being massive bodies, really beset with numerous interior cracks.

Schreibersite is present in most iron meteorites but absent from, e.g. the low Ni members of group IVA and from Arlington, Ysleta and Santiago Papasquiario. When the bulk P content in group IVA increases above 0.06 %, the first minute schreibersite particles usually become visible. Particularly rich in schreibersite crystals are Tombigbee River, La Primitiva, Bellsbank, Soper, and Ni-rich members of group IIIB, such as Chupaderos, Grant and Narraburra.

Alabandite, brezinaite and daubréelite

Alabandite (table 10) is a very minor mineral in iron meteorites, occurring in intimate contact with troilite and sometimes sphalerite in the following irons: Cañon Diablo, Cape York, Kendall County, Landes, Mundrabilla, Odessa and Toluca. Essentially it is manganese sulphide, but replacement of up to one-tenth of the manganese atoms with iron, chromium and zinc is apparently the rule rather than the exception (Bunch *et al.* 1970; Ramdohr *et al.* 1975).

Brezinaite (table 10). An extremely rare sulphide, it was reported only from Tucson and New Baltimore (Buchwald 1975*a*, p. 896). It is a monoclinic chromium sulphide with a few percent of iron and occurs as anhedral, (5–100) μm grains in kamacite. In Tucson, brezinaite is intergrown with silicates and is the only sulphide present; in New Baltimore, there are no silicates, and troilite and the chromium-bearing minerals carlsbergite and daubréelite are common.

Daubréelite. Although a minor accessory it is widespread in iron meteorites, being particularly common in groups I, IIA, IIIA and IVA. Besides occurring as exsolution lamellae parallel to (0001) of troilite, it is common as discrete grains 10–500 μm across in kamacite. Daubréelite crystals may contain minor amounts of manganese and zinc in solid solution (see, for example, Schwarcz *et al.* 1975). Sometimes cosmic shocks and associated reheating have entirely destroyed the regular troilite–daubréelite aggregates. Thereby the daubréelite has been scattered and partially dissolved in the sulphide melt, and now occurs as dispersed subangular particles in the solidified iron–sulphur eutectic.

Artificial reheating of troilite–daubréelite aggregates may cause the minerals to recrystallize and to show secondary exsolution reactions (figure 8, plate 1). Very fine exsolution lamellae may thus be observed in the artificially reheated meteorites Rodeo, Verkhne Udinsk and Yanhuitlan. Ramdohr (1973, p. 39 and Plate B47) has described the peculiar morphology of a similar assemblage from Hoba. Perhaps the ‘mysterious’ appearance was also here caused by artificial reheating; it is known that material from the giant 60 t meteorite in the beginning could only be removed by cutting with an oxy-acetylene blowtorch. The violent heat-input into the relatively small cut sections would in the author’s opinion suffice to create the observed structures.

Troilite and other iron and iron copper sulphides (table 10). There has always been some confusion with respect to these sulphides in iron meteorites. There are several polymorphs and some of them occur only as minute grains or as exsolution lamellae, so that under certain conditions it is difficult to decide whether they are primary or alteration productions of terrestrial weathering. It is therefore only in recent years, since the advent of the powerful electron microprobe technique, that the definitive identification seems to be near at hand. Of the following minerals, which have been reported through the years – pyrite, pyrrhotite, troilite, chalcopyrite, chalcopyrrhotite, cubanite, valleriite, mackinawite, djerfisherite and gentnerite – only one, troilite, is really present in significant quantities. Djerfisherite, mackinawite and chalcopyrite have each been reported from one, two or three iron meteorites, and it is not yet entirely clear

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TABLE 11. COMPOSITION OF TROILITE. SELECTED ANALYSES

(Most of them are electron microprobe (ref. 1, 3, 4, 7, 10) and mass spectrometric analyses (ref. 6, 8). Mass %)

name	group	S	Fe	Ti	V	Cr	Mn	Co	Ni	Cu	Zn	ref.
Campo del Cielo	I	36.6	62.4	0.08	< 0.02	0.29	0.15	—	—	—	0.30	1
Cañon Diablo ^(c)	I	—	62.8	—	0.004	0.71	—	0.005	0.017	0.015	< 0.005	2
Cañon Diablo ^(c)	I	35.0	61.6	—	—	1.18	—	0.025	0.081	—	—	9
Gladstone	I	37.0	63.8	0.01	0.01	0.59	0.00	0.02	0.00	—	0.04	10
Landes	I	36.3	62.8	0.08	< 0.02	0.21	0.09	—	—	—	0.12	3
Linwood	I	37.2	62.5	0.09	n.d.	0.60	0.16	—	—	—	0.31	1
Morasko ^(a)	I	35.0	63.0	0.05	—	0.8	n.d.	—	—	—	n.d.	4
Morasko ^(b)	I	37.0	57.0	n.d.	—	n.d.	0.3	—	—	—	0.4	4
Odessa	I	37.3	62.4	0.09	< 0.02	—	0.17	—	—	—	0.30	1
Odessa ^(c)	I	—	—	—	0.004	0.66	0.063	0.025	0.23	0.0083	< 0.002	7
Toluca	I	37.0	62.5	0.10	n.d.	0.41	0.18	—	—	—	0.28	1
Copiapo	I-An.	36.4	62.7	< 0.03	< 0.02	0.29	< 0.02	—	—	—	< 0.02	1
Four Corners	I-An.	37.5	62.9	< 0.03	n.d.	0.05	< 0.02	—	—	—	< 0.02	1
Persimmon Creek	I-An.	36.6	63.1	< 0.03	n.d.	0.17	< 0.02	—	—	—	< 0.02	1
Pine River	I-An.	36.4	62.6	< 0.03	0.02	0.30	< 0.02	—	—	—	< 0.02	1
Pitts	I-An.	36.4	62.6	< 0.03	n.d.	0.30	< 0.02	—	—	—	< 0.02	1
Coahuila ^(c)	IIA	—	55.7	—	0.041	5.57	—	0.013	0.55	0.10	0.01	2
Sikhote-Alin ^(c)	IIB	—	62.6	—	0.007	1.58	—	0.012	0.91	0.08	< 0.005	2
Sikhote-Alin ^(c)	IIB	34.5	62.7	—	—	1.07	—	—	—	0.06	—	5
Cape York ^(a)	IIIA	—	—	—	0.0008	0.10	0.017	0.0036	0.034	0.005	< 0.002	6
Cape York ^(a)	IIIA	—	—	—	0.023	5.0	0.053	0.0063	0.064	0.047	0.13	6
Colomera	Anom.	36.4	63.0	< 0.03	< 0.02	< 0.02	< 0.02	—	—	—	< 0.02	1
Kendall County	Anom.	38.0	61.7	0.08	0.13	0.51	0.16	—	—	—	0.30	1
Kodaikanal	Anom.	36.3	63.1	< 0.03	< 0.02	< 0.02	< 0.02	—	—	—	< 0.02	1
Mundrabilla ^(c)	Anom.	—	—	—	0.0094	1.1	0.12	0.0013	0.075	0.023	< 0.008	7
Netschaevo	Anom.	36.3	62.6	< 0.03	n.d.	0.05	< 0.02	—	—	—	< 0.02	1
Soroti	Anom.	—	—	0.002	0.0005	0.073	0.086	0.004	0.057	0.013	n.d.	8
Weckerroo Station	Anom.	36.4	63.1	< 0.03	< 0.02	< 0.02	< 0.02	—	—	—	< 0.02	1

References: (1) Bunch *et al.* 1970; (2) Nichiporuk & Chodos 1959; (3) Bunch *et al.* 1972; (4) Dominik 1976; (5) Dyakonova 1958; (6) Jochum *et al.* 1975; (7) Jochum & Hintenberger, personal communication; (8) Mason & Graham 1970; (9) Weinke *et al.* 1971; (10) Schwarcz *et al.* 1975.

(a) Troilite without inclusions; (b) troilite adjacent to sphaerite; (c) probably contaminated by daubréelite, chromite, etc. but no details; (d) contaminated by chromite and phosphates; (e) contaminated by graphite (2%) and chromite.

whether all of the occurrences are primary. The remaining minerals are absent or need re-definition, and chalcopyrrhotite itself has been discredited as a distinct species by Cabri (1967).

Troilite is non-magnetic, if pure; its microhardness is 250 ± 30 V.h.n. It often contains parallel daubréelite lamellae; grains of chromite, sphalerite, graphite and various phosphates and silicates may occur in profusion. Particularly the troilite nodules of chemical group I are rich in accessory minerals; moreover, graphite and silicates appear to be almost exclusively restricted to group I and to a few anomalous iron meteorites.

The bulk sulphur content of iron meteorites ranges from about 0.02 % to 12 % (Buchwald 1975*a*, Table 30). The sulphur is present in solid solution (up to about 100 parts/10⁶) in the metallic phases, and beyond this as troilite and other sulphides. Tucson and Nedagolla are remarkable by the absence of troilite; instead, Tucson has brezinaite and Nedagolla daubréelite as minor accessories.

Almost pure stoichiometric troilite has been reported from the three related anomalous meteorites, Colomera, Kodaikanal and Weekeroo Station. As may be seen from tables 11 and 12, most other troilite occurrences are less pure. Apparently, the troilite very often carries in solid solution up to about 1 % total of the elements Cr, Mn, Zn, Ni, Cu and Co, while what is reported beyond this level may be present in finely disseminated inclusions of metal, chromite, sphalerite, daubréelite and phosphates. The two analyses quoted from Cape York were performed upon material from opposite ends of a 73 mm long troilite inclusion. Apart from the chromium content, one troilite end is as pure as that of Colomera and compares favourably with Copiapo and other irons of group I–Anomalous. The opposite end, however, is enriched in Cr, Zn, Mn, Cu, Ga and Se; most likely these elements are concentrated in chromite and phosphate inclusions in the troilite, the presence of which was confirmed by optical microscopy.

Uranium contents of troilite samples from Sikhote-Alin, Augustinovka and Cape York have been determined by Levsky & Komarov (1975). The average uranium content is 0.4–0.7 part/10⁶; the Cape York (Agpalilik) sample is even below 0.04 part/10⁶.

Older analyses, e.g. by Buddhue (1945), have often shown substantial amounts (1–3 %) of nickel and cobalt in troilite. The author has been suspicious about these reports, and the data of table 11 on well-documented material confirm that nickel is usually well below 0.1 % while cobalt is on a ten times lower level. This is in agreement with the experimental data of Kullerud (1963), who for the Fe–Ni–S system reported troilite to contain less than 0.5 % Ni at 900 °C and less than 0.2 % at 650 °C. The reason for the previous high figures was, in particular, the analytical failure to separate quantitatively finely disseminated metal and sulphide phases from shock-melted troilite nodules.

