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INTERPRETATION OF ORE TEXTURES

BY
EDSON S. BASTIN
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PREFACE

Since the happy days of apprenticeship in Economic Geology on the U. S. Geological Survey when under the inspiring guidance of Waldemar Lindgren and Frederick L. Ransome he first acquired an interest in the application of geology to mining, the writer has been especially fascinated by the interpretation of ore textures under the microscope. Through twenty-five years of subsequent teaching of Economic Geology in the laboratory at The University of Chicago this interest has increased, as he has realized more and more fully how much the critical interpretation of the textures of ores as seen under the microscope contributes to a sound diagnosis of the origin of the ore deposits.

In the study and interpretation of microscopic ore textures, as in most relatively new fields, a certain period of seasoning must be passed through before sufficient agreement develops among the research workers to warrant the preparation of a text. In the present case the writer believes that season has been sufficient and that the needs of students warrant the presentation in a brief book of the more salient information on ore textures now mainly scattered through the technical journals and governmental reports. Widespread agreement does not yet exist with respect to the significance of all ore textures. The book is intended to be not a manual but a laboratory aid to graduate and advanced undergraduate students in their microscopic studies of ores, and the writer hopes it may also be of aid to research workers.

In this as in most textbooks the function of the author is to an important degree that of a critical editor in dealing with the contributions of a host of fellow workers. In a very real sense therefore this book has many authors; as is natural, however, the writer has drawn heavily upon his own experience. In 1931 a committee on Mineral Paragenesis of the National Research Council prepared a report (Bastin *et al.*, 1931) that has been of particular aid in the preparation of this volume. To the other members of this committee, L. C. Graton, the late Waldemar Lindgren, W. H. Newhouse, G. M. Schwartz, and M. N. Short, the writer acknowledges his especial indebtedness.

The writer has depended as heavily upon illustrations as upon words having in mind a recent remark of President Stoddard of the University of Illinois: "A Principle is not really understood in the absence of examples."

Pioneer efforts in a field such as this are sure to contain errors and omissions. The writer hopes his colleagues in the field of Ore Deposits will freely offer suggestions that may make any later book on this subject more nearly complete.

During the first stages of critical editing of this book there has appeared from the other side of the world a book entitled *Textures of the ore minerals and their significance* by Dr. A. B. Edwards of the University of Melbourne, Australia (1947). The plan of organization of Edwards' excellent book is sufficiently different from that of the present book so that the two volumes should supplement each other usefully without undue duplications or conflicts of viewpoint.

EDSON S. BASTIN

Ithaca, New York
February 1, 1950.

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INTRODUCTION

The term *texture* as used here refers to the smaller features of ores which are dependent on the size, forms, and arrangement of the component minerals and in some cases upon such features as mineral cleavage, mineral parting, mineral contacts, and minor fracturing. These features can be observed in hand specimens either with the unaided eye or with the microscope. For those larger features of ore deposits that are dependent upon folding, faulting, jointing, bedding, etc., the term structure is applied. Since the distinction between texture and structure is in important part one of size, the writer reserves some latitude in the descriptions. This usage of the terms texture and structure conforms to that commonly adopted by petrographers, and this is desirable since ores are merely unusual rock types.

Some textural terms well established among petrographers, are also appropriately applicable to ore textures and have been widely used in ore descriptions. Petrographic terms used in describing the shapes of crystals are:

Idiomorphic or *euhedral* if the crystal contours are more or less perfectly developed.

Xenomorphic or *anhedral* if crystal contours are not developed.

Hypidiomorphic or *subhedral* if crystal contours are partially developed.

In all interpretations of ore textures it is important that the texture be visualized in three dimensions, although a single thin section or polished section displays only two dimensions. In magmatic ores of granular texture it is usually safe to assume that the third dimension is similar to the other two, but in magmatic ores showing either graphic or flow textures several sections cut in diverse directions may be needed for complete textural interpretation. In ores of pneumatolytic and hydrothermal origins directional differences in texture may be considerable and significant. For example, in one section mineral B may appear to be a simple inclusion in A, but when the third dimension is studied the "inclusion" may be found to represent a prong from a large area of mineral B projecting into mineral A.

The techniques of the preparation of polished specimens of ores and of the identification of metallic minerals in polished specimens by etching and other tests are basic to the interpretation of ore textures. In this country these techniques have been described in an excellent text (Short, 1940) which also includes a section on microchemical methods. In German there is the comprehensive and profusely illustrated text of Schneiderhöhn and Ramdohr (1931), which deals also in part with the interpretation of ore textures.

In Chapter I the Magmatic Ores are considered separately because they are a variant of the great class of Igneous Rocks. The textural terms established by petrographers for the commoner igneous rocks and applicable also to magmatic ores form an appropriate background for the consideragion of textures formed by solutions that were *not* magmas though many of them were hot. Adherence to the widely accepted distinction between magmas and hydrothermal solutions does not negate the possibility of transitional relations between them. It *does* emphasize certain significant contrasts between most magmas and most hydrothermal solutions. These contrasts are particularly great between hydrothermal solutions and the basic or

ultrabasic magmas with which most magmatic ores are associated. The two most significant features of hydrothermal alteration are (1) the intimacy with which the altering solutions have penetrated the rocks, often for long distances, and (2) the hydrous nature of many of the new minerals formed by their action. From this it is logical to infer that the solutions were highly mobile or tenuous and were highly aqueous. Magmas do not penetrate the rocks they intrude with an intimacy at all comparable, and there is ample evidence that the basic and ultrabasic magmas with which most magmatic ores are associated are relatively poor in water.

Ores precipitated from hydrothermal solutions may be deposited (1) by precipitation in pre-existing open spaces, (2) by replacement, or (3) by a combination of these two processes.

CHAPTER I. TEXTURES OF MAGMATIC ORES

GENERAL CONSIDERATIONS

Magmatic deposits are unique among ore deposits in being direct crystallizations from magmas, usually deep-seated (plutonic), rarely volcanic. It has long been recognized that magmas are complex mutual solutions and obey the laws of solutions (Grout, 1932, p. 209–210. Some geologists (Spurr, 1923) have emphasized the supposed igneous relations of the tenuous and highly aqueous solutions from which many ores were deposited by terming them "Ore Magmas." It seems more useful, however, to stress the *contrasts* between a magma, such as a lava flow, and a hot tenuous mineralizing solution that may penetrate a body of solid rock for tens or hundreds of feet replacing some or all of its minerals by new ones. Most magmatic ores are differentiation products from *basic* magmas, and, as these are characteristically poor in volatile components, we should expect that pneumatolytic and hydrothermal processes would be lacking or rare in association with such deposits. This appears to be the case.

Ores can be proved to be magmatic only if it can be shown that their valuable metallic minerals were direct crystallizations from magmas and not later—perhaps much later—replacements of *solid* igneous rocks accomplished by solutions from some external source. In such diagnosis larger fields relations may be more valuable than small textural features. Most magmatic ores occur *within* bodies of undoubted igneous rock and show transitions one to the other.[1] The microscope is usually of great aid in confirming such transitions. In any one district magmatic ores are usually restricted to a single type of igneous rock or at most to closely similar igneous types. They may occur in the peripheral or in the central portions or may be scattered through the mass. The mineral composition of the ore usually aids in diagnosis, for worldwide observation has shown that most magmatic ores are valuable mainly as sources of iron, nickel, copper, chromium, titanium, and the platinum group. Negative criteria, such as the absence or rarity of hydrothermal alteration and other replacement phenomena, may be of confirmatory value but must be used with caution since certain high-temperature replacements may take place without destroying the fresh "unaltered" megascopic appearance of the rock.

Ore textures seen under the microscope are seldom conclusively diagnostic of magmatic origin but may have confirmatory value. In a few ores microscopic study shows that the metallic minerals are essentially contemporaneous with certain nonmetallic minerals such as olivine that are known to be magmatic. In the chromite ore for example, shown in Figure 1 of Plate 1, some chromite grains (black) show automorphic crystal faces against olivine (or its alteration rim of serpentine), but some of the olivine also shows automorphic outlines against chromite. This reversible relation indicates that chromite and olivine are roughly contemporaneous. As the olivine is magmatic the chromite is also.

[1] Less commonly dikelike masses of ore that are believed to be magmatic traverse, with sharp contacts, rocks that may or may not be genetically related to the ore, as in the case of some magnetite dikes.

While the metallic minerals are, in some magmatic ores, contemporaneous with the silicates, much more commonly they are either earlier or later. Figure 2 of Plate 1 shows an ore in which euhedral chromite is enclosed in olivine (partly altered to serpentine). Such a texture might be due to automorphic hydrothermal replacement of an olivine rock by chromite analogous to the replacement of schist by pyrite cubes shown in Figure 4 of Plate 8. If such replacement were of the diffuse type it would be difficult to rule out this possibility on textural grounds alone. Commonly, however, automorphic replacements betray their replacement origin by being controlled in their distribution by textural features of the host such as fractures, mineral cleavages, and mineral contacts. In the absence of specific evidences of replacement the alternative is that the chromite is magmatic crystallizing first and then being enclosed by the characteristically magmatic mineral olivine. This is here the preferred explanation and is harmonious with the fact that only in a few rare occurrences has chromite been shown to be hydrothermal (Sampson, 1931; Ross, 1929).

In other ores such as the sulphide-bearing diabase (Pl. 2, fig. 1) the metallic components are the youngest and form the matrix between the silicates. The silicates are unaltered, and the metallic minerals—pyrrhotite, chalcopyrite, and magnetite—fill the interspaces around the plagioclase laths and clearly formed last. These relations might be interpreted as the result of replacement of the silicates by the metallics through hydrothermal solutions from an outside source. However, the euhedral outlines of the silicates against the metallics indicate that such replacements did not take place. The alternative explanation, entertained by Schwartz (1925), is that the metallics like the silicates crystallized from the magma.