Troilite nodules are often wrapped in successive sheets of schreibersite, cohenite and kamacite, and many authors have considered that these minerals were formed by exsolution from the cooling troilite. Various lines of evidence suggest, however, that they were rather formed by exsolution from the metal phase, and that the troilite mainly served to provide heterogeneous nuclei for the precipitating phosphides and carbides. When, in some meteorites, such as Henbury and Gibeon, the troilite is virtually free of schreibersite and cohenite wrappings, it is because the bulk phosphorus and carbon contents in these meteorites are so low, that all P and C atoms can remain in solid solution in the metal. On the other hand, in Chupaderos, with a high bulk phosphorus concentration (0.5 %), or in Odessa, with high phosphorus (0.25 %) and carbon content (> 0.2 %), the metallic phases on cooling will become super-saturated, and the rim zones are then developed to perfection.

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TABLE 12. COMPOSITION OF TROILITE. ADDITIONAL ANALYSES FOR SOME MINOR ELEMENTS. MASS CONCENTRATIONS IN PARTS/10⁶

	Na	P	K	Ga	Ge	Se	Mo	Te	W	Ir	Pt	Au	ref.
Cañon Diablo ^(a)	—	180	—	—	—	262	3.0	—	—	0.01	—	0.005	1
Odessa ^(a)	13	430	17	< 0.06	< 0.13	33	1.1	0.6	—	—	—	—	2
Cape York ^(a)	2.1	0.9	3.4	0.2	0.1	38	8.6	0.5	< 0.8	< 0.9	< 0.7	< 0.1	3
Cape York ^(b)	2.1	4.0	2.8	12.0	0.19	92	15	1.2	< 0.7	< 1.0	< 0.8	< 0.2	3
Mundrabilla ^(a)	51	5.1	32	< 1.6	< 0.7	99	< 15	< 0.5	—	—	—	—	2

(a) Pure troilite; (b) troilite contaminated by chromite and phosphates; (c) troilite contaminated by graphite and probably also chromite, silicates and phosphates.

References: (1) Weinke *et al.* 1971 (neutron activation); (2) Jochum & Hintenberger, personal communication (mass spectroscopy); (3) Jochum, Hintenberger & Buchwald 1975 (mass spectroscopy).

The exterior wrapping of kamacite, which to varying degrees is present whether or not the troilite has schreibersite and cohenite rims, must again reflect a nucleation phenomenon. When upon primary cooling the high-temperature austenite (taenite) phase became unstable, the transformation to ferrite (kamacite) took place heterogeneously upon inclusions, so that these became enveloped in 'swathing kamacite'. Away from the inclusions, however, the transformation had to occur homogeneously which resulted in the well-known Widmanstätten structure.

Mackinawite (table 10). This mineral was named from terrestrial occurrences by Evans *et al.* (1964). It is a tetragonal iron sulphide that in polished sections closely resembles valleriite, a hexagonal layer-lattice mineral $(\text{CuFeS}_2)(\text{Mg}(\text{OH})_2)$ (table 20). When Ramdohr in the early 1960s examined a number of meteoritic sulphides, he evidently saw mackinawite several times, but wrongly called it valleriite, because mackinawite at that time was undescribed. Most of his observations were, however, on stony meteorites (Ramdohr 1963, 1973, 1975). The only occurrence of mackinawite in iron meteorites seems to be that in Delegate, a medium octahedrite, in which the mineral appears as minute grains inside troilite together with chalcopyrite, pentlandite, alabandite and chalcopyrrhotite.

Chalcopyrite (table 10) is extremely rare in iron meteorites, having been observed only by Ramdohr (1963, 1973) and El Goresy (1965). The occurrences seem restricted to minute lamellae and specks in troilite, and they are usually associated with pentlandite, native copper or mackinawite. It is not clear whether the minerals are primary or were formed in secondary reactions during terrestrial weathering. The occurrences on record are all from iron meteorites known to have been exposed to corrosion for thousands of years.

Chalcopyrrhotite (table 18). Ramdohr (1963, 1973, 1975) and Yund & Kullerud (1966) have described this 'mineral' as an undercooled metastable solid solution, of cubic structure, of CuFeS_2 and FeS in varying proportions. The stability varies from case to case and is perhaps influenced by foreign atoms such as nickel. On slow cooling chalcopyrrhotite will decompose to chalcopyrite and mackinawite, or to other combinations of sulphides; on rapid cooling, however, it may survive metastably. Ramdohr (personal communication) and El Goresy (1965) have reported it from Cape York, Delegate, Odessa, Soroti and N'Goureyima.

Djerfisherite (table 10) is a rare potassium- and chlorine-containing sulphide mineral of a surprising composition. It was first recorded from enstatite chondrites (Fuchs 1966), then also from achondrites and the iron meteorite Toluca (El Goresy *et al.* 1971). El Goresy examined more than 150 troilite nodules from 15 iron meteorites and found only a single occurrence in Toluca, the mineral being squeezed in 100–300 μm patches between troilite and graphite. Recently, djerfisherite has also been found in Cape York, where it occurs in small irregular specks inside troilite (A. Kracher, personal communication). After its discovery in meteorites it has been reported as a rare accessory in some terrestrial ore bodies (Sokolova *et al.* 1970).

Sphalerite

Sphalerite (table 10) has been reported as 20–150 μm grains, associated with troilite and often also with alabandite and daubréelite, from Bogou, Campo del Cielo, Cañon Diablo, Cape York, Charcas, Gladstone, Landes, Morasko, Mundrabilla, Odessa, Pan de Azucar, Sardis, Toluca and Udei Station, i.e. so far mainly from groups I and IIIA iron meteorites. Its composition is variable and exemplified by the following analyses: 58 % Zn, 9 % Fe, 3 % Mn, 32 % S (Cape York (IIIA); A. Kracher, personal communication); 40 % Zn, 21.5 % Fe,

4.5 % Mn, 34 % S (Bogou (I); Schwarcz *et al.* 1975); 33 % Zn, 28 % Fe, 4 % Mn, 36 % S (Landes (I); Bunch *et al.* 1972) and 38 % Zn, 28 % Fe, 0.3 % Mn, 34 % S (Mundrabilla (anomalous); Ramdohr & El Goresy 1971). Other reports are those of Ramdohr (1963, 1973), El Goresy (1965) and Bunch *et al.* (1970). Sphalerite has received renewed attention after the suggestion that its iron content is an indicator of the pressure when the mineral cooled through the interval 500–350 °C.

Schwarcz *et al.* (1974, 1975) thus argue that the ZnS–FeS–Fe system is sensitive to both temperature and pressure, but only little influenced by the presence of manganese. Therefore, if we estimate the temperature when diffusional processes came to a stop, and measure the composition of sphalerite in contact with troilite and kamacite, the pressure can be deduced. Higher mole fractions of FeS correspond at the same temperature to lower pressures. The preliminary reports suggest that the final equilibrium of the sphalerite occurred at pressures ranging from 0.2 to about 3 kbar (20 to 300 MPa), corresponding to formation at centres of chondritic objects from 100 to 400 km in radius, or in non-central regions of correspondingly larger objects.

TABLE 13. MINERALS IN IRON METEORITES. III: OXIDES

	composition	crystal system	source	meteorite example
chromite	FeCr ₂ O ₄	cubic	Rose 1864; Bunch & Olsen 1975	Sikhote-Alin
cristobalite	SiO ₂	tetragonal	Marvin 1962	Carbo
ilmenite	FeTiO ₃	hexagonal	Bunch & Olsen 1968	Weekeroo Station
rutile	TiO ₂	tetragonal	El Goresy 1965	Odessa
tridymite	SiO ₂	monoclinic	Marvin 1962; Götz 1962	Colomera
zircon	ZrSiO ₄	tetragonal	Laspeyres 1897	Toluca

OXIDES

Chromite (table 13) was in the beginning thought to be rare, but has now proven to be present in all major types of meteorites, and in all the important chemical groups of iron meteorites. It occurs as minute euhedral or rounded grains in troilite of, e.g. Cape York and N'Goureyima, as extended platelets in, e.g. Monahans, Cape York and Duchesne, and as rare, large crystals in Bendego and Sacramento Mountains, even reaching the surprising dimensions of 8 × 4 cm in Sikhote-Alin (Kvasha 1963). Burns & Burns (1975) and Bunch & Olsen (1975) have recently reviewed the mineralogy of chromium, the latter authors with special reference to meteorites.

Chemically, chromite shows a wide range. Reports by Bence & Burnett (1969), Bunch & Keil (1971), Bunch *et al.* (1970, 1972) and the author's own data suggest that the range can be given as: 49–74 % Cr₂O₃, < 0.03–17.6 % Al₂O₃, < 0.03–0.75 % V₂O₃, < 0.02–4.4 % TiO₂, 2.5–35.2 % FeO, < 0.02–14.2 % MgO, 0.5–4.2 % MnO and < 0.02–2.3 % ZnO. Individual chromite grains of Cape York are on the average nearly stoichiometric (68 % Cr₂O₃, 32 % FeO), but show under the electron microprobe small contents of V₂O₃ (0.3 %), MnO (0.6 %) and ZnO (0.2 %), the zinc being enriched in an outer zone. Similar results have been presented from another IIIA iron, Bagdad (Bunch & Keil 1971).

The so-called Reichenbach lamellae were seen by Reichenbach (1861) and named by Brezina (1880). When best developed they form very thin, < 0.05 mm, but large, 20 mm × 30 mm, platelets parallel to three planes through the metallic matrix, generally parallel to {100}_γ. Additional planes are often present, e.g. in Cape York. Contrary to general opinion, many of

these lamellae have turned out to be, not pure troilite, but troilite deposited upon a very thin backbone of chromite. It appears therefore that it is the crystallographic relationship of chromite to high-temperature taenite that has determined the shape and morphology of the precipitates, and that troilite – and often schreibersite – somewhat later found an appropriate substrate for nucleation here.

Cristobalite and tridymite (table 13) occur as accessory minerals in a few iron meteorites. Cristobalite is the rarer, having been reported only from Carbo (Marvin 1962), associated with glass and troilite, and probably having been formed by shock during cosmic collisions. Tridymite occurs in Colomera, Kendall County, Weckerroo Station, Gibeon and Steinbach, but an old report of its presence in Kodaikanal (Klein 1906) has not been confirmed. According to analyses by, e.g. Bunch *et al.* (1970) and Mason (1972*b*), the minerals are more than 96.5 % SiO₂, the rest being mainly oxides of Al, Ca, K and Na. Mason (1972*b*) notes that it is remarkable that the tridymite and cristobalite, in meteorites and in lunar rocks, have survived for aeons without inverting to quartz, as has occurred in most ancient terrestrial rocks. He speculates that it is due to the extremely anhydrous nature of the materials; the catalytic action of water in promoting recrystallization and equilibration is well documented.

Ilmenite has been recorded from a large number of stone meteorites and mesosiderites (Ramdohr 1973), but apparently only from two iron meteorites. In Edmonton (Kentucky), it appears in a rare silicate nodule (Ramdohr 1973), and in Weckerroo Station ilmenite occurs with chromite. Its composition is 53.0 % TiO₂, 36.1 % FeO, 6.7 % MgO, 2.7 % MnO, 0.3 % Cr₂O₃ and 0.4 % V₂O₃ (Bunch *et al.* 1970).

Rutile was reported by El Goresy (1965) and Buseck & Keil (1966) from Odessa, and a few stones and mesosiderites, then also found as a very minor accessory in the anomalous iron meteorites Colomera (Bunch & Olsen 1968), Kodaikanal (Bence & Burnett 1969), Weckerroo Station (Bunch *et al.* 1970) and Mundrabilla (Ramdohr & El Goresy 1971).

Finally, rutile was reported also from another normal group I meteorite Toluca (El Goresy 1971). El Goresy pointed out that the rutile from iron meteorites is enriched in niobium, showing 1.6–2.9 % NbO₂. There is also a little Cr₂O₃ (0.8–1.2 %) and FeO (0.1–1.0 %).

Zircon (table 13). Laspeyres (1897) reported zircon from the insoluble residue of Toluca, but later investigators assumed that the zircon came from terrestrial contamination. Marvin & Klein (1964) proved, however, that the zircon was indigenous to Toluca, and they found 10 µm grains, fluorescent under the electron beam, in the Vaca Muerta mesosiderite. Having thus been reported only twice, and then only after extensive search through much material, it must be assumed that zircon is extremely unusual in meteorites.

PHOSPHATES

The *phosphates* (table 14) have virtually all arrived on the scene after the advent of the electron microprobe. The only one known prior to this was merrillite (table 20), described by Wherry (1917), and later discredited in favour of the more precisely analysed and defined whitlockite (compare, for example, the brief summary by Mason (1967*a*, p. 319)). Most of the phosphates are small, i.e. < 0.5 mm across, and they are usually associated with troilite and chromite, but may also be found in the metallic matrix. Characteristically, the phosphates are hydroxyl-free, but they may contain halogens. Much of the chlorine which is present in unweathered iron meteorites seems to be bound in chlorapatite and similar compounds; some

may be located in djerfisherite (table 10). Detailed analyses of chlorapatite and whitlockite from Copiapo and other irons have been reported by Bunch *et al.* (1970); the chlorapatite of group I irons, e.g. Odessa and Toluca, was found to contain 3.6–4.7 % Cl and 0.6–1.3 % F. Chlorapatite from Campo del Cielo, also a group I iron, contained 6.0 % Cl (Wlotzka & Jarosewich, personal communication).

TABLE 14. MINERALS IN IRON METEORITES. IV: PHOSPHATES

	composition	crystal system	source	meteorite example
brianite*	$\text{Na}_2\text{CaMg}(\text{PO}_4)_2$	rhombic	Fuchs <i>et al.</i> 1967	San Cristobal
buchwaldite*	NaCaPO_4	rhombic	Olsen <i>et al.</i> 1976	Cape York
chlorapatite	$\text{Ca}_5(\text{PO}_4)_3\text{Cl}$	hexagonal	Fuchs & Olsen 1965	Odessa
farringtonite*	$(\text{Mg, Fe})_3(\text{PO}_4)_2$	monoclinic	Bild 1974	Barranca Blanca
grafonite	$(\text{Fe, Mn, Mg})_3(\text{PO}_4)_2$	monoclinic	Olsen & Fredriksson 1966	Wiley
panethite*	$\text{Na}_2(\text{Mg, Fe})_2(\text{PO}_4)_2$	monoclinic	Fuchs <i>et al.</i> 1967	Dayton
sarcopsidite	$(\text{Fe, Mn})_3(\text{PO}_4)_2$	monoclinic	Olsen & Fredriksson 1966	Wiley
whitlockite	$\beta\text{-Ca}_3(\text{PO}_4)_2$	hexagonal	Fuchs 1962, 1969	Colomera

Until the early sixties potassium, sodium, magnesium, calcium and manganese in meteorites were considered to be exclusively present in the various silicate minerals. From table 14 it will, however, be noted that the situation has changed significantly. It appears that the phosphates, besides a few sulphides such as alabandite, djerfisherite and sphalerite, rather actively have concentrated these minor elements, while the metal phase and troilite have become depleted in them. Phosphates generally occur with phosphide and thus can provide a measure of the equilibrium oxygen fugacity at the time of their formation. Olsen & Fuchs (1967) have performed some of these calculations for iron meteorites.

Present research continues to bring up new phosphates, such as farringtonite and buchwaldite, and no doubt many more phosphates will be reported in coming years. Particularly the troilite nodules of Cape York (IIIA) and many group I irons provide complex polymineralic aggregates from which new minerals constantly peep out (Marshall & Keil 1965; El Goresy 1965; Buchwald 1975*a*, Figure 176).

SILICATES

Silicates (table 15) are a very characteristic component of coarse octahedrites of group I and of some anomalous irons. They are virtually absent from groups II, III and IV, although occasional silica polymorphs are known from troilite nodules of, e.g. Cape York, Gibeon and Bishop Canyon. The important silicates are olivine, pyroxenes and feldspars, while the remaining ones are very rare accessories, each occurring in minute grains in only one or two iron meteorites. A survey of silicates in irons, giving the names of the actual meteorites with inclusions, will be found in Buchwald (1975*a*, Appendix 1).

Mason (1967*b*) noted that the minerals of the inclusions of Woodbine, a group I iron, were essentially the same as those in bronzite (H) and enstatite (E) chondrites, and the bulk Si/Mg atomic ratio of 1.09 was also similar. Furthermore, Wlotzka & Jarosewich (1975, and personal communication) have shown that the amount of silicates in the inclusions of Campo del Cielo (table 16) is quite close to the average silicate content of bronzite chondrites, which is 76 %. The albitic character of plagioclase in groups I and I–Anom (table 17) also relates the inclusions to ordinary chondrites and distinguishes them from mesosiderites and achondrites, which contain

anorthitic plagioclase. The main difference from normal H and L chondrites is the FeO content of olivine and pyroxene which in group I and I-Anom inclusions is significantly lower than that of H-chondrites.

TABLE 15. MINERALS IN IRON METEORITES. V: SILICATES

	composition	crystal system	source	meteorite example
olivine	$(\text{Mg, Fe})_2\text{SiO}_4$	rhombic	J. Lawrence Smith 1855	Campo del Cielo
orthopyroxene	$(\text{Mg, Fe})\text{SiO}_3$	rhombic	Shepard 1854; Cohen 1894	Woodbine
clinopyroxene	$(\text{Mg, Fe, Ca})\text{SiO}_3$	monoclinic	Cohen 1894	Netschaevo
plagioclase, albite	$\text{NaAlSi}_3\text{O}_8$	triclinic	Genth 1855; Cohen 1900	Copiapo
plagioclase, anorthite	$\text{CaAl}_2\text{Si}_2\text{O}_8$			
potash feldspar	$(\text{K, Na})\text{AlSi}_3\text{O}_8$	triclinic	Bunch & Olsen 1968	Kodaikanal
ureyite*	$\text{NaCrSi}_2\text{O}_6$	monoclinic	Laspeyres 1897; Frondel & Klein 1965	Toluca
krinovite*	$\text{NaMg}_2\text{CrSi}_3\text{O}_{10}$	monoclinic	Olsen & Fuchs 1968	Youndegin
fluorichterite	$\text{Na}_2\text{Ca}(\text{Mg, Ti})_5\text{Si}_8\text{O}_{22}\text{F}_2$	monoclinic	Olsen 1967; Olsen <i>et al.</i> 1973	Cañon Diablo
roedderite*	$(\text{Na, K})_2\text{Mg}_2(\text{Mg, Fe})_3\text{Si}_{12}\text{O}_{30}$	hexagonal	Olsen 1967	Wichita County
yagiite*	$(\text{Na, K})_{1.5}\text{Mg}_2(\text{Al, Mg, Fe})_3(\text{Al}_2\text{Si}_{10}\text{O}_{30})$	hexagonal	Bunch & Fuchs 1969 <i>b</i>	Colomera

TABLE 16. MINERAL COMPOSITION OF THREE SILICATE INCLUSIONS IN THE GROUP I-IRON CAMPO DEL CIELO

(Wlotzka & Jarosewich 1975; and personal communication)

mass %	inclusion 1	inclusion 4	inclusion 5
olivine	33	21	43
orthopyroxene	35	44	21
clinopyroxene	2.5	0.8	2.7
plagioclase	5.4	11.5	7.8
sum of silicates	75.9	77.3	74.5
graphite	22	14	0.8
troilite	0.1	3.4	16.5
nickel-iron	0.6	5.5	4.9
iron oxide†	1.7	—	3.2
total	100.3	100.2	99.9
points counted	1300	1000	1000

† Formed by terrestrial weathering.

Having thus suggested a relationship between silicate inclusions of group I and group I-Anom on one side and H-chondrites on the other side, there remains the question whether this relationship is a genetic one, whether for example these irons can have formed by adding chondritic material to an iron pool. The chondritic nature of the silicates is no longer evident; they now display a metamorphic texture with equiaxial and allotriomorphic grains, suggesting recrystallization and possibly melting. If metallic and chondritic fragments were mixed and we assume a sustained thermal source, supporting temperatures of 1200–1400 °C for long times, it is possible that compression and sintering, without bulk melting, could have led to the

observed meteorite types, with angular silicate fragments of mm–cm size solidly embedded in a metallic matrix (figure 9, plate 2). Low-melting eutectics of Fe, Ni, P, C and S, liquid to well below 900 °C, may have been present and promoted the production of massive bodies; on the other hand, the absence of liquid phases in bulk quantities could have prevented the physical separation of light silicate fractions. On cooling, but still at temperatures above 1000 °C, the metallic phases would have homogenized and grain growth occurred; however, the silicate inclusions in, e.g. Campo del Cielo (El Taco), have clearly pinned the grain boundaries so that the (taenite) grains never grew beyond diameters of 5–50 μm , as opposed to the metre sized crystals found in, for example, Cape York, a meteorite which is silicate-free and probably at one time was entirely molten.

If the original stony matter was chondritic in composition the presently observed low FeO content of olivine and pyroxene could have been produced during the thermal excursion associated with mixing. Reduction by the ubiquitous graphite and transfer by diffusion of some iron from the silicates to the metallic matrix would bring about the change, while the abundant metallic matrix would hardly be affected at all. Wlotzka (personal communication), who has proposed this reaction, has found in olivines a 10 % lower FeO content along the rims than in the centres, supporting the diffusional character of the process.

Netschaëvo (table 17 and figure 9) has angular silicate fragments and in some of these the chondritic texture is still apparent (Buchwald 1967; Olsen & Jarosewich 1971). The silicates display FeO contents very close to those of H-chondrites, so perhaps the anomalous Netschaëvo meteorite represents a case of mixing, where time was insufficient to erase the chondrule structure and reduce the iron content of the silicates appreciably. Perhaps, in addition, reducing agents such as graphite were only present in limited quantities. Unfortunately, the Netschaëvo meteorite has been almost lost, and what remains has been severely altered by its early possessors (Buchwald 1967), so we apparently have a tough case to solve.