COMMON TEXTURAL TERMS

As magmatic ores are igneous rocks, in general the accepted textural terms developed for the description of the common igneous rocks may be applied also to magmatic ores. Only in a few instances does the great abundance of metallic minerals lead to textures not found in igneous rocks poor in metallics. Familiar textures of ordinary igneous *rocks* (Johannsen, 1931) that have also been recognized in magmatic ores are:

(1) Granular
    (a) Xenomorphic
    (b) Hypautomorphic
        Poikilitic and Ophitic
(2) Porphyritic
(3) Micrographic
(4) Banded textures
    Flow banding (fluidal or fluxion)
    Banding due to gravitative settling
(5) Orbicular and nodular textures
(6) Synneusis

Xenomorphic (or Allotriomorphic) *granular*. Crystal grains are without facets and tend to be fairly uniform in size. This texture is seldom well developed in mag-

matic ores but is approximated in the chromite ore shown in Figure 1 of Plate 1 in which olivine and chromite indent each other and in which crystal facets are rare and rounded grain outlines prevail.

Hypautomorphic (or Hypidiomorphic) *granular*. Some of the crystals show facets. The chromite ore pictured in Figure 2 of Plate 1 well illustrates this texture, as does also the sulphide diabase shown in Figure 1 of Plate 2.

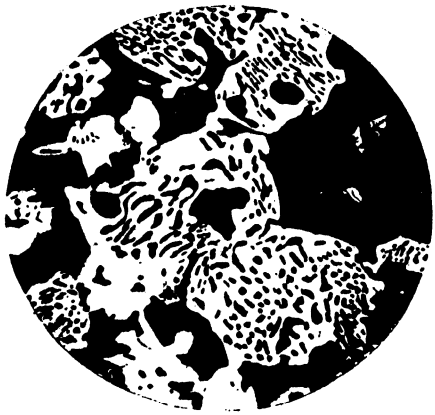

FIGURE 1.—*Micrographic intergrowth of olivine (white) and titanomagnetite (black)* From Foskaa, Norway. × 20. After Vogt.

Poikilitic granular. Optically continuous crystals of one mineral form the matrix for diversely oriented crystals of another mineral.

Ophitic granular. A variety of poikilitic texture in which the diversely oriented crystals are plagioclase laths.

Porphyritic. Large automorphic crystals (phenocrysts) of one or more minerals in a finer groundmass; in ores the groundmass appears to be always crystalline.

Graphic and Micrographic. Two minerals mutually so intergrown as to resemble ancient cuneiform writing. Some micrographic intergrowths tend to radiate from a common center. In igneous rocks the minerals of such intergrowths are usually regarded as contemporaneous. Micrographic intergrowths of olivine and titanomagnetite are shown in Figure 1, and a micrographic intergrowth of platinum and osmium is shown in Figure 3 of Plate 1. Both are unquestionably magmatic.

Banded textures. Banding due to dynamometamorphism subsequent to ore deposition will be considered in a later section but banding is fairly common in ores unaltered by metamorphism, particularly in ores of iron and chromium. In many cases such banding is clearly the result of flowage in the magma during crystallization. An excellent example is given by Takeo Kato (1921) from the Akaishi mine in Japan in Figure 5 of Plate 1. In this ore layers of chromite alternate with olivine. The microscope reveals that many of the crystals in the chromite layers have been crushed along their borders but that the crushed-off grains have also been corroded. Kato

concludes that "the banded ore was formed by a flowing motion of the crystallizing dunite magma, in which settling of the chromite crystals was going on." Banded textures due to flowage are also well developed in the magnetite ores of Kiruna, Sweden.

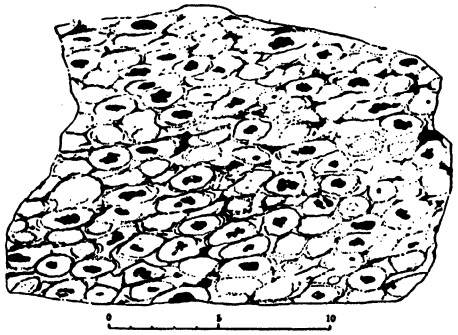

FIGURE 2.—*Orbicular chromite*

Tracing from a polished slab. Octopus mine, Siskiyou County, Calif. Black is chromite. White is olivine partly altered to serpentine. Scale in centimeters. After W. D. Johnston, Jr.

In some chromite ores individual bands are so straight and persistent for long distances that the rocks at first sight resemble sediments. Johnston and de Souza (1943) have described such ores from Brazil (Pl. 1, fig. 4). At one mine 93 layers of chromite occur in a thickness of 20 meters. The chromite bands range from a few millimeters to a meter or more in thickness, and there is little or no regularity in the thickness or spacing of the bands such as is found in diffusion banding or in varves. At some of the Brazilian mines the bands are horizontal; elsewhere, as a result of faulting, they are steeply inclined or even vertical. Chromite bands never cut across olivine bands or vice versa. Johnston and de Souza attribute the banding to variable gravitative settling of both olivine and chromite crystals during cooling and crystallization of the sill-like bodies of magma. Under this concept all of the bands were originally nearly horizontal. Many of the chromite deposits of the Bushveld complex in the Transvaal are also flat-lying and bedlike though clearly magmatic. Banded chromite ores have also been described from Siskiyou County, California (Johnston, 1936, p. 417–427).

Orbicular and Nodular Textures. Textures resembling the orbicular forms occasionally found in the commoner types of igneous rocks have been described (Johnston, 1936, p. 418–426) in chromite ores from Siskiyou County, California (Fig. 2). Johnston reports (p. 418–419)

"The orbicular ore is made up of many spheroidal units, each of which consists of a nucleus of chromite surrounded by a shell with drawn-out pointed ends, composed, in the main, of partially serpentinized olivine. Each nucleus is composed of a number of chromite crystals. . . . Each sheath of olivine is surrounded by a thin outer sheath of chromite and commonly contains several concentric layers of chromite. Larger masses of chromite occupy the interstices between abutting units. In a few units, the chromite nucleus is absent. . . . The drawn-out spheroids are fitted together in parallel orientation. The elongation of the orbicular ore is linear and not platy, for cross-sections normal to the long axis of the units show only circular patterns."

The parallel elongation of the spheroids is attributed to flowage before crystallization was complete.

In the orbicules chromite nuclei appear to have acted as growth centers for olivine, and the interstitial chromite followed the olivine crystallization.

Nodular structures are much more common than orbicular structures and have been described in chromite ores from Oregon (Fig. 3), California, Cuba, and Quebec.

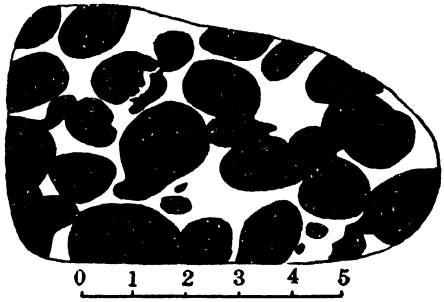

FIGURE 3.—*Nodular chromite ore*

Briggs Creek, Josephine County, Oregon. White matrix is serpentine. Scale in centimeters. After Diller and Johnston.

These nodules have been interpreted as nuclei to which chromite was attracted during crystallization.

Synneusis and Chain Structure. Synneusis (Vogt, 1921) is peculiar to igneous rocks and to the magmatic stage in their development. Vogt (p. 321) states:

> "The individuals of a mineral, segregated from a magma at an early stage, frequently swam together to assemblings or aggregates, the result of which is a structure, for which I propose the term together-swimming structure or synneusis structure."

Vogt observed the texture not only in ores but in slags. In ores synneusis is a useful criterion of magmatic origin and moreover the mineral involved is *early* magmatic. Vogt's illustration of synneusis texture is reproduced as Figure 3 of Plate 2. On textural evidence alone this relationship might be explained as an automorphic replacement of olivine by chromite. However, fractures in the olivine and contacts between olivine grains exerted no control on the chromite distribution. It is more likely that olivine was still liquid when the chromite crystallized as Vogt concluded.

The chain texture described by Sampson (1932, p. 130) from chromite ores in South Africa (Fig. 4) may be a variety of synneusis texture. Sampson does not reach a final conclusion as to its origin. The prevailingly automorphic outlines of the chromite plus the fact that in some specimens the chromite grains and chains have clearly been corroded seem to show that the chromite is older than the enclosing minerals.

ORDER OF CRYSTALLIZATION

It should thus be clear that in ores that are products of normal crystallization from magmatic liquid solutions it is commonly possible to infer the order of crystal-

lization of the several minerals from their textural relations. The principal criteria are (1) degree of automorphism of various minerals, (2) inclusions, (3) graphic intergrowths. (4) Synneusis texture, though rare, may also be of value. If the rock texture is xenomorphic granular (Fig. 1) the minerals are interpreted as essentially contemporaneous. In magmatic ores showing graphic or micrographic textures

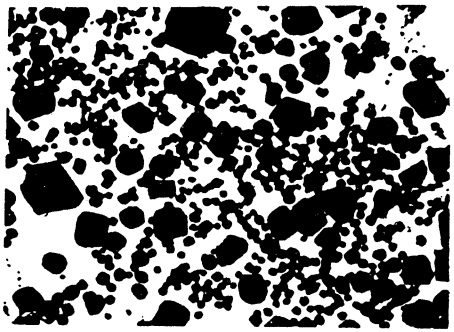

FIGURE 4.—*Chain texture in chromite ore*
Enclosed by labradorite. Forest Hill platinum mine, Bushvelt, S. Africa. × 17. After Sampson.

(Pl. 1, fig. 3), the components of the intergrowth are generally interpreted as contemporaneous. In ores that are not magmatic, graphic textures may be formed by quite different processes, and the two components may or may not be contemporaneous.

If the texture is porphyritic or hypautomorphic granular (including the poikilitic and ophitic), minerals completely or largely enclosed by others are interpreted as older than the enclosing minerals. Usually the enclosed mineral is euhedral or subhedral. Thus, for example, the phenocrysts of a porphyry are older than the minerals of the matrix, and the subhedral crystals of chromite in the chromite ore shown in Figure 2 of Plate 1 are older than the olivine that conforms to their crystal outlines.

TEXTURES RESULTING FROM LATE MAGMATIC CHANGES

A crystallizing magma is a system characterized by changing conditions of equilibrium in the still liquid portions. Such changes in equilibirum may lead to changes in the composition of the minerals deposited and/or to re-solution of minerals already formed. The most familiar example of depositional changes in composition is zoning in plagioclase feldspars; the outer zones usually are progressively richer in soda. Comparable phenomena in the metallic minerals of ores when of clearly magmatic origin seem not to be common. Zoning has, however, been observed in platinum grains in the Urals, in which sharply demarked outer and inner zones differ notably in the amount of iron alloyed with the platinum.

One of the most familiar examples of re-solution is the corrosion of quartz phenocrysts observed in porphyries. Analogous phenomena are occasionally observed in

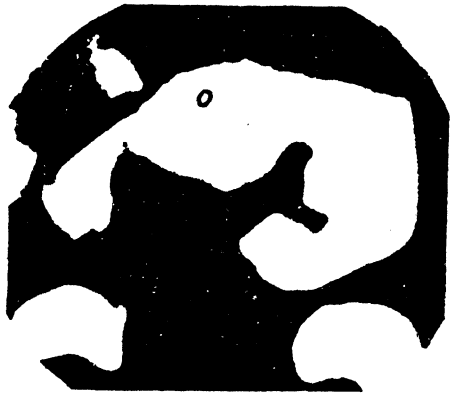

FIGURE 5.—*Corrosion of orthoclase (O) phenocryst by chalcopyrite (black)*
In chalcopyrite syenite from Park County, Montana. × 65. After Lovering.

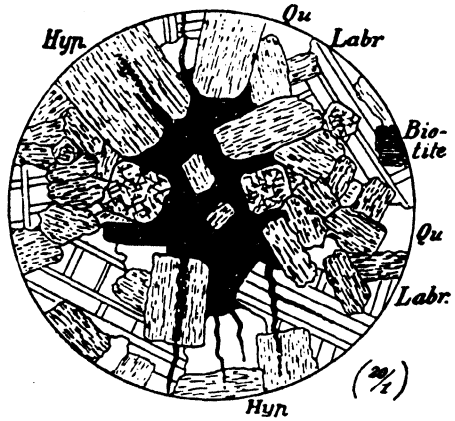

FIGURE 6.—*Late pyrrhotite (black) in magmatic ore*
Penetrating earlier labradorite (Labr), hypersthene (Hyp), and quartz (Qu) along fractures. Quartz norite from Romsaas, Norway. × 16. After Vogt. There is no evidence of corrosion of the silicates by pyrrhotite.

ores as for example the corrosion of feldspar phenocrysts by chalcopyrite (Fig. 5). Figure 2 of Plate 2 shows ore from Feragen, Norway, in which rounded and embayed chromite grains are surrounded by serpentine (after olivine). Donath (Abb. 20) interprets this as corrosion of chromite by still liquid olivine.

Movements within the crystallizing magma may produce minor fracturing of crystals, and into these fractures the still molten components enter and crystallize forming veinlets of late magmatic minerals in the earlier magmatic minerals. Vogt (p. 643) describes this in a norite from Romsaas, Norway (Fig. 6). There is no tex-

tural evidence of either corrosion or replacement of the silicates, and Vogt interprets the relations as a late magmatic injection of pyrrhotite.

Although corrosion or resorption phenomena have been referred to as "replacement by resorption" the writer believes the magmatic conditions under which corrosion takes place differ sufficiently from those characteristic of replacement to justify separate terms. These differences will be more apparent after we have discussed replacement phenomena.

TEXTURES PRODUCED BY CHANGES IN EQUILIBRIUM AFTER THE MAGMATIC STAGE

After complete crystallization of a magma there may still be changes in equilibrium due mainly to the continued cooling of the rock. In magmatic ores, nearly all of which are plutonic rocks, cooling is usually slow, and minor textural modifications may take place in the solid state with the formation of new minerals.

Reaction Rims. One of the best examples of such changes is furnished by reaction rims, usually recognizable only under the microscope (Fig. 7). They are restricted, so far as known, to basic plutonic rocks and are fairly common in such rocks and the associated ores. They are particularly well developed between calcic plagioclase and olivine and in such situations are double, one rim having developed at the expense of plagioclase and the other at the expense of olivine. The boundary between the two rims is smooth and in places straight, and it clearly marks the original boundary of olivine against plagioclase. Between plagioclase and metallic minerals such as magnetite or pyrrhotite there is only a *single* reaction rim which has developed mainly at the expense of the plagioclase. Reaction rims are lacking between chemically similar minerals such as olivine and pyroxene. In contrast to replacements they never occur between grains of the same mineral species or along fractures or cleavages in a single mineral.

Such rims are interpreted as phenomena of the solid stage immediately following complete crystallization from the magma. They record an interchange of substance between certain minerals of notably contrasting composition in response to declining temperatures; the new mineral or minerals being more stable at the lower temperatures than were the original minerals. The new minerals are intermediate in composition between the minerals from which they were formed. There is no evidence of addition of material from outside as in the process of replacement with which such reactions must not be confused.

Unmixing or ex-solution textures. Solid solutions are common and are well exemplified among the common rock-forming minerals by the plagioclase feldspars which form a continuous isomorphous series from albite to anorthite. The student of ore deposits is of course concerned mainly with solid solutions of metallic minerals. However, the concepts of solid solutions and their unmixing seem first to have been developed by the metallurgist and later extended to the study of rock minerals and the metallic minerals of ores. It may be well, therefore, to consider first what the metallurgist means by a solid solution in alloys. If a melted alloy of copper with a relatively small amount of zinc is allowed to cool and solidify, microscopic examination

shows that it is apparently homogeneous and that its properties differ only slightly from those of pure copper. If a series of other melted copper alloys with increasing proportions of zinc (but not exceeding 35 per cent) are prepared each alloy upon solidifying is homogeneous, and their properties depart from those of pure copper in

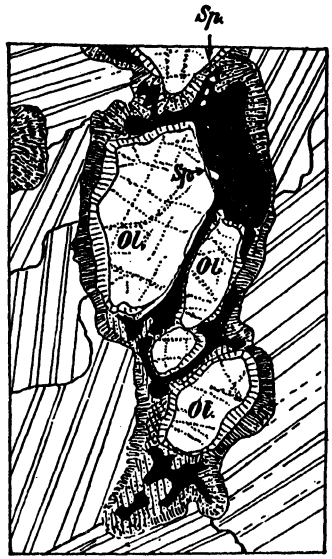

FIGURE 7.—*Reaction rims in olivine gabbro*

From Langö, Norway. Between magnetite (black) and labradorite (ruled) there is a single reaction rim of hornblende. Between olivine (Ol) and labradorite the reaction rim is double; the part bordering olivine is hypersthene, and that bordering labradorite hornblende. Sp is serpentine. × 28. After J. H. L. Vogt.

progressively greater degree as the zinc content increases. Such homogeneous solid alloys whose properties vary progressively with varying composition are termed *solid solution.*

The limits of solid solubility in alloys vary with the nature of the components, with temperature, and probably also with pressure. In general the solubility increases with increase in temperature although a few exceptions are known. If, therefore, a solid solution is near saturation with respect to one component at temperature T and the temperature declines to t, a separation of that component may take place. This is known as ***unmixing*** or ***exsolution.*** In alloys one or both components resulting from unmixing may be (a) metals, (b) compounds, or (c) solid solutions. The textures developed are varied but characteristic.

The interpretation in recent years of certain textures observed in ores as due to unmixing was based on the close resemblance of such ore textures to those that result from unmixing in alloys. The argument from analogy is particularly close in the case of those magmatic ores which solidified from relatively dry melts. Origin

by unmixing of certain of these ore textures has been verified experimentally. Thus certain ore intergrowths of magnetite and ilmenite if heated above 800°C. merge to form a homogeneous solid.

In ore deposits unmixing phenomena have been observed in certain members of four mineralogic classes—the native metals, the oxides of the metals, sulpho-salts of the metals, and sulphides of the metals. Some solid solutions occurring in ores may show no unmixing textures if the natural changes in temperature and pressure have not transgressed the limits of solid solubility. In such cases only analysis will reveal the mixed character of the minerals. This is notably true, for example, with natural solid solutions of gold and silver.

In magmatic ores, unmixing of solid solutions is a phenomenon of the post-magmatic stage when initially high temperatures have declined through a considerable range.

Presumably well-authenticated instances of unmixing in magmatic ore deposits involve the following pairs of minerals.

(1) Magnetite $FeO \cdot Fe_2O_3$	Ilmenite $mFeTiO_3 \cdot nFe_2O_3$
(2) Hematite Fe_2O_3	Ilmenite $mFeTiO_3 \cdot nFe_2O_3$
(3) Chalcopyrite $Cu_2S \cdot Fe_2S_3$	Cubanite $CuS \cdot Fe_2S_3$
(4) Pyrrhotite FeS	Pentlandite FeNiS

The commonest type of unmixing texture is the grating type (Pl. 2, fig. 4) in which the crystal structure of one constituent controls the distribution of the other. The "inclusions" may be needles, blades, plates, or blebs.

The following criteria seem valuable in the recognition of unmixing textures in *ores* (Schwartz, 1931).

(1) The first criterion is mineralogic rather than textural. The products of unmixing are usually minerals possessing certain mineralogic or chemical similarity. In the plagioclase feldspars the components are chemically analogous and are isomorphous. Gold and silver are both isometric, crystallize in a face-centered lattice, and are similar in many of their properties. Arsenic and antimony are both rhombohedral and have similar crystal structures. Among the oxides and sulphides listed above as occurring as unmixing products in magmatic ores magnetite and ilmenite though crystallizing in different systems are chemically similar except that ilmenite contains TiO_2. Hematite and ilmenite are both rhombohedral and differ chemically only in that ilmenite contains titanium. Chalcopyrite and cubanite though belonging to different crystal systems are chemically very similar. Pyrrhotite and pentlandite though crystallizing in different systems differ chemically only in that pentlandite contains nickel. Solid solutions and ex-solution phenomena seem never to have been

demonstrated between members of different mineral groups such as oxides and sulphides or native metals and sulphides. It has been suggested that in some instances gold, in particles too small to be seen under the microscope, is in solid solution with the pyrite, but such a relationship has not been demonstrated (Edwards, 1947, p. 92–93).

(2) The inclusions may be blade-, plate-, rod- or bleb-like. There is no enlargement where the blades or plates join or cross; on the contrary many blades *contract* in width at intersections (Pl. 3, fig. 1). Replacement veinlets on the other hand characteristically enlarge at their intersections. (*Compare* Figure 6 of Plate 9.)

(3) The boundaries of the included mineral are always sharp and smooth. Boundaries formed by replacement may be either smooth or ragged.

(4) The included mineral is commonly absent outside the intergrowths.