The physical process of mixing metal and silicates is difficult to visualize. Some interesting speculations concerning the Tucson iron meteorite have been presented by Miyake & Goldstein (1974), and other models have been discussed by, for example, Olsen & Jarosewich (1970).

Olivine

Olivine (tables 15–17) constitutes one-third to one-half by mass of the silicate inclusions in group I and group I–Anom. It is more sparsely represented in other irons and it is absent in Kendall County and Weckeroo Station. Within most classes the range of composition is small (table 17), but in mesosiderites and carbonaceous chondrites there is a wide range, both within the classes and within any single meteorite. In most olivines, see for example Bunch *et al.* (1970), only SiO₂, FeO, MgO and MnO are present in measurable quantities, and MnO is very minor, showing a range of only 0.17–0.46 %. Olivine of almost pure forsteritic composition has been reported from Tucson by Bunch & Fuchs (1969*a*).

The trace element content of olivine has mainly been studied on pallasitic material. Here, Buseck & Goldstein (1969) reported 50–70 parts/10⁶ Ni, 15–20 parts/10⁶ Ti and 20–75 parts/10⁶ Ca. Particularly the low nickel content is noteworthy, because terrestrial olivine, from dunite and periodotite, generally has as much as 0.1–0.4 % Ni. This apparent anomaly is best interpreted as an effect of thermal equilibration between olivine and nickel-iron in pallasites, the extremely siderophilic nickel largely diffusing into the metallic phase. In terrestrial rocks no metallic recipient for nickel was directly available.

TABLE 17. AMOUNT AND COMPOSITION OF SILICATES IN VARIOUS METEORITE CLASSES AND IN SOME IRON METEORITES WITH SILICATE INCLUSIONS

symbol	silicate amount†	olivine‡ Fa	Pyroxenes										metallic iron¶
			Ca-poor§			Ca-rich§				plagioclase			
			En	Fs	Wo	En	Fs	Wo	Ab	An	Or		
enstatite chondrites	65-71	÷	98-100	0-1	0-1	98-100	0-1	0-1	81.0	15.0	4.0	19-28	
bronzite chondrites	74-82	16-20	81.7	16.8	1.5	48.8	6.1	45.1	81.9	12.3	5.8	14-19	
hypersthene chondrites	84-92	22-26	77.5	20.7	1.8	47.7	7.7	44.6	84.2	10.2	5.6	4-11	
amphoterites	88-93	27-32	73.3	24.5	2.2	46.7	9.7	43.6	85.9	10.5	3.6	0.5-4	
carbonaceous chondrites	93-98	0-60(100)										0-5	
aubrites	95-98		95-100									0.5-3	
pallasites	5-55	10-20	÷	÷	÷	÷	÷	÷	÷	÷	÷	28-88	
mesosiderites	20-65(80)	14-51	50-80	18-52	2-6	22-45	15-30	40-48				30-55	
coarse octahedrites	< 0.1-16	2-5	92.2	6.9	1.9	52.3	3.0	44.7	83.4	12.5	4.1	90-92	
octahedrites, anomalous	< 0.1-15	1-8	90.0	7.9	2.1	52.6	3.5	43.9	85.3	11.4	3.3	70-92	
Kendall County	< 1	÷	98.3	1.0	0.7	54.4	0.8	44.8	50-87	12-49	0.4-1.5	91	
Kodaikanal	5	21	80.7	16.6	2.7	51.1	9.0	39.8	÷	÷	÷	85	
Mundrabilla	< 0.1	2.6	93.4	6.6		54.3	2.4	43.3	84.4	11.4	4.2	82	
Netschaëvo	13	14	84.4	14.2	1.4	49.0	5.7	45.3	82.0	13.8	4.2	76	
Tucson	3.7	0.2	99.0	0.4	0.6	54.8	0.5	44.7	0	100	0	85	
Weckerroo Station	1.8	÷	75.0	22.2	2.8	46.1	17.6	36.3	(85)	11	(4)	89	

† Silicate fraction of whole meteorite; range in mass %.

‡ Fa, mol % fayalite, Fe₂SiO₄; rest is forsterite, Mg₂SiO₄.

§ En, mol % enstatite, MgSiO₃; Fs, mol % ferrosilite, FeSiO₃; Wo, mol % wollastonite, CaSiO₃.

|| Ab, mol % albite, NaAlSi₃O₈; An, mol % anorthite, CaAl₂Si₂O₈; Or, mol % orthoclase, KAlSi₃O₈.

¶ Metallic iron content of whole meteorite; range in mass %.

Sources: Own data; Keil 1962; Massalski & Park 1964; Keil 1965; Mason 1967*b*; van Schmus & Ribbe 1968; Bunch & Fuchs 1969*a*; van Schmus 1969; Bunch *et al.* 1970; Olsen & Jarosewich 1970; Wasson & Wai 1970; Olsen & Jarosewich 1971; Powell 1971; Mason & Jarosewich 1973; Scott & Bild 1974; Rambaldi *et al.* 1974; Ramdohr *et al.* 1975; Wlotzka & Jarosewich 1975; Weigand 1975; Baedeker & Wasson 1975.

Pyroxenes

Pyroxenes (tables 15–17) comprise one-third to one-half by mass of the silicate inclusions in group I and group I–Anom and range in composition from pure enstatite to augite.

Considering the present uncertainty with respect to the nomenclature of pyroxenes – it is under review by the International Mineralogical Association – it may be appropriate to comment briefly upon the situation. Prior (1920) divided meteoritic orthopyroxenes into enstatite (< 10 mol % FeSiO₃), bronzite (10–20 mol %), and hypersthene (> 20 mol % FeSiO₃). On this basis he suggested the three major stony meteorite classes, enstatite, bronzite and hypersthene chondrites. This classification has been extremely useful, since in meteorites it is accompanied by hiatuses in both bulk chemistry and mineralogy, and so most meteoriticists adhere to the pyroxene nomenclature proposed by Prior. However, in recent years Poldervaart & Hess (1951), Ramdohr (1954, p. 576), Deer *et al.* (1962) and Strunz (1970, p. 414) have each put forward other definitions, more or less arbitrarily redefining the composition limits of bronzite. As long as the situation is ambiguous there seems to be no way out of specifying what is meant by ‘bronzite’ and ‘hypersthene’, e.g. by giving the composition in terms of mole percent enstatite (MgSiO₃, En), ferrosilite (FeSiO₃, Fs), and if necessary, wollastonite (CaSiO₃, Wo), the end members of the ternary system.

Orthopyroxenes in iron meteorites usually contain only a little CaO; they may also contain some Al, Cr, Mn, Na and Ti, all these oxides totally between 1.5 and 3.7 % by weight, according to Bunch *et al.* (1970). Kendall County and Tucson display the purest pyroxenes, while Netschaëvo’s pyroxenes are close to those of H-chondrites.

Calcium-rich clinopyroxenes are usually quite low in iron, but they contain about 1 % Cr₂O₃ wherefore they are properly termed chromediopsides. The chromediopside of Mundrabilla has as much as 1.66 % Cr₂O₃ and 0.89 % TiO₂ (Ramdohr *et al.* 1975), while that in Toluca has 0.92 % Cr₂O₃ and 0.67 % TiO₂ (Bunch *et al.* 1970). Tucson contains an exceptional diopside, almost without Cr₂O₃ and TiO₂, but with 7.3 % Al₂O₃, i.e. an augite proper (Bunch & Fuchs 1969*a*).

Unlike terrestrial pyroxenes, those of meteorites contain little or no ferric iron, because they have been rather thoroughly equilibrated with the omnipresent free metallic iron. Equally, the aluminium oxide content is usually below 1 %, aluminium having diffused into the feldspars. For this reason, augite *sensu stricto*, i.e. with Fe⁺³ and Al substituting for Mg and Si in the CaMg[Si₂O₆] molecule, occurs but rarely within the silicates; however, since the terminology is still under dispute, references to augite in the silicates of iron meteorites are not infrequent.

For further information on pyroxenes in meteorites, the reader is referred to reviews by Mason (1968, 1974), Binns (1970), Powell (1971), Wasson (1974) and Bunch & Olsen (1975). The special conditions concerning shock transformations of pyroxenes and other silicates have been reviewed by Stöffler (1972, 1974).

Feldspars

Plagioclase (tables 15–17). Feldspars are common minerals of the Earth’s crust; however, in iron meteorites they are less abundant than olivine and pyroxenes. Plagioclase is the commonest, but other feldspars may be found in minute quantities. The plagioclase of meteorites is normally an oligoclase (An_{10–30}), but some albite (An_{0–10}), bytownite (An_{70–90}) and anorthite (An_{90–100}) also occur. Almost all plagioclase crystals display low contents of K₂O

(0.2–0.8 mass %) which technically is usually reported as mole percent orthoclase (KAlSi_3O_8), compare the analyses of table 17. In some iron meteorites with silicate inclusions, plagioclase may show up to 8 mass % K_2O , but since the meteorites in question, Odessa and Toluca, are terrestrially old, the K-enrichment has probably been caused by circulating groundwater (Bunch *et al.* 1970).

Peristerite is a plagioclase of albite to oligoclase composition (An_{5-17}) which has unmixed to almost pure albite and a slightly more calcium-rich oligoclase (An_{18-30}). Since the exsolution phenomenon is pretty difficult to observe and requires good resolution on the electron microprobe, unmixing may be more common in the albite–anorthite system than previously thought. Recently Scott & Bild (1974) demonstrated peristerite-unmixing in albite grains of the unusual iron meteorite San Cristobal (group I–Anom), the unmixing having produced 3–10 μm albite blebs (An_3) in an oligoclase matrix (An_{14}).

Potash feldspar (table 15) in iron meteorites is mostly formed as the result of another unmixing effect, in a different part of the ternary albite–anorthite–orthoclase system. Albite and orthoclase form a continuous series of solid solutions at high temperatures, but decompose below 660 °C into potassium-rich (microcline) and sodium-rich (albite) phases. Depending on whether the parent feldspar crystal was rich in orthoclase or albite, the product on cooling is perthite or antiperthite, respectively. Antiperthite has been reported from Kodaikanal (Bunch & Olsen 1968) and from San Cristobal (Scott & Bild 1974). The K-rich lamellae of San Cristobal are up to $3 \times 50 \mu\text{m}$ in size and can be represented by the formula $\text{An}_{50}\text{An}_{<2}\text{Or}_{50}$.

Potash feldspar has further been reported as micron-sized blebs in Weekeroo Station and Colomera (Bunch & Olsen 1968), Mundrabilla (Ramdohr *et al.* 1975) and Campo del Cielo (Wlotzka, personal communication), but it remains a very rare mineral. The reports cited here all indicate that the mineral far from being a pure potash feldspar, might in fact better be described as a potassium-rich feldspar of wide variation, showing a $\text{Na}_2\text{O}:\text{K}_2\text{O}$ mole ratio between 0.1 and 5. This ratio, the albite:orthoclase ratio, is in the average H-chondrite about 14.