(5) The included mineral tends to be rather evenly distributed.

(6) The included mineral usually occurs as disconnected units instead of forming a network as in replacement along mineral cleavages.

(7) Orientation of the included mineral differs for each grain of the enclosing mineral depending upon the crystal orientation of the latter.

(8) The orientation of the included mineral is not related to the grain boundaries of the enclosing mineral (Pl. 2, fig. 6).

(9) In general the absence of evidence of replacement indicates the possibility of unmixing.

Several of these relations must be recognizable in order to make a safe diagnosis.

Exsolution phenomena in ores formed hydrothermally are considered later in this paper.[2]

SELECTED REFERENCES

References of particular importance as bearing on magmatic ore textures are listed in the end bibliography as follows: Per Geijer (1910), Schwartz (1931b), Singewald (1913), Vogt (1921), Warren (1918).

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[2] For a more detailed discussion of solid solutions in ores the student is referred to Chaper IV in Edwards: *Textures of the ore minerals*. This chapter includes lists of (1) oxides, (2) sulpho-salts, and (3) sulphides that have been shown experimentally to form solid solutions or in which unmixing has been inferred from textures.

CHAPTER II. COMMON TEXTURES INDICATIVE OF DEPOSITION IN OPEN SPACES

GENERAL

In Chapters II to IV textures indicative of deposition in rock openings are considered with respect to deposition of (1) crystalline and (2) microcyrstalline and amorphous substances. Although one type of deposition usually dominates, both crystalline and amorphous substances differing in kind may be deposited simultaneously or in quick sequence.

Almost any type of opening in rocks may become the locus of ore deposition. Ores have been found in the following openings.

(1) Pores (in sediments and pyroclastics mainly)
(2) Vesicles
(3) Plant cells
(4) Miarolitic cavities
(5) Solution cavities
(6) Fissures
(7) Faults
(8) Joints
(9) Bedding-plane fractures and "flats and pitches"
(10) Tension fractures in folds
(11) "Ladder vein" fractures
(12) Irregular minor fractures

Only certain of these require discussion from the standpoint of the textures developed in the ores.

As a general rule minerals deposited in open spaces are younger than those forming the walls of such spaces. There may, however, be occasional exceptions. For example the minerals of a fissure filling in granite are younger than the minerals of the granite, but if the granite has been altered hydrothermally they are not necessarily younger than these alteration products.

If the fillings show layering or "crustification" they may be very significant in indicating the sequence of mineral deposition.

CRUSTIFICATION

Crustification is the deposition of mineral matter in successive layers or crusts upon the walls of openings in rocks. There may be many such layers, and usually the layer next the wall was deposited first.[3] In fissures the bands are symmetrical in kind on opposite walls but not necessarily symmetrical in width. The youngest bands may meet in the center of the fissure, or unfilled vugs, usually lenticular, may remain. Minerals that tend to crystallize in elongate forms, notably quartz, are often arranged with their long axes perpendicular to the base of the band, producing

[3]Reopened veins often have a new layer next one or both walls.

what is known as *comb structure*. There is no better illustration of crustification than the familiar one of von Weissenbach first published in 1836 and shown in Figure 3 of Plate 3. This vein is unusual, however, in the large number of bands present.

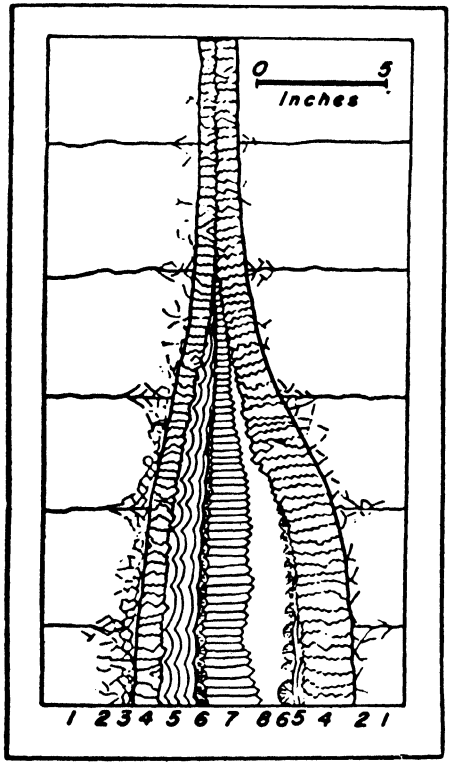

FIGURE 8.—*Crustification in vein on Port au Port Peninsula, Newfoundland*

1. Massive limestone, 2. Limestone traversed by calcite seams, 3. Limestone breccia cemented by calcite, 4. Galena, 5. Colloform sphalerite, 6. Mainly marcasite, 7. Calcite showing comb structure, 8. Vug. After K. deP. Watson.

Such complete symmetry in kind between the bands of opposite walls appears to imply deposition from solutions that *completely* filled the fracture.

The writer has observed typical crustification in veinlets as small as a centimeter in width though the bands were few in number. In veins of microscopic dimensions crustification is not observed, presumably because openings were quickly filled. Figure 8 illustrates this principle and also shows interesting variations from the regular crustification of Figure 3 of Plate 3. Some of the bands (notably 4 and 7) are crystalline, while others (notably 6) were deposited either in a colloidal or a microcrystalline state. Fillings of small fractures, while seldom showing crustification, occasionally show a segmented character.

In some crustified ores many of the bands may be narrow and finely crystalline. This is notably true in the well-known "Schalenblende" zinc ores of Aachen, Germany (Pl. 4, fig. 1). The texture of this specimen is a combination of crustification and cockade or ring structure (top portion). The banded layers are mainly sphalerite in varying shades, but in many specimens there are occasional bands of pyrite. The

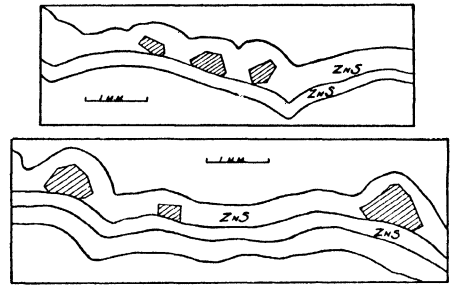

FIGURE 9.—*Crustification in ore from Aachen, Germany*

Camera lucida drawings from parts of the specimen shown in Figure 1 of Plate 4, showing galena crystals deposited on curved surfaces of an older sphalerite band and enveloped by a younger sphalerite band whose surface curvature is controlled by the forms of the galena crystals. After Bastin.

sphalerite bands show little or no regularity either in width or in spacing. Study by the writer of typical specimens in the collections of Cornell University shows that the sphalerite is crystalline showing, under low magnification, cleavage faces when freshly fractured. The cleavage faces show that the sphalerite crystals are oriented nearly perpendicular to the banding and range up to 0.1 mm in length. The occasional pyrite bands are much more coarsely crystalline. The bladelike crystals of pyrite are oriented perpendicular to the banding and range from 2 to 5 mm long; a few extend the full width of the band. Ehrenberg (1931), using hydrogen iodide as an etching reagent to distinguish sphalerite from wurtzite, has shown that much of the sphalerite is pseudomorphic after original crystalline wurtzite (hexagonal ZnS). Lengths of single wurtzite crystals range from 0.5 to 0.8 mm.

Scattered crystals of galena up to 5 mm in diameter cap or are embedded in a few of the sphalerite bands as shown in Figure 1 of Plate 4 and more clearly in Figure 9 which shows camera-lucida sketches from the same specimen. The bottoms of many of these galena crystals conform to the curved outline of the underlying sphalerite band. The galena crystals grew therefore on the curving free upper surface of a sphalerite band. The succeeding band of sphalerite enveloped the galena crystals, and the curvature of its upper surface was controlled in part by the shapes of the galena crystals.

The textures exhibited by the Schalenblende of Aachen somewhat resemble those produced by diffusion into colloids (*compare* Pl. 5, fig. 4), but the regularity in spacing characteristic of diffusion banding is absent, and the curvature of the sphalerite bands depends not on surface-tension phenomena but upon the shape of the underlying surface. In the specimens seen by the writer or described in the literature there seems to be no evidence that even the finest of the minerals—sphalerite—was de-

posited as a colloid. Shrinkage cracks so common in ores deposited as colloids appear to be absent. It is important to distinguish such finely crystalline crustification from the diffusion banding of colloids.

COCKADE OR RING STRUCTURE

This structure is a variant of crustification. Where a fissure becomes partly filled with a rubble of fragments torn or dropped from the walls these may become crusted with successive layers of mineral matter (Pl. 3, figs. 5, 6). Typical occurrences in the Harz Mountains show the same succession of mineral bands on the walls and around many of the fragments. The absence of the oldest bands around certain fragments is explainable if these fell from the walls after mineralization had begun. In some cockade ore many fragments seem not to be in contact with other fragments or with the walls of the cavity. This absence (?) of support is probably apparent rather than real (Talmage, 1929). Also new minerals coating fragments may by their force of crystallization produce a moderate amount of dislodgement (Beck, 1909, p. 248, Emmons, 1940, p. 73).

FILLINGS OF MINOR FRACTURES

The filling of minor fractures is an important mode of mineralization; as replacement may also follow minor fractures it is important to distinguish the two processes.

Openings formed by fracturing may be very irregular because of brecciation or somewhat regular as a result of fissuring. In either case the filling has sharply defined though often sinuous boundaries. Fissure fillings are commonly characterized by matching walls—that is, by walls that would fit together at least moderately well if the filling were removed. Perfect matching is not always found, even in undoubted fracture fillings, since one wall may have moved in a direction inclined to the plane of the section available for study, and fragments torn from the wall in the fracturing may also have moved out of that plane. Typical fracture fillings are shown in Figure 10 and in Figure 2 of Plate 3. The prevalence of matched walls in fissure fillings is in marked contrast to the usual absence of matching in replacement veinlets such as those shown in Figures 3 and 4 of Plate 9.

The fillings of *minor* fractures such as are observed under the microscope seldom if ever show crustification. The filling minerals are usually irregularly associated (Fig. 10), and it may not be possible to recognize age differences among them.