Other silicates

Ureyite (table 15) is a deep emerald-green pyroxene, first described from Toluca by Laspeyres (1897). The original name, kosmochlor, referred to the green colour of the cosmic mineral. Precise analyses and X-ray powder data were provided by Frondel & Klein (1965), who in view of the uncertainty of the old report suggested the new name ureyite. After some dispute (Neuhaus 1967; Strunz 1970, pp. 412, 585), the Commission on New Minerals and Mineral Names of the International Mineralogical Association has now ruled in favour of the name ureyite. So far, the mineral has only been reported, as 0.1–0.5 mm crystals associated with troilite and daubréelite, from the group I meteorites Cañon Diablo, Smithville and Toluca, and from the group IIA meteorites Coahuila and Hex River (Bunch & Olsen 1975, and personal communication).

Krinovite (table 15) is only known as minute, rare grains in the troilite-graphite nodules of the group I irons Cañon Diablo, Wichita County and Youndegin (Olsen & Fuchs 1968). *Richterite*, or fluorichterite (table 15) is the first amphibole to have been found in any meteorite. It is a soda tremolite, occurring as up to 1 mm long grains in the graphite nodules of Cañon Diablo and Wichita County. The original, surprising report that the molecule should contain a hydroxyl group could not be supported by later examinations. Instead it turned out that the amphibole was fully fluorinated and quite rich in chromium (0.7–3.5 % Cr_2O_3)

(Olsen *et al.* 1973; Bunch & Olsen 1975). *Roedderite* (table 15) was first reported from the enstatite chondrite Indarch (Fuchs *et al.* 1966), but subsequently found, together with richterite, in Wichita County (Olsen 1967). *Yagiite* (table 15) occurs interstitially in pyroxene grains of Colomera.

TABLE 18. MINERALS IN IRON METEORITES. VI: SOME UNEQUILIBRATED PHASES, PHASE-TRANSFORMATION PRODUCTS AND MINOR ACCESSORIES INDUCED BY REHEATING, SHOCK, OR COMBINATIONS OF THESE

	composition	remarks	references
α_2 -iron*	Fe-Ni	thermally altered kamacite	Reichenbach 1858; Buchwald 1975a
ϵ -iron*	Fe-Ni	shock-altered kamacite	Maringer & Manning 1962; Buchwald 1975a
magnetite	Fe ₃ O ₄	main component of the fusion crust	Cohen 1894; Kolomenskij & Yudin 1958
wüstite	FeO	component of the fusion crust	Kolomenskij & Yudin 1958; Marvin 1963
diamond	C, cubic	shock-altered graphite	Foote 1891; Nininger 1956; Lipschutz & Anders 1961
lonsdaleite*	C, hexagonal	shock-altered graphite	Frondel & Marvin 1967; Hannemann <i>et al.</i> 1967
awaruite	Ni ₃ Fe	shock-produced in troilite	Kullerud & El Goresy 1969
heazlewoodite	Ni ₃ S ₂	shock-produced in troilite	Kullerud & El Goresy 1969
maskelynite	glass	shock-altered plagioclase	Tschermak 1872; Engelhardt & Stöffler 1968
chalcopyrrhotite	CuFeS ₂ , FeS	quenched cubic form of cubanite	Ramdohr 1963, 1966, p. 48; 1975, p. 543
barringerite*	(Fe, Ni) ₂ P	produced by artificial reheating	Buseck 1969; Buchwald 1975a, p. 105

UNEQUILIBRATED PHASES AND VARIOUS REACTION PRODUCTS

In the preceding pages quite a few minerals have been mentioned that occur in trace quantities or have – so far – only been identified in a very few meteorites. However, the significance of a particular mineral or phase is in no way determined by its abundance. ‘In practically all meteorites the elements Fe, Mg, Si and O make up over 90 %, both in mass and in number of atoms; hence the common minerals are metallic nickel-iron, olivine and pyroxenes. A rare mineral is rare either because it contains a rare element or because it has a very limited stability field. A mineral reflecting the latter situation is of especial significance, because it may enable us to deduce the conditions of origin of the meteorite more precisely than could be established from the common minerals’ (Mason 1972a). Since some of the principal objectives of meteorite research are to determine the conditions under which these enigmatic bodies crystallized and later were divided into meteorite masses, further study of the rare minerals should be particularly rewarding.

In this perspective, the numerous unequibrated phases and metastable phase-transformation products in meteorites arouse considerable interest. In table 18 the more important forms have been compiled. Of these, the very common metallic α_2 and ϵ phases, produced respectively by rapid cooling and by rapid depressurizing, were discussed on p. 460, while the rare *diamond* and *lonsdaleite* occurrences were noted on p. 463, and *chalcopyrrhotite* on p. 468.

Although repeatedly reported as a primary mineral, the author believes that *magnetite* in iron meteorites only occurs as a late secondary product. It was either formed as a reaction product of oxygen with the surface of the ablating meteorite when it rapidly penetrated the atmosphere (see, for example, Marvin 1963; Buchwald 1975a), or it was formed during terrestrial weathering, such as reported, e.g. from Odessa by Marshall & Keil (1965) and in a

number of irons by Buchwald (1975*a*). *Wüstite* is only known as a minor component of the oxidic fusion crust on iron meteorites. According to the Fe–O equilibrium diagram, wüstite is metastable below 560 °C; it does, however, survive for long times in a terrestrial environment, just as the metastable martensite.

Awaruite and *hezlewoodite* (table 18) were reported by Kullerud & El Goresy (1969) as minor accessories from the troilite–graphite nodules of Odessa. The thorough study of the mineral associations indicated that formation and preservation had required pressures of at least 40 kbar, and relatively short thermal excursions to minimum 750 °C where hezlewoodite was present, and maximum 550 °C where awaruite was present. The entire occurrence thus suggested transient high pressures and temperatures, with steep gradients, such as may prevail under shock conditions.

Barringerite has only been described from the pallasite Ollague, where it locally forms rims 10–20 µm wide along the contacts between schreibersite and troilite. The occurrence of (Fe, Ni)₂P within a meteorite containing abundant metal is indicative of a metastable condition, induced by some thermal treatment of relatively brief duration. From similar occurrences in artificially reheated iron meteorites, the present author concludes that the barringerite of Ollague was produced by artificial reheating, probably when one of the early owners experimented with the nature of the pallasite. Incidentally, it is almost certain that Ollague is a transported fragment of the Imilac pallasite and has had a turbulent story (Buchwald 1973, 1975*a*, p. 1393; Scott & Wasson 1975). Polished sections through the Ollague pallasite unambiguously display phases from transformations that witness to artificial reheating to about 800 °C.

Maskelynite (table 18) was first described from the achondrite Shergotty and was for a long time believed to be fused glass. However, Milton & De Carli (1963) were able to produce maskelynite in shock-loading experiments; they redefined the mineral in more precise terms to mean a shock-transformed amorphous phase, produced by solid-state transformation and retaining the external features of crystalline plagioclase. Maskelynite can be regarded a special case in the group of diaplectic glasses, so abundantly formed from rocks in terrestrial and lunar cratering events. For a penetrating treatment of maskelynite, diaplectic glasses and shock-fused glasses, see Stöffler (1972, 1974).

Maskelynite is abundant in some heavily shocked chondrites (see, for example, Lange *et al.* 1974), while it is not often reported from iron meteorites. Good descriptions and analyses concern blebs and veins in Kodaikanal and Colomera (Bunch & Olsen 1968; Bence & Burnett 1969; Bunch *et al.* 1970). Slight changes in Si, Fe, K, Al and Na contents have been detected, so the shock-transformation event has apparently allowed for some diffusion and/or decomposition.

MINERALS FROM WEATHERING

Iron meteorites represent extremely reduced and anhydrous systems, displaying mixtures of troilite, graphite and hydroxyl-free minerals with native iron, carrying nickel in solid solution. They evidently have had an origin where the oxygen and water pressures were extremely low, either comparable to an ultra-high vacuum or with an excess of hydrogen. Olsen & Fuchs (1967), for example, computed that the oxygen fugacity at 1000 K had been close to 10⁻²¹ atm (10⁻¹⁶ Pa) at the time and place of formation.

Therefore, once arrived on Earth meteorites start to corrode and ultimately disappear as

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recognizable masses. How long this process takes will depend upon depth of burial, porosity and homogeneity of the soil, temperature range and average rainfall, and but little on the meteorite itself. However, the morphology of the corrosion attack may somewhat depend on the state of stress and the nature of the included minerals. In shock-hardened meteorites with hatched ϵ -structure, e.g. a stress-corrosion attack may develop, displaying fine branching networks of cracks; and in irons with shock-melted troilites, intense pitting may develop, ultimately producing deep, undercut, hemispherical holes as in Willamette and Sacramento Mountains. Although the details are far from clear, it appears that the microcrystalline troilite aggregates of iron and various unequilibrated sulphides are very sensitive to corrosion; the resulting sulphuric acid may concentrate locally and excavate the pits, leaving adjacent areas virtually unattacked.

TABLE 19. MINERALS IN IRON METEORITES. VII: PRODUCTS OF WEATHERING

	composition	crystal system	major references
akaganecite	β -FeOOH	tetragonal	Marvin 1963
cassidyite*	$\text{Ca}_2(\text{Ni, Mg})(\text{PO}_4)_2 \cdot 2\text{H}_2\text{O}$	triclinic	White <i>et al.</i> 1967
copper	Cu	cubic	El Goresy 1965; Ramdohr 1973
goethite	α -FeOOH	rhombic	Buddhue 1957; Marvin 1963
hematite	α -Fe ₂ O ₃	hexagonal	Buddhue 1957; Marvin 1963
jarosite	$\text{KFe}_3(\text{OH})_6(\text{SO}_4)_2$	hexagonal	White <i>et al.</i> 1967
lipscombite	$(\text{Fe, Mn})\text{Fe}_2(\text{OH})_2(\text{PO}_4)_2$	tetragonal	White <i>et al.</i> 1967
maghemite	γ -Fe ₂ O ₃	cubic	White <i>et al.</i> 1967
magnetite	Fe ₃ O ₄	cubic	Buddhue 1957; Marvin 1963
pentlandite	$(\text{Fe, Ni})_9\text{S}_8$	cubic	Ramdohr 1966, 1973
reevesite*	$\text{Ni}_6\text{Fe}_2(\text{OH})_{16}\text{CO}_3 \cdot 4\text{H}_2\text{O}$	hexagonal	White <i>et al.</i> 1967
sulphur	S	rhombic	Buddhue 1957, p. 53

In an extensive study of the oxidation of meteorites, Buddhue (1957) reported a large number of secondary minerals, goethite and magnetite (table 19) being the most important. He also noted the difficulty in precisely defining the minerals. He found the corroded crust to be rather complex and unequilibrated, partly dehydrated and with more or less NiO in the molecules of the component oxides. Characteristically, the oxides were found to contain 7–15 mass % total water (hygroscopic plus combined), but there were not too many good analyses in the literature.

After the additional works by Marvin (1963), Marshall & Keil (1965) and White *et al.* (1967), the data by Buddhue (1957) may best be interpreted as showing that the disintegration products mainly consist of goethite with small amounts of NiO (about 1 %) in solid solution, and of maghemite with significantly more NiO (up to about 9 %). In addition, substantial amounts of magnetite, i.e. a not yet fully oxidized iron oxide, are present. The specific gravity of typically massive oxides from iron meteorites is 3.7–4.2, i.e. somewhat below that of the pure oxides. This is no doubt due to the porous, microcracked nature of the corrosion products. The magnetic character of the oxidation products is essentially due to the presence of maghemite and magnetite, to a very minor degree also to inclusions of imperfectly corroded metallic blebs.

The other minerals of table 19 are also well documented, but occur only in very limited quantities, or, rather, because of their accessory nature and minor interest, they have only been reported from a few intensively studied meteorites.

Finally, the following products of weathering have been reported as very minor occurrences; they are not all equally well substantiated: apatite, bornite, chalcopyrite, collinsite, heazlewoodite, lepidocrocite, opal, siderite, trevorite, violarite and vivianite. On a specimen of Cape York, recovered from the midden of an Eskimo settlement, the author identified a crust of vivianite, probably formed in a reaction between the meteorite and the disintegrating bones of the midden (Buchwald 1975 *a*, p. 425).

On the recognition of old iron meteorites

Several iron meteorites have been recovered from deposits of a considerable geological age. For example, Susuman and Elga were recovered from auriferous gravels at respectively 32 and 20 m depth. The meteorites were, however, surprisingly well preserved, possibly because of a low oxygen pressure and a low average temperature at the place of deposition.

Other irons have been almost entirely converted to oxide shale, for example numerous masses of Cañon Diablo and Wolf Creek, or to loosely adhering masses, that partially or wholly disintegrated when they were excavated and removed to the museum, for example Sardis, Ider, Lucky Hill and Dorrigo. The recognition of the meteoritic character poses few problems here, because the Widmanstätten pattern usually is still visible, the corroded fragments even falling apart along octahedral planes.

Ultimately we may be confronted with such completely oxidized material as Coldwater from Kansas or Monturaqui from Chile. The identification as a meteorite, which in some cases, despite the corroded character, can lead to full classification, rests in these cases upon (i) the recognition in polished sections of tiny inclusions of unaltered meteoritic minerals, (ii) the presence of NiO in the bulk oxide, and (iii) the recognition of fossil traces of the Widmanstätten structure (about 80 % of all iron meteorites display Widmanstätten structure).

The mineral most resistant to terrestrial corrosion is schreibersite. Even when everything else is converted to oxide, tiny fragments of the phosphides may still be identified optically and analysed by the electron microprobe. Chromite and daubréelite are also quite resistant, but occur only in limited quantities. Somewhat less resistant are high-nickel taenite rims, monocrystalline troilite and cohenite. The martensitic part of plessite fields, although disappearing rather early, leaves a very characteristic microscopic ghost-pattern in the oxides and helps to verify the other conclusions. Troilite nodules that have been exposed to shock-reheating are, in contrast, very prone to oxidation because the finely disseminated iron and sulphide particles constitute a microcracked galvanic cell easily accessible to terrestrial water.

In his survey of terrestrial oxidation, Buddhue (1957) was perhaps too uncritical. If he had selected only the better analyses, among them his own, he might have seen that most iron oxide shale was somewhat depleted in nickel, relative to the Fe–Ni ratio of the unweathered meteorite. He would also have noted that *some* NiO evidently was always retained in the oxides, even in such completely weathered irons as Coldwater (4.6 % NiO). In the iron shale of Cañon Diablo, recalculation of Buddhue's data suggests that the Fe–Ni ratio of the oxides is the same as that of the coarse octahedrite itself.

Iron in aerated solutions will soon oxidize to Fe⁺³ and precipitate as ferric hydroxide; very often some nickel will coprecipitate with the iron as nickelous hydroxide. These mixed hydroxides in time dehydrate and age to the characteristic minerals. The nickel-bearing solutions may also attack adjacent troilite and in a replacement process form pentlandite (Fe,Ni)₉S₈. The pentlandite crust may be 5–20 μm thick and will typically be located along cracks in the

troilite, e.g. in Hoba, Odessa and Cape York. Sometimes, in severely weathered meteorites, the troilite is almost wholly converted to pentlandite and violarite, e.g. in Santa Catharina and Twin City. The remaining nickel is quite stable in solution and may be removed as Ni^{+2} , to be found in the soil up to a few metres distance from the corroding mass (Buddhue 1957, p. 132). On the other hand, much oxide can also be formed in situ in a replacement process, so that very little iron and nickel is carried away, and the original Fe–Ni ratio is retained. This process is perhaps particularly common in arid places.

In other words, even very long exposure – 50 000 years or more – to the terrestrial environment need not erase the meteoritic character of massive irons entirely. As far as is known today, the oxides will always retain some nickel; much of the Cañon Diablo shale has even retained the original Fe–Ni ratio. The lowest value of any meteoritic oxide shale seems to be that of Brenham, with only about 2 % NiO; this came from a mass excavated from a pond filled with water during at least part of the year, so that conditions were unusually conducive to rusting and leaching (Monnig 1948; Buddhue 1957).

TABLE 20. MINERALS IN IRON METEORITES.
VIII: DISCREDITED OR WITH AN UNCERTAIN POSITION, 1976

	reported by:	discredited by:	remarks
chalypite*	Shepard 1867	Buchwald 1975 <i>a</i> , p. 113	misinterpreted carbide
cliftonite	Fletcher 1887	Brett & Higgins 1967	graphite aggregate
gentnerite	El Goresy & Ottemann 1966	I.M.A. 1967	inadequate identification
gold	Edwards 1953	Mason 1967 <i>a</i> , p. 320	inadequate identification
kosmochlor*	Laspeyres 1897	Fronzel & Klein 1965	identical with ureyite
lawrencite	Smith 1855; Daubrée 1877	Buchwald 1975 <i>a</i> , p. 113	terrestrial contamination
merrillite*	Wherry 1917	Fuchs 1962	identical with whitlockite
moissanite†	Moissan 1904; Kunz 1905	Mason 1967 <i>a</i> , p. 320	contamination by SiC
nickel serpentine	White <i>et al.</i> 1967		more data needed
perryite	Fredriksson & Henderson 1965	Buchwald 1975 <i>a</i> , p. 113	more data needed
valleriite	Ramdohr 1963	Evans <i>et al.</i> 1964	misinterpreted mackinawite
W-carbide	Scott & Agrell 1971		more data needed
weinbergerite*	Berwerth 1906	Olsen & Mueller 1964	mixture of feldspar and pyroxene
zaraitite	Buddhue 1957, p. 85	Mason 1967 <i>a</i> , p. 323; Strunz 1970, p. 247	inadequate identification

DISCREDITED MINERALS AND MINERALS WITH AN UNCERTAIN POSITION

As for 1876, we may also prepare a list in 1976 of dubious minerals (see table 20).

Chalypite was reported from the Ovifak irons, Greenland. These masses of up to 30 t were at the time supposed to be iron meteorites. The mineral was assumed to be Fe_2C and it is still

so listed by Strunz (1970, p. 97). No doubt, the type occurrence was, in fact, cohenite, Fe_3C with some nickel; the old report must have been based upon inferior analyses and misinterpretations. The mineral chalypite should be deleted.

Nickel serpentine was reported as a product of weathering in the iron meteorite Wolf Creek by White *et al.* (1967). The authors inferred that the numerous previous reports, e.g. by Buddhue (1957), of bright yellowish-green grains of zaratite, a nickel carbonate, should in fact be interpreted as nickel serpentine, the analogue of clino-chrysotile. Apparently, the definitive work on this mineral controversy has not yet appeared.

Lawrencite

Lawrencite requires a separate discussion. The ferrous chloride, FeCl_2 , has been called lawrencite and has only been reported from meteorites. Described by various authors and then named by Daubrée (1877), this cosmic ghost now is celebrating its centenary. There have been very few good reports on lawrencite; most workers, after having noted the exudation of greenish droplets on freshly prepared sections through meteorites, have jumped to the conclusion that it is the cosmic mineral lawrencite that has started its decomposition. In one serious report (Keil 1968), a tiny grain of lawrencite was verified in an enstatite chondrite. On the basis of this paper, lawrencite might be retained as a genuine mineral intrinsic to one, not too common class of stone meteorites.

However, it is in iron meteorites that lawrencite has always stirred the imagination. Cohen (1897), Buddhue (1957), Mason (1962), Marvin (1963), Reed (1963) and White *et al.* (1967) discussed the corrosion of iron meteorites on the assumption that lawrencite was present in substantial amounts, and Olsen & Fuchs (1967) and Bloch & Müller (1973) discussed the formation of iron meteorites on a similar background.

The alarming rate with which some irons disintegrate when moved from the field to the museum has invariably been ascribed to a critical amount of lawrencite (Perry 1944, p. 25; Chalmers 1948). When the Sardis meteorite was excavated, a fragment was found that weighed about 2.3 kg. Invasive oxidation in this case had produced a thick coating of oxide not entirely unlike a shale ball. This specimen disintegrated completely in about three months (Buddhue 1957). The present author noted a similar case when excavating the 20 t Cape York meteorite in 1963 and a 390 g fragment spalled off the lower side (figure 10, plate 2). It was considered a rather good specimen, despite its cracked surface, and rested on the author's desk for a couple of months, before it started to disintegrate. Before a year was over, the massive sample was nothing but a heap of loose scale and dust, the largest fragment being about 1 cm in diameter. It is estimated that the Cape York meteorites fell at least 1000 years ago, and probably much earlier; the violent change in behaviour on moving is therefore in need of an explanation.

The author has searched for lawrencite in this material and in all other iron meteorites where it allegedly occurred, but with negative result. However, chloride was always detected in qualitative tests, and sometimes minor chlorine-bearing accessories, such as chlorapatite and djerfisherite, were also noted.

There is a remarkably good correlation between lawrencite reports and the terrestrial age of meteorites. Never was lawrencite identified in a fresh fall. It has unfailingly been reported from such meteorites which from other evidence may be judged to have been well buried for many thousands of years. This has always been interpreted as *because* of the cosmic lawrencite content the meteorite disintegrated, such as Cranbourne now protected in a glass case in the

British Museum. But it is no doubt the other way around: Because the meteorite has rested for a long time in terrestrial saline soil before being found, chloride from circulating groundwater has invaded it. Once there, it is like a pest, chlorine almost acting as a catalyst and serving to accentuate the attack on the meteorite, even after its removal to a museum.

Likewise, scrutinizing the numerous reports, the lawrencite is always said to be in near-surface parts of the weathering meteorite, often associated with troilite inclusions, with cracks along the Widmanstätten structure or with grain boundaries in polycrystalline hexahedrites, such as Forsyth County. However, in the author's opinion, it is quite a coincidence that lawrencite should be concentrated along the rims of a meteorite, since the final shape of the meteorite was first determined in the atmosphere.