In a few ores many of the small veinlets seen under the microscope are *segmented veinlets*; their component minerals alternating along the trend of the vein. Commonly only two vein minerals are involved, the vein at one place being composed entirely of mineral *A* while the succeeding segments are composed of mineral *B*. Occasionally a third mineral is present. The contacts between the segmented vein minerals are sharp and usually are nearly at right angles to the vein walls. Segmented veinlets are common in the ores of the Cobalt and South Lorraine districts, Ontario, where the segments are commonly silver and calcite (Pl. 4, fig. 3), though in some cases silver and niccolite (Pl. 4, fig. 2). Sharwood (1911) has figured a segmented veinlet of gold and pyrrhotite filling a fracture in arsenopyrite in the Homestake Mine.

Many of the segmented veinlets of calcite and native silver in the ores of Cobalt and neighboring camps in Ontario, Canada, are unquestionably the result of late hypogene replacement of portions of the silver by calcite. The evidence of such an origin is conclusive in the ores of Gowganda and is identical in character with the

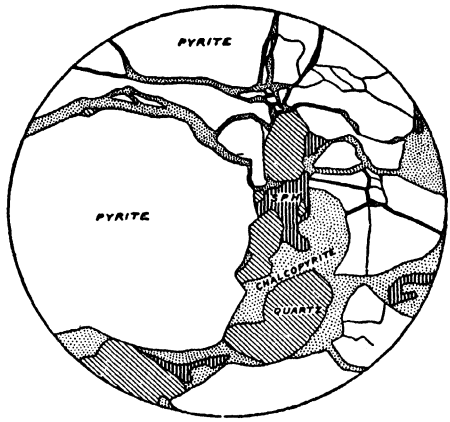

FIGURE 10.—*Fracture filling in ore deposition*

Fractures in pyrite filled with quartz, sphalerite (SPH.) and a little chalcopyrite. The three minerals of the filling seem to be contemporaneous, and there is no evidence that any of them have replaced the pyrite. × 16. Specie Payment Mine, Clear Creek County, Colorado. After Bastin.

evidence given in this paper (Pl. 13) that calcite has replaced dendritic forms of native silver; indeed, some of the segmented veinlets are offshoots from partially replaced silver dendrites.

Whether all segmented veinlets are the result of partial replacement as at Gowganda is uncertain. In the absence of evidences of replacement an alternative explanation would be the filling of fractures by two or more minerals whose grains crystallized from different centers until they finally interfered.

Veinlets with Automorphic Crystalline Walls. An exceptional type of veinlet, well displayed in the cobalt-nickel-silver ores of Cobalt and South Lorraine, Ontario, is illustrated in Figure 3 of Plate 4 and Figures 1 and 2 of Plate 5. The walls of the veinlets tend to match roughly, implying that they are fracture fillings. Examination of the walls of the veinlet under high power reveals, however, that they are made up of automorphic crystals of the mineral that forms the walls of the veinlet. These relationships in ore from South Lorraine, Ontario, are shown at low magnification in Fig. 3 of Plate 4. The segmented veinlets of calcite (black) and silver (white) traverse safflorite. In the larger veinlet the tendency of opposite walls to match is conspicuous. When, however, the walls of such veinlets are more highly magnified (Pl. 5, fig. 1) it is seen that they are composed mainly of automorphic crystals of safflorite. The central part of Figure 1 Plate 5 is shown more highly magnified in Figure 2 of Plate 5. It is evident that no replacement of safflorite by the silver and calcite of the veinlets has taken place.

These unusual veinlets are interpreted as ruptures produced *during* the primary mineralization and while the arsenides were still crystallizing. In these ruptures silver and calcite were deposited though not necessarily simultaneously. Very similar relations are shown in Figure 2 of Plate 4. Many of the veinlets just described are segmented veinlets. Still other features prove that these veinlets of the Cobalt ores are fracture fillings. In places the silver veinlets pass along their trends into breccias of angular fragments of arsenides enclosed by silver. One silver veinlet was observed to give way along its trend to a succession of silver-filled gash fractures oriented diagonally to the trend of the vein—a typical rupture phenomenon. Finally, where a silver-calcite veinlet traverses ore showing the "tubercle" texture, the cap or outer layer of one of the tubercles lies on one side of the veinlet, while the main part of the tubercle lies on the opposite side.

Dendrites (Swartzlow, 1934) may occupy minor fractures or may be developed by replacement. The delicate, branching, crystalline growths deposited along very small fractures are well illustrated by the manganese oxides on joint planes at Leadville, Colorado (Pl. 3, fig. 4). Replacement dendrites are shown in Figures 1 and 3 of Plate 12.

Most fracture fillings in ores are crystalline, but some in crystalline material are colloidal and show colloform textures. Figure 5 of Plate 7 shows small fissures in crystalline smaltite filled with colloidal native arsenic.

PORE FILLINGS

Filling of pores as a mechanism of ore deposition becomes important mainly in rocks in which the percentage of pore space is high enough to give ample space for ore deposition and in which the pores are large enough to make the rock easily permeable. Limonite and hematite often fill the pores of sandstones, but the volume of pore space is almost never sufficient to produce workable deposits. Perhaps the best examples of ores that are in part pore fillings are the copper, vanadium, and uranium deposits of the so-called "Red Beds" type (Pl. 1, fig. 6). In such sandstones one must be sure the material now filling the spaces is not a replacement of an older matrix. Remnants of any older matrix usually remain.

PLANT CELL FILLINGS

The commonest examples of the filling of plant cells by metallic minerals are the pyrite nodules so often associated with coal beds. Examples are also found in association with disseminated copper deposits in sandstones, the so-called "Red Beds" type of copper deposits (Pl. 5, fig. 3). The uncollapsed condition of the cells in many of these occurrences shows that the pyrite filling was deposited before the plant material was buried under any considerable load. Not uncommonly chalcosite has replaced the pyrite. There is good evidence that in some such ores chalcocite has directly filled cell cavities as well as directly replaced wood (Papenfus, 1931, p. 318).

CHAPTER III. PRINCIPLES OF COLLOIDAL DEPOSITION

GENERAL

Deposition of mineral matter in a very finely divided state may produce ore textures very different from those that characterize coarse cyrstalline precipitates from "true" or electrolytic solutions such as have thus far been considered. Brief consideration of some general characteristics of fine dispersions is necessary before proceeding to discuss their role in ore deposition.

When a cube is divided into eight cubes of half the original diameter the surface exposed is thereby doubled. If the same cube is divided into cubes of one-fourth the original diameter the surface is quadrupled. Thus without change in mass the amount of exposed surface is notably increased, and all phenomena dependent upon the total amount of exposed surface of the solid are thereby greatly amplified. If the process of comminution is continued, ultimately a condition is reached in which the particles will be so small that they will not settle readily in any fluid medium in which they are suspended, and may remain suspended for months. Such systems are called *colloidal dispersions* or *sols;* the particles are the *dispersed phase*, and the fluid is the *dispersing medium*. This general principle applies regardless of whether the substance subdivided be solid or liquid and regardless of whether the surrounding medium be gas or liquid. Thus the importance of large surface areas of exposure may be manifest in such diverse phenomena as tiny drops of water suspended in air as fog, in small solid particles suspended in air as smoke, in fats suspended in water as cream, and in particles of solid gold suspended in water.

Characteristics of colloidal dispersions are: (1) buoyancy, (2) brownian movement, (3) adsorption, (4) electrical charge, and (5) chemical reactivity. These will require comment in later paragraphs.

The economic geologist is concerned almost exclusively with dispersions in liquid water which, however, is not pure water but carries electrolytes in solution and may also carry more than one substance in suspension. Further, he is concerned mainly with dispersions of solids in water but subordinatley with dispersions of liquids insoluble in water. In further discussions these will be the types of dispersions principally emphasized. Colloidal dispersions or sols have already been defined, and coagulates or flocculates from such dispersions will be referred to as *colloidal flocculates*, regardless of whether the flocculated particles be solid, liquid, or mixtures. The term *gels* is restricted to colloidal flocculates or coagulates from emulsoids which, like gelatine or colloidal silica, may show somewhat elastic, jellylike properties. Flocculates from suspensoids are not jellylike. The textures characteristic of colloidal flocculates may be termed colloidal texture or, with certain limitations, *colloform* textures following the usage proposed by A. F. Rogers (1917, p. 518); Colloform refers only to the shape or form of the deposit regardless of whether the material is now collidal or exhibits only a relic colloidal or to use Wherry's (1914, p. 112) term, a metacolloidal texture.

Most particles in colloidal dispersions are too small to be seen with the compound microscope, but with proper illumination, in the ultramicroscope, reflections of light from the particles can be seen, and it can be determined that they are in rapid motion.

BUOYANCY

As the particles in most colloidal dispersions in water are much heavier than water we may inquire why they do not rapidly settle. If a cube of unit diameter is divided into cubes of a fourth that diameter each of these has one sixty fourth the mass of the original cube but has one sixteenth as much surface. Thus as materials become progressively finer-grained the ratio of surface to mass increases, and buoyancy tends more and more to delay settling. Thus it can be calculated (Burton, 1938, p. 4) that under quiescent conditions a sphere of silver 1 cm in radius in water will settle 1 cm in five-millionths of a second. A sphere 0.00001 cm in radius will require about half a day to settle the same distance, and a sphere 0.000001 cm in radius will require about 58 days. Thus if no factors other than buoyancy were involved colloidal particles would tend to settle slowly. There are, however, other factors that add to the effects of buoyancy.

"BROWNIAN MOVEMENT"

The ultramicroscope shows that the motion of particles in colloidal suspensions lacks regularity, and that it changes abruptly in direction. This so-called Brownian movement is believed to be due to unequal bombardment of the small dispersed particles by the molecules of the medium. At one instant more fast-moving molecules may strike from one quarter, the next instant from a different quarter, causing the particles to move in successively different directions. In very fine particles the magnitude of the random motion may be many times the motion due to gravitation and consequently may be much more important than buoyancy in keeping the particles in suspension.

According to Burton (1938, p. 5) the feature common to all colloidal dispersions is the small size of the particles; they are small enough not to settle out under gravity (about 10^{-5} cm in diameter) and large enough to have a very slow rate of diffusion (about 10^{-7} cm in diameter).

ADSORPTION

In the interior of a liquid or a solid, molecules are attracted equally in various directions, and these attractions tend to neutralize each other. In a surface, however, molecules have their attracting forces neutralized on one side only, and surfaces therefore exert strong attractions on molecules as well as on atoms and ions in their close vicinity. This phenomenon, known as adsorption, is particularly well displayed in colloidal dispersions because of the large amounts of surface involved. Adsorption is often highly selective.

ELECTRICAL CHARGE

Not only do fine particles dispersed through liquids adsorb molecules of the liquid, but they may adsorb ions as well, and as these ions carry electrical charges the solid particles also become electrically charged. Some substances adsorb positive ions more readily than negative ions, and vice versa. In water solutions the hydrogen and hydroxide ions are principally involved. In dispersions of a single substance in a liquid

the particles become similarly charged and tend to repel each other. This is an important factor in keeping colloidal particles in suspension because flocculation takes place readily when the charges are neutralized.

An important consequence of the electric charges on the particles of colloidal dispersions is that when such dispersions are brought in contact with other colloidal dispersions whose particles are oppositely charged neutralization of both charges may occur and flocculation ensue. Similarly when a colloidal dispersion is brought in contact with a true or electrolytic solution carrying an abundance of *oppositely* charged ions neutralization with resultant flocculation may also occur. The peculiar effectiveness of aluminum sulphate in water clarification is due to the fact that it yields both negative and positive ions and is therefore effective in flocculating both positively and negatively charged colloidal particles.

CHEMICAL REACTIVITY

The large amount of surface that characterizes colloidal dispersions of solids in either gases or liquids is in general favorable to chemical interaction between such solids and the enclosing fluids or with substances dissolved in such fluids. The principle is particularly well illustrated in oxidation phenomena as for example the explosibility of magnesium dust in air as contrasted with the stability of a bar of magnesium.

Thus colloidal dispersions have many unique characteristics, and these should be reflected in characteristic textures.

In the naturally occurring colloidal dispersions in water the water contains dissolved mineral matter. Deposition from such dispersions takes place by flocculation of the dispersed particles and by crystallization of the dissolved material. There seems to be no *a priori* reason why the two processes should not proceed concurrently, and indeed the ore textures indicate that such has often been the case. Furthermore, if chemical reactions in electrolytic solutions are the starting point in the formation of most natural colloidal dispersions, associations of this sort should be common.

FORMATION OF COLLOIDAL SUSPENSIONS

In the laboratory the two general methods of getting substances into the colloidal state are: (1) Larger particles are disintegrated until colloidal proportions are attained. This can be done in an electric arc or by fine grinding in the presence of some means of preventing flocculation of the particles. The formation of fine clays which have many of the properties of colloids in perhaps the most familiar geologic analogue of this method, but there seem to be few examples in the processes of ore deposition. (2) Substances in the molecular condition, usually in "true" or electrolytic solution, are built up to colloidal particles; precautions must be taken to prevent the particles from growing too large. This method involves a reduction in surface with consequent liberation of energy. It might, therefore, be expected to occur spontaneously and rather commonly in nature; and actually most natural colloids have formed in this way. Any type of chemical reaction whether of oxidation, reduction, double decomposition, or hydrolysis may initiate the building up of colloidal dispersions.

SUSPENSOIDS AND EMULSOIDS

Thus far features common to all colloidal dispersions have been stressed. Some significant contrasts, partly qualitative exist, however, between (1) dispersions of solids in water, and (2) dispersions of liquids in water or dispersions of certain substances which though solid have many of the properties of liquids, as, for example, agar agar and gelatine. Recognition of these classes is of great importance to the chemist and the economic geologist.

Dispersions of solids in a liquid are commonly termed *suspensoids;* in nature the liquid is usually water. Such dispersions are nonviscous, nongelatinizing, tend to be non-stable and easily flocculated, and once flocculated are not readily redispersed. To this class belong nearly all colloidal dispersions of the metals and their compounds.

Dispersions of liquids in other liquids in which they are insoluble and dispersions of substances such as gelatine in liquids are usually termed *emulsoids.* They are often viscous and tend to gelatinize. They are likely to be more stable than suspensoids for they are not so readily flocculated by electrolytes. Gelatine and like substances tend to disperse readily and spontaneously in water. Redispersion after flocculation can be easily accomplished. Emulsoids of oil in salt water are of much importance to the petroleum geologist. Colloidal dispersions of SiO_2 in water can be prepared in the laboratory from soluble silicates, and under certain conditions such silica dispersions coagulate to form stiff gels. In ore deposits silica *now* gelatinous is very rare, as we might anticipate, since silica gels are unstable and most ore deposits were formed long ago. Nevertheless some occurrences are known as in the upper portions of the Great Australia mine in Queensland, Australia (Levings, 1912, p. 478). That much of what is now quartz in certain types of ore deposits was deposited originally from colloidal dispersion may be inferred, however, from relic colloidal textures still visible in the quartz (Pl. 5, fig. 5).

STABLIZING AGENTS

Despite the dispersive effects of Brownian movement and of like electrical charges on the particles, colloidal dispersions are unstable systems. The dispersed particles tend to flocculate easily, and the flocculated material tends to settle and separate from the fluid in which it was suspended. After flocculation redispersion is very difficult and under natural conditions usually impossible. However, certain substances facilitate the formation of colloidal dispersions and tend to perpetuate or "stabilize" or "protect" them once they are formed.

The presence of certain emulsoid colloids, such as gelatine, facilitates the formation of suspensoids, renders them more stable, and often make it possible to obtain more concentrated suspensoids and to redisperse a flocculate that otherwise could not be dispersed. Seemingly, the emulsoid is adsorbed on the surface of the solid particles of the suspensoid and this layer inhibits flocculation both by preventing the electrical discharge of the suspensoid particles and by offering a buffer to their coalescence. Thus the suspensoid acquires some of the properties of an emulsoid.

In the laboratory certain common ore minerals are effective precipitants of gold and silver from solutions of gold chloride and silver sulphate (Palmer and Bastin,

1913). Furthermore, with addition of a gelatine emulsoid to the solutions, the gold and silver can be obtained in beautiful colloidal suspensions instead of as crystalline precipitates. Whether analogous processes operate in the natural transport and precipitation of gold and silver is not known. The emulsoid most likely to function in nature as did the gelatine in these experiments (Bastin, 1915) is a silica emulsoid. In further experiments silica was substituted for gelatine, and colloidal solutions of gold or silver were obtained with many of the ore minerals.

Ordinarilly, flocculated suspensoids cannot be redispersed, but in the presence of "peptizers" some flocculates can be redispersed. Among the substances that might function as a peptizer in nature is hydrogen sulphide (Clark and Menaul, 1916) which is known in the laboratory to facilitate the formation of colloidal suspensions of certain sulphides and sulpho-salts in alkaline solutions.

CHAPTER IV. COLLOIDAL ORE TEXTURES

GENERAL

While it is now generally accepted that certain ores were deposited as colloids the proof of such an origin in many cases offers inherent difficulties. Examples of present-day deposition of such ores are not numerous and have not been adequately studied. Most of the ores thought to be of colloidal origin were formed long ago and have crystallized and developed new textures. In the siliceous ore from the Talisman Mine, New Zealand, the microscope reveals a granular texture in polarized light (Pl. 5, fig. 6), but the same field in ordinary transmitted light (Pl. 5, fig. 5) shows an older and very different and much finer texture characterized by spheroidal forms outlined by bands of dark, opaque, accessory minerals. The obvious extreme fineness of this earlier texture and the presence of outlines that are almost perfect circles or arcs thereof suggest that the silica and probably the accessory minerals were originally deposited as very finely divided material of such low viscosity that surface tension could operate to produce spheroidal surfaces. Such deposits are commonly interpreted as colloidal. In general the finer the original grain, the more nearly circular the outlines, and the smaller the scale on which the phenomena are displayed the stronger is the probability of colloidal origin. Typical deposits (Pl. 5, fig. 5) show occasional vugs bordered by spheroidal surfaces and other evidences of deposition in open spaces.

To clarify the surface-tension explanation of such textures it is appropriate to recall some general characteristics of surface-tension phenomena. The spheroidal outlines assumed by one liqid when suspended in another of equal density, the spheroidal forms of small rain drops, and the incompletely spherical form of the bubbles in various kinds of foams are all "surface-tension" phenomena. The molecules in the center of a body of liquid are attracted equally in all directions by the surrounding molecules, but towards any free surface of the liquid the outward attraction in a direction normal to that surface decreases becoming zero when the surface is reached. Thus molecules at any point on the surface are being pulled parallel to the surface and toward the interior but *not* outward. The effect is somewhat as if the mass of liquid were enclosed in a tightly stretched flexible membrane. The molecules of liquid tend to draw together as much as possible until the whole mass has the smallest possible surface compatible with its volume. Unless distorted by forces outside the liquid, this surface is spherical.

Not only may wholly liquid substances assume rounded outlines because of surface tension, but liquids carrying much finely divided solid or semisolid matter in suspension (colloidal suspensions) may also assume spheroidal outlines under the influence of surface tension. Good examples of rotund surfaces are found among nonmineral substances in such familiar colloidal coagulates as glue and cottage cheese; the latter contains usually about 70 per cent of water.

Not only have spheroidal deposits of silica like that shown in figure 5 of Plate 5 been interpreted as colloidal, but many deposits of oxides, sulphides, and a few elements that exhibit globular, botryoidal, reniform, or mammillary forms even on a relatively large scale have also been so interpreted. When the materials in their pres-

ent state are fine-grained, when the spheroidal forms are developed on a small scale and approach sphericity, and when the material is notably hydrous, this interpretation may be valid; for others, alternative explanations must be considered before final diagnosis.

Several textural features of colloidal flocculates depend directly on their high water content which in turn is due to the large amount of surface available for water adsorption. The water content is a maximum in gels formed by the coagulation of *emulsoids*. A fairly stiff agar agar gel may be 99 per cent water, and gelatine gels that are 85 to 90 per cent water may be stiff enough so that their modulus of elasticity can be measured. In the laboratory silica gels can be prepared stiff enough to be cut yet contain over 90 per cent water. Natural silica gels occur at some hot springs, and a few have been recorded in mines, but no determinations of their water content seem to have been made. Natural opal is probably the solidified product of such gels, but during solidification the water content becomes greatly reduced. Ninety-four analyses of opal quoted by Hintze (1915, p. 1535–1538) show water contents mainly between 5 and 16 per cent.