There seems to be no way of escaping the conclusion that iron meteorites contain very little primordial chlorine, possibly in the order of 10 parts/10⁶ (Berkey & Fisher 1967; Hintenberger, personal communication), and that this is partitioned between the grain boundaries of the metal and a few inclusions, notably chlorapatite and djerfisherite. The much higher amounts sometimes reported, for example 0.17 % Cl in Forsyth County (Cohen 1897), must be due to chloride introduced with aggressive terrestrial groundwater while the meteorite for countless generations was buried in the soil.

The difference between the state of preservation of masses belonging to the same fall, for example Brenham and Cañon Diablo, has surprised many observers; they have normally been forced to conclude that lawrencite was very heterogeneously distributed over the parent meteorite mass. This is of course in conflict with the other observation that it should be located along the exterior rim zone. When it is first realized that lawrencite was not present at all, there remains to emphasize that the different states of corrosion were conditioned – for the same fall – mainly by degree of burial of individual masses, permeability and chemistry of the local soil, and nature and quantity of cracks in the exposed meteorite. Masses which came to rest on top of a rocky surface thus have had a much better chance of surviving than partly or fully buried masses. All so-called shale balls from Meteor Crater, i.e. totally converted Cañon Diablo masses, have for example been found at grass-root depths.

The important thing about lawrencite is thus not its possible addition to our mineral inventory, but that we may be able to formulate our thoughts about formation and, especially, corrosion of iron meteorites quite differently when we realize that lawrencite is a cosmic ghost; that virtually no chlorine is indigenous to the irons, the bulk having been introduced into the material during a long sojourn in terrestrial soil.