Little information is available concerning the water content of artificial flocculates from *suspensoids*—the class that most concerns the mining geologist. Noncrystallized varieties of hydrous iron oxide listed by Hintze (1915, p. 2073) show for 77 analyses water contents ranging from 5 to 26 per cent; these were presumably more hydrous when first deposited.

In the earlier studies of colloidal ore deposition emphasis was laid upon the rounded, more or less spherical forms and in 1917 Rogers proposed the term "colloform" for such forms when developed *in open spaces* (p. 518).

> "From the standpoint of physical chemistry amorphous solids are liquids. Now, the shape of a liquid unaffected by gravity or other external influence is spherical, and so we often find the hydrogel minerals in spherical, botryoidal, reniform, stalactitic, and mammillary forms. These forms intergrade, so that one is often at a loss which term to use. I therefore propose the term *colloform* for the rounded, more or less spherical, forms assumed by colloidal and metacolloidal substances in open spaces. Some crystalline, not merely microcrystalline, minerals such as smithsonite, also occur in colloform crusts, and it should be emphasized that this term refers only to the shape or form, and not to the condition of the material"

Lindgren (1925) pointed out that colloidal minerals can also be deposited by replacement and cited marcasite "with colloform structure" that had replaced gangue minerals as well as pyrite and sphalerite (Pl. 5, fig. 4). He also cited as perhaps the most familiar example of colloidal replacement the development of rounded chert nodules in limestones and dolomites—especially those nodules in which oölitic textures of the limestone were preserved in the chert (Pl. 8, fig. 6).

Other textural features besides "rounded, more or less spherical forms" suggest though they do not always demonstrate colloidal deposition. By definition Rogers' term "colloform" was restricted to spheroidal forms developed in open spaces, yet a comprehensive term is needed to include all textures that point to the original deposition of the material in the colloidal state. The writer would be willing to see Rogers' term "colloform" broadened to serve this purpose, but in the present book "colloidal textures" is used as the comprehensive term.

The following textural features at least suggest colloidal deposition: (1) rotund or

spheroidal forms (in part), (2) syneresis or shrinkage cracks (in part), (3) pellet texture (in part), (4) framboidal texture, (5) interference surfaces (in part).

ROTUND OR SPHEROIDAL FORMS (IN PART)

By no means all of the rounded, more or less spheroidal forms found in ores indicate colloidal deposition. In the stalactites in certain zinc mines cross sections show rounded *outer* surfaces, and the successive bands of sphalerite and pyrite or marcasite of which they are made up also show rounded outlines. In many such stalactites the minerals are now crystalline, and there is no evidence that they were colloidal when deposited. In certain veins rounded outlines are developed by the successive deposition of bands of different minerals around rock fragments as illustrated in the cockade or ring ore shown in Figures 5 and 6 of Plate 3. All the minerals of this ore seem to have been deposited in the crystalline rather than the amorphous state. In still other ores rounded outlines may be developed by the diffusion of the mineralizing solution outward from certain centers often combined with rhythmic precipitation (Pl. 14, fig. 4; Pl. 15, fig. 1). In some such banded ores colloidal deposition is probable, but in others the ore minerals were crystalline as deposited.

Rotund outlines combined with quite delicate banding (Pl. 4, fig. 1) are sometimes developed by layered deposition or crustification on the walls of openings or around angular fragments of wall rock or older ore minerals—a delicate cockade texture. In Fig. 1 of Plate 4 galena forms crystals up to 5 mm across, and the associated sphalerite though much finer is far above colloidal dimensions.

There are, however, many other ores, mostly finer-grained, in which spheroidal outlines are developed in such delicacy, smoothness of outline, and close approach—in cross section—to a circle or an arc thereof as to strongly suggest that *surface tension* caused the spheroidal forms. Such textures are well displayed on a microscopic scale in the gold-bearing quartz from the Talisman mine, New Zealand. Figure 5 of Plate 5 indicates that nearly complete reorganization of the silica to a microcrystalline aggregate has not obliterated the original spherulitic forms.

Original colloidal textures are most easily demonstrated in transparent ores, that can be studied under the microscope in transmitted light. In opaque ores that can be studied only in polished specimens by reflected light evidence of colloidal deposition is likely to be presumptive rather than conclusive. In the covellite-chalcocite ore from Kennecott, Alaska (Pl. 6, fig. 1), both minerals are now crystalline. However, spheroidal outlines suggest surface-tension phenomena, and the chalcocite seems to have been guided in its replacement of the covellite by fractures that were (1) radial, (2) concentric, or (3) netlike and hence strongly suggestive of the shrinkage cracks so characteristic of colloidal materials. Lasky believes the covellite was originally colloidal and became fractured with shrinkage. Subsequently it crystallized to its present radiating pattern and was partially replaced by chalcocite.

Colloids deposited by replacement may exhibit rotund forms somewhat similar to those just described but apparently due to diffusion and rhythmic precipitation and not to surface tension. Figure 4 of Plate 5 shows partly oxidized ore from the Bigbug District, Arizona, in which fine-grained, curved concentric bands of marca-

site are enclosed in a relatively coarse granular aggregate of sphalerite, pyrite, and gangue minerals. In places the banding of the marcasite transects the grain boundaries of the matrix. According to Lindgren (1926, p. 140–141) this represents a replacement of sphalerite and gangue by colloidal iron sulphide which later crystallized to marcasite.

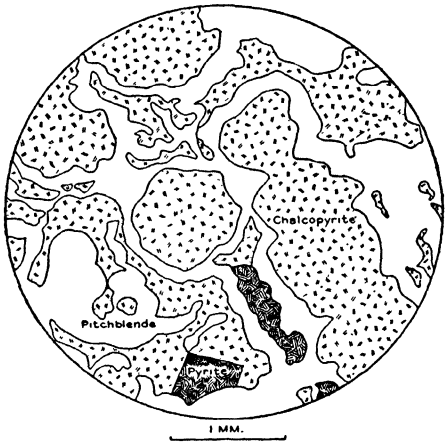

FIGURE 11.—*Colloform pitchblende*

Association of crystalline pyrite and chalcopyrite with pitchblende that was deposited as a colloid and shows spheroidal outlines. The three minerals are believed to be of nearly the same age. Wood mine, Central City, Colorado. × 17. After Bastin.

Lindgren described excellent examples of replacement of limestone by highly siliceous ores from the Gemini mine in the Tintic District, Utah (Pl. 15, fig. 2). These are discussed at more length in the Chapter on Replacement. He believed (1915, p. 231) limestone was first replaced by colloidal silica; metal-bearing solutions diffusing into this silica gel then deposited galena and sphalerite rhythmically to form the darker bands. In the lighter bands there are chalcedony spherules. The banding is due to diffusion and rhythmic precipitation.

As might be expected, colloidal substances are weak materials and easily deformed by external stresses. Figure 3 of Plate 6 shows an ore from Silverton, Colorado, which is a breccia of pyrite fragments in a matrix of quartz with accessory sphalerite and galena. The pyrite fragments show the rotund outlines and the radial and concentric shrinkage cracks characteristic of colloids.

All or only some of the minerals of an ore may have been deposited as colloids. In the rather unusual ore shown in Figure 5 of Plate 7 older crystalline smaltite has been brecciated and in the fractures was deposited native arsenic having spheroidal outlines suggestive of colloidal origin. Even among essentially contemporaneous ore minerals some may be deposited as crystalloids and others as colloids. In the ore shown in Figure 11 the three-component minerals are nearly contemporaneous, yet the chalcopyrite and pyrite were crystalline whereas the pitchblende was deposited as

a colloid as shown in the polished specimen both by spheroidal outlines and shrinkage cracks. The explanation of such associations is inherent in the nature of the minerals. Just as some minerals such as barite and pyrite have a strong tendency to crystallize so the oxide of uranium is, in ore deposites, almost invariably deposited as a colloid.

SYNERESIS OR SHRINKAGE CRACKS (IN PART)

Fracturing, usually on a small scale and often microscopic, is common in ores whose other characters suggest colloidal deposition. Such fracturing is often either parallel to or approximately perpendicular to spheroidal outlines. When, colloidal and crystalline minerals are associated in the same ore, this type of fracturing is restricted to the colloidal minerals (Pl. 6, fig. 5). Such relations indicate that the fractures are not due to external stresses but are of internal origin. Cracks similar to those under discussion can be developed in the shrinkage of artificial gels and have been termed *Syneresis* cracks. In colloidal ores as in the artificial gels the cracks seem to have been formed by loss of water.

In some ores the shrinkage cracks form a fine irregular pattern somewhat resembling that of crackled porcelain (Pl. 6, fig. 2). This ore is composed of very fine-grained chalcocite with a steely luster and a conchoidal fracture. Etching the polished surface with nitric acid brings out the fracture pattern.

Shrinkage cracks developed in colloidal pitchblende from Joachimsthal, Bohemia, are shown in Figure 4 of Plate 6. Some of these fractures trend either parallel to or perpendicular to the curved boundaries of the pitchblende but others show diverse trends. The cracks in the pitchblende do not penetrate the surrounding mineral. The brecciated colloidal iron sulphide of Figure 3 of Plate 6 shows both radial shrinkage cracks and curving cracks that parallel spheroidal surfaces in fact most of the fragments are bounded by such fractures.

Shrinkage cracks may traverse an association of two or three colloidal minerals. In the ore from the Cornwall mine, Missouri (Pl. 6, fig. 5), shrinkage cracks traverse globular intergrowths of chalcocite and covellite but do not extend into the surrounding chalcopyrite. Rust (1935) interprets all these minerals as primary (hypogene).

A fairly common type of shrinkage crack in colloidal ores is the bifurcating form shown in Figure 6 of Plate 6 also from the Cornwall mine, Missouri. The crack traverses an aggregate of tiny chalcopyrite spherules (often with pyrite cores) in a bornite matrix. The sulphides have replaced a nodular area of dolomite in sandstone. Rust believes that the sulphides were deposited in colloidal form and that the cracks developed by shrinkage on drying.

In cross sections of banded presumably colloidal ores, tapering cracks of the sort shown in Figure 6 of Plate 7 are sometimes present. This figure shows banded chalcopyrite ore deposited on the wall of a fracture in limestone in the Cornwall mine, Missouri. These cracks are undoubtedly due to shrinkage, and those in the lowest (and oldest) band formed before the overlying band was deposited. The tapering forms show that shrinkage was greatest at the former surfaces of the successive bands.

Microscopic shrinkage or syneresis cracks in very fine-grained materials suggest

deposition of the material in a colloidal state. Such cracks are not, however, confined to colloids, for familiar mud cracks are of this class.

PELLET TEXTURE (IN PART)

In certain ores, most if not all of which have been developed by replacement, some of the metallic minerals when viewed under the microscope are tiny spheroids which may be homogeneous or may show concentric or radial structures. For such forms the term pellet texture is proposed. The pellets resemble oölites in form but are on a much smaller—usually a microscopic— scale and have been formed by quite different processes.

Not all pellets are of colloidal origin. They are a common feature for example of the ores of Mount Isa, in Queensland (Grondijs and Schouten, 1937) which are replacements of very fine sediments. There the spheroidal pellets usually have as their core a crystal of pyrite, and there are gradations from spheres to automorphic crystals of pyrite (Pl. 11, fig. 1).

In other ores the pellets appear to have been deposited in a colloidal state. In ore from the Cornwall mine, Missouri (Pl. 6, fig. 5), a colloidal origin is favored by the shrinkage cracks so well developed within the pellets. Kidd and Haycock (1935) have described in the ores of the Eldorado mine, Great Bear Lake, very small pellets of pitchblende in quartz. These are clearly colloidal. They are often found between much larger spherules of pitchblende.

FRAMBOIDAL TEXTURE

The first description of this texture in ores from the United States appears to be of ore from the Cornwall mine, Missouri (Rust, 1935). They are shown in Figures 1 and 3 of Plate 7. To this unusual texture Rust applied the term "Framboidal" (after Framboise, the French for Raspberry) because of its resemblance to the aggregate fruits of the raspberry. The texture was recognized earlier by Schneiderhöhn (1923) in the copper shales of Mansfield, Germany, but because of the extremely fine grain of these ores was not very clearly figured nor was a distinctive name given to it. A magnification of $\times 800$ was necessary to make these Mansfield structures visible. They consist of globular (rarely rodlike) masses 4 to 15 μ in diameter mainly of chalcopyrite, though bornite or chalcocite may substitute for chalcopyrite. Each globular mass is made up of many tiny chalcopyrite grains, and a few of the grains exhibit triangular outlines. Their maximum diameter is about 0.2 μ.

Schneiderhöhn believed these ores were deposited in colloidal form as a gel of mixed sulphides which later became very finely crystalline. He suggested ore deposition on the sea bottom and interpreted the small chalcopyrite spherules as representing colonies of sulphur-depositing bacteria and the tiny grains of which they were composed as "fossil bacteria." Not only the form of the spherules but the order of size he regarded as supporting the bacterial hypothesis.

The "framboidal" spherules described by Rust from the Cornwall mine, range in diameter from 18 to 48 μ or about 3 to 12 times as large as those in the Mansfield deposits. Most of them are pyrite rather than chalcopyrite.

Schouten (1946) has published excellent illustrations of framboidal textures in the Copper shales of Mansfield and has figured similar textures from the ores of Meggen, Germany, from Redruth in Cornwall, England, and from Rio Tinto, Spain. A typical occurrence at Meggen is shown in Figure 2 of Plate 7. Schouten discusses Schneiderhöhn's theory of a bacterial origin of this texture and rejects it in favor of an inorganic origin.

The recognition of framboidal textures in undoubted hypogene ore deposits such as those of the Cornwall mine, Missouri, Rio Tinto, Spain, and Redruth in Cornwall, England, rules out a bacterial origin. According to Rust (1935) the framboidal textures as displayed in the Missouri occurrences

> "are probably metacolloidal in origin, having been formed by the crystallization of a globule of pyrite gel, crystallization starting simultaneously at many points. In an alternative hypothesis, the texture might be considered to have been formed by the bunching of tiny pyrite grains as they floated in a chalcopyrite gel. On the basis of uniformity of grain size, gradational types between the grain clusters and solid spheroids, and the fact that in non-circular, irregular pyrite masses, the 'framboidal' texture may be produced artificially by etching with nitric acid, the writer prefers the former interpretation." (p. 408).

Some of the pellets illustrated in Figure 5 of Plate 6 show incipient framboidal texture, and the shrinkage cracks indicate a probable colloidal origin.

The distribution of the small pyrite crystals shown in Figure 2 of Plate 7 is not suggestive of a floating together of the pyrite grains; it seems more likely that the framboidal masses are composite concretions, the tiny pyrite crystals having developed at their present locations. Notable variations in the size of the pyrite crystals in closely adjacent framboidal clusters also favor this interpretation. (Pl. 7, fig. 3).

Known framboidal textures are confined to fine-grained ores many if not all of which are of hypogene origin. At the Cornwall mine they are in ores that, on the basis of other evidences, were almost certainly deposited in colloidal form. Whether all framboidal occurrences were colloidal when deposited is uncertain, and such uncertainty obviously limits the value of this texture as a criterion of colloidal origin.

In the Cornwall ores the framboidal clusters are of about the same order of size as the pellets already described. In some cases the cluster has a matrix of primary chalcopyrite; in others the matrix is secondary. Within most clusters there is fair uniformity in the size of the units, but in some clusters there is large size diversity. The average size of the units may differ considerably even in adjacent clusters. Pellets and framboidal aggregates are often associated in the Cornwall ores, and there are suggestions of transitions between types.

INTERFERENCE SURFACES

In some minerals, judged on other grounds to have been deposited as colloids, as for example the pitchblende of Figure 4 of Plate 7, the contacts between different portions of the colloidal mineral are straight and in marked contrast to its curving outlines against other minerals. These boundaries resemble the nearly straight contacts of biscuits in a pan as contrasted with their rotund free surfaces. These phenomena in the pitchblende suggest interference in a yielding medium growing from many centers. The interference surfaces, if such they be, belong to the period of growth of the colloid and are slightly older than the shrinkage cracks which formed

after growth had ceased. In Figure 4 of Plate 7 many shrinkage cracks are restricted to the individual pitchblende units and do not cross the interference surfaces that separate these units. Shrinkage cracks may follow along the interference surfaces as they may also follow spheroidal surfaces. Interference surfaces have usually not been distinguished from shrinkage phenomena but seem to be quite distinct, as suggested by a comparison of Figure 4 of Plate 7 with Figure 4 of Plate 6, both showing pitchblende.

Interference surfaces are not confined to colloidal deposits but have been observed between the finely crystalline spherulites of certain lavas where neighboring spherulites interfere in their growth (Harker, 1909, p. 276–280). Such spherulites may show radial and/or concentric growth patterns.

SELECTED REFERENCES

References of particular importance on colloids and colloidal textures are listed in the bibliography at the end of this paper as follows: Boydell (1924–1925), Burton (1938), Lindgren (1924), Rogers (1917), Rust (1935), Zsigmondy (1917).

CHAPTER V. REPLACEMENT TEXTURES

GENERAL

Replacement is the dissolving of one mineral or group of minerals and the *immediate* deposition of another mineral or group of minerals in the place thus vacated. There is no intervening formation of any open spaces of a magnitude visible with the compound microscope. Theoretically minute spaces must exist briefly. This phenomenon is one of the most widespread and varied with which the geologist deals. Although its *practical* importance seems to be limited mainly to the geologic field, probably because of the slowness with which replacement proceeds, it holds much of theoretical interest to the physicist and chemist whose co-operation is needed for a full understanding of the process. Replacement is particularly well displayed in the modification of limestones and dolomites through the silicification of their contained fossils, in the development in them of chert nodules, and in the formation within them of large bodies of ore composed mainly of sulphides of the metals. In the process of downward enrichment of ores, which is of major economic importance, replacement is the dominant mechanism. The substance replaced is referred to as the *host*, and the replacing substance as the *guest*.[4]

The clear relationship of most replacement phenomena to fractures, mineral cleavage, contacts between mineral grains, etc., shows that solutions were the dissolving and depositing agency. These solutions were sufficiently mobile to be guided by openings of microscopic dimensions so water was probably their principal component. Many replacements have taken place at ordinary temperatures, particularly those accomplished by waters of surface origin. Other replacements have been accomplished by thermal solutions at temperatures ranging up to 500°C. and possibly higher. The solutions responsible for the higher-temperature replacements were presumably gaseous.

Essentially every mineral or rock species has under some circumstances undergone replacement. Seemingly the only general rule is that rocks such as the carbonate rocks—limestones and dolomites—that are readily soluble under a great variety of natural conditions have been subject to replacement most commonly and on the largest scale. Minerals like quartz and rocks like quartzites, which under most natural conditions are difficulty soluble, may nevertheless yield to replacement under special circumstances.

In most cases the replacing (or guest) mineral is now crystalline and was probably crystalline when it was deposited. In other cases the guest mineral, though now crystalline, is believed to have been colloidal when deposited. Evidences for this belief are mainly spheroidal outlines assumed by the guest mineral and the presence of concentric and radiating shrinkage cracks filled with later minerals. These features suggest crystallization of a gel and have already been described.

In most situations where replacement has been active the replacing solutions have circulated along channels of submicroscopic size. These usually connect, however,

[4] A synonymous term is metasome.

with larger openings nearer the source of the solutions. In the submicroscopic openings the molecular attractions between the minerals and the solution must be important.

In almost all cases there has been no detectable volume change in replacement. This of course would be expected if in a rigid mineral or rock there were no visible open spaces to start with and space for the guest minerals became available only by dissolving of the host. In many instances no portion of the replaced substance enters into the replacing substance, as in the replacements of limestone by quartz, galena, etc. In other instances chemical interchange is involved as when limestone is replaced by calcium silicates such as wollastonite or when, in the enrichment of a copper ore, chalcopyrite is replaced by chalcocite.

In replacement we have a mechanism whereby large bodies of ore may be deposited without the previous existence of any considerable cavities. The material dissolved during replacement is of course eventually deposited but often far from its source. Other features of replacement can best be considered with the aid of specific illustrations.

PSEUDOMORPHIC REPLACEMENT TEXTURES

Pseudomorphism is the presence, in a mineral or mineral aggregate, of forms, textures, or structures *not* characteristic of itself but characteristic instead of *other* minerals, or mineral aggregates, or even of organisms. The recognition of pseudomorphism presupposes a knowledge