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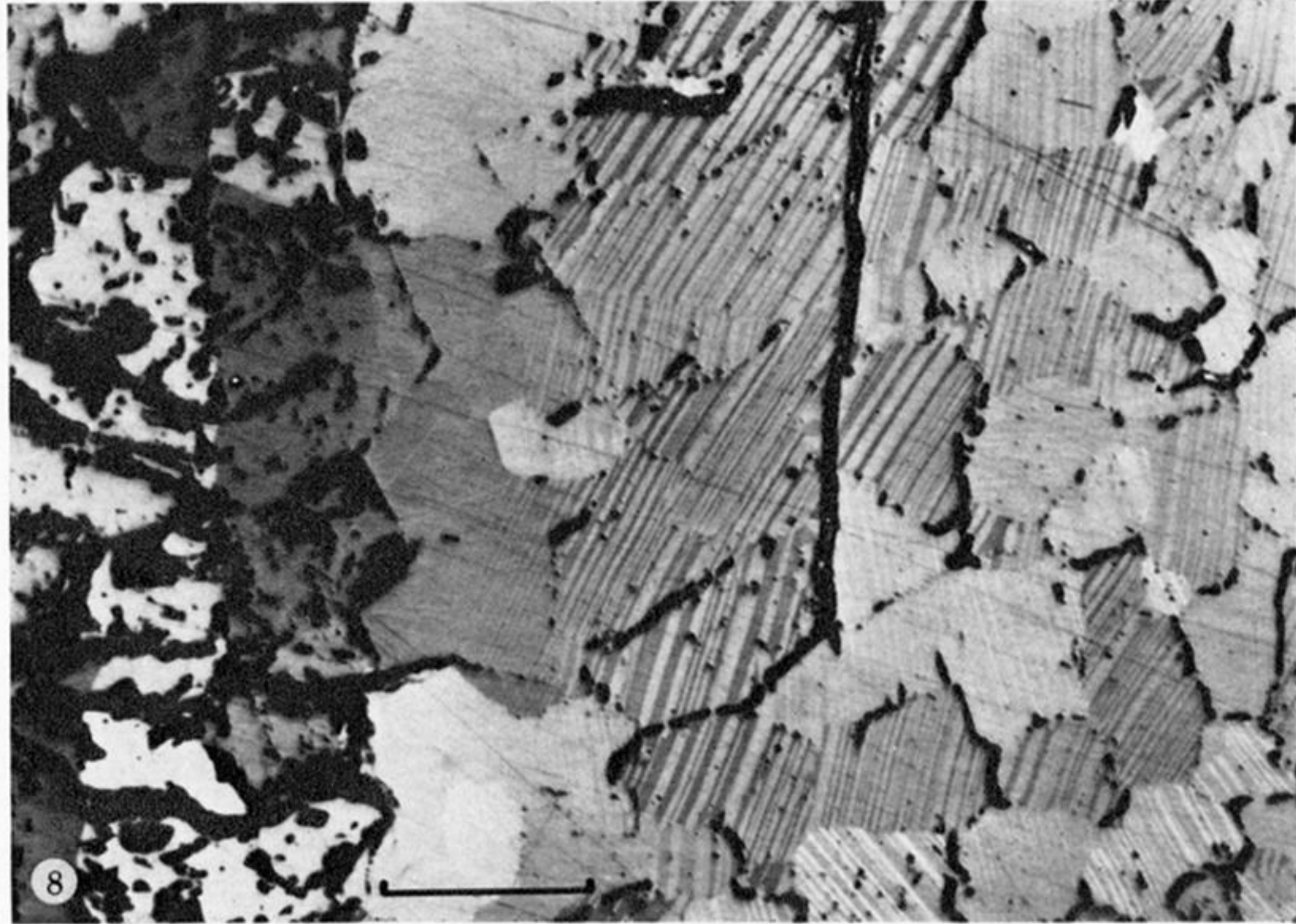
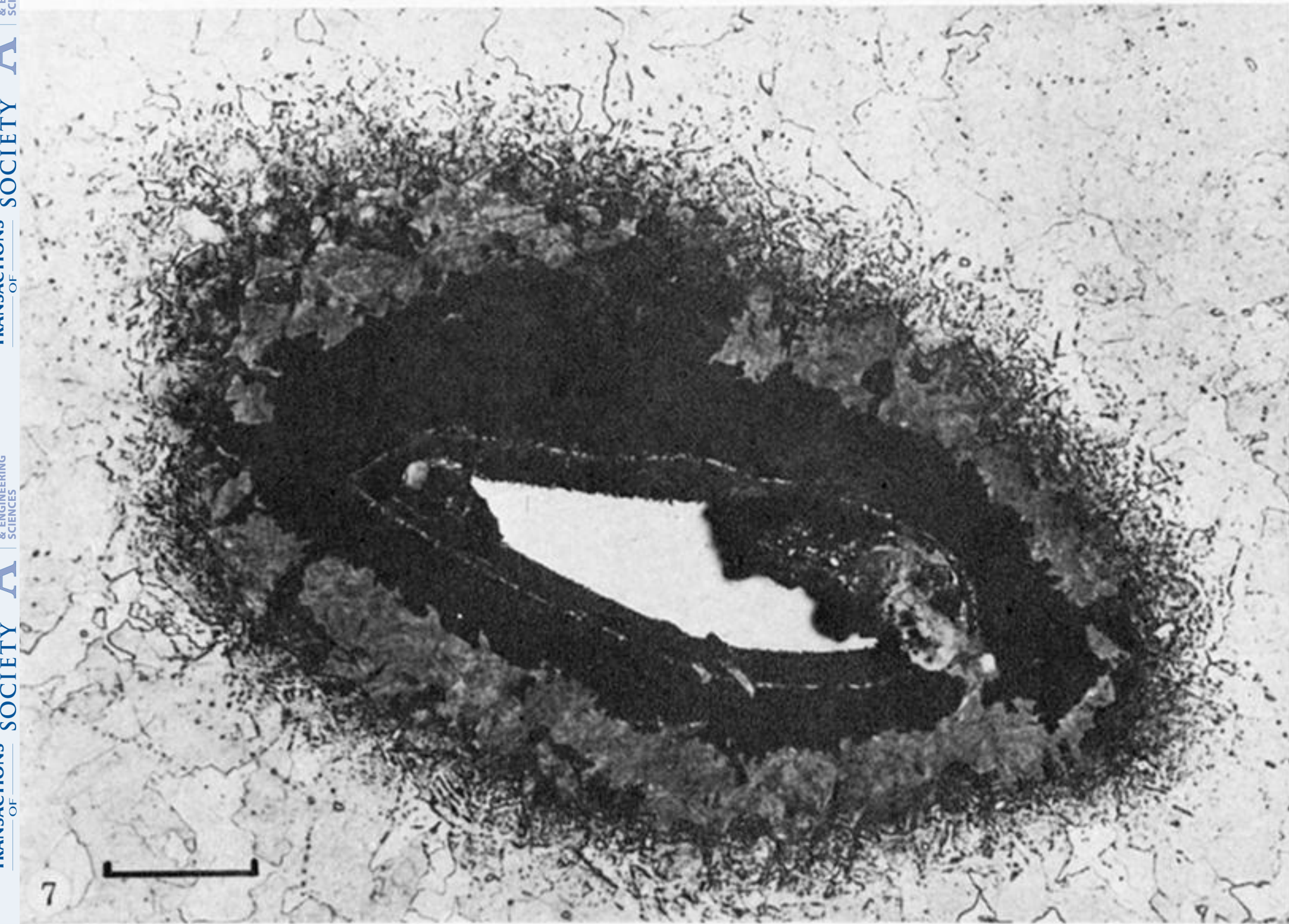
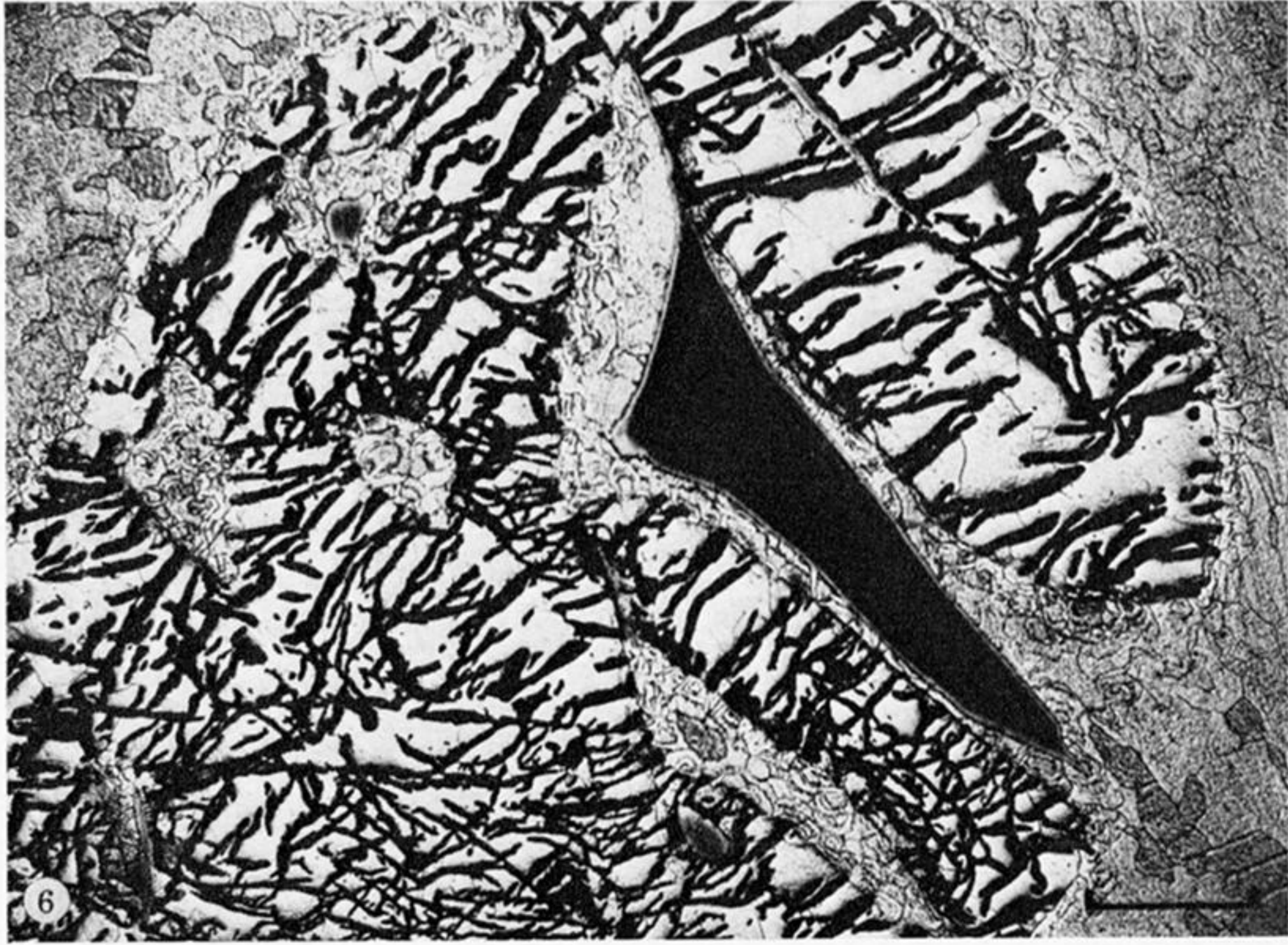
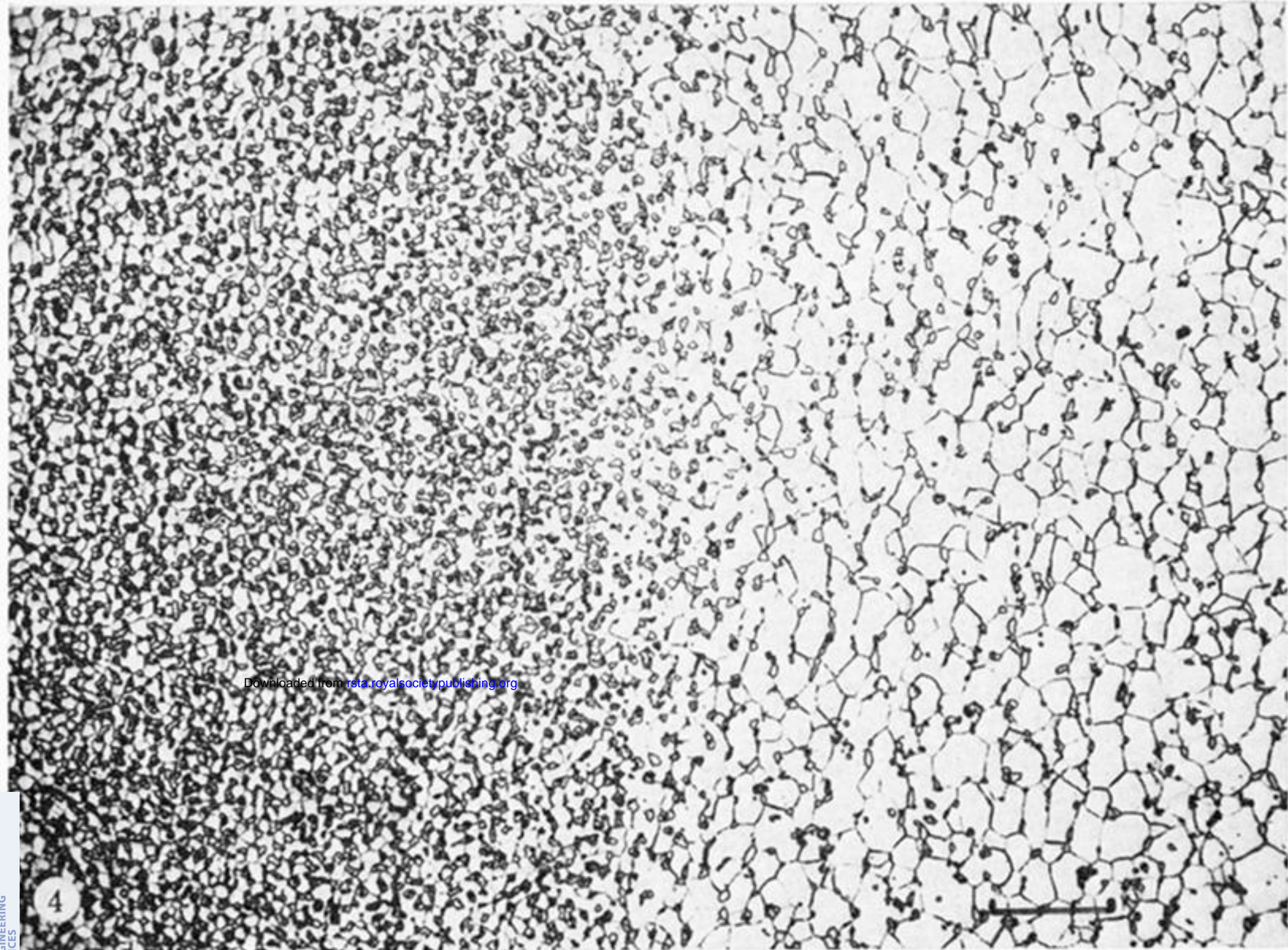
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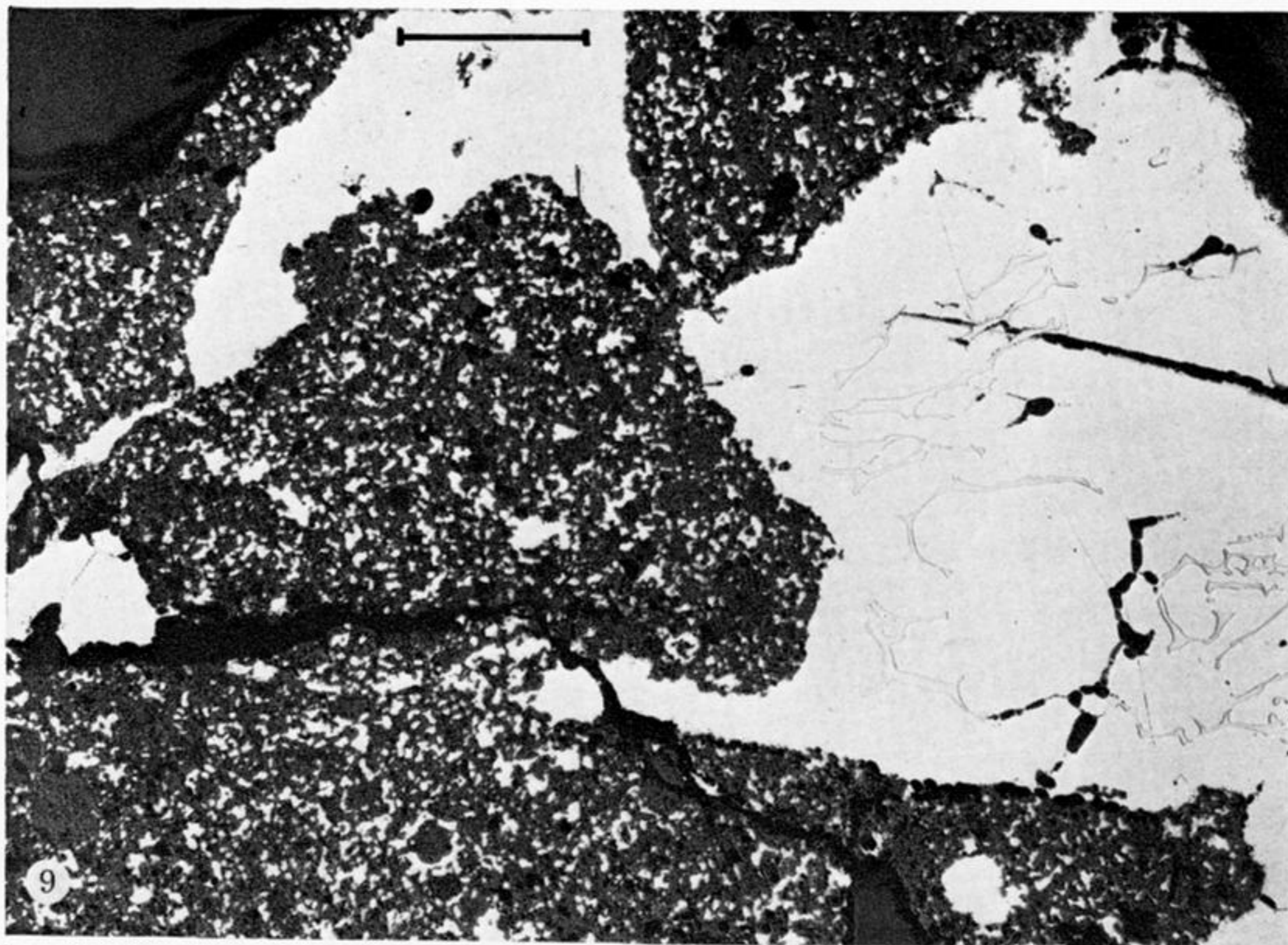
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FIGURES 4 AND 6–8. For description see opposite.



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FIGURE 9. Netschaevo, Russia (Brit. Mus. No. 33, 953). An anomalous medium octahedrite with angular silicate inclusions. The inclusions are chondritic and display nickel-iron, olivine and enstatite as major, and plagioclase and chromite as minor minerals. The bulk metal has 8.46% Ni and Ga-Ge-Ir contents that place Netschaevo outside the chemical groups. Scale bar 3 mm.

FIGURE 10. A corroded surface fragment (390 g) of the medium octahedrite Cape York (Agpalilik), Greenland (Copenhagen No. 1967, 410). While the upper part of the 20-ton mass was well preserved, such parts that had been in contact with the soil were corroded. The average density of this sample, when taken in 1963, was 5.0, but on removal to my office in Copenhagen it rapidly disintegrated and was after 12 months nothing but a loose heap of scale and dust. Scale in centimetres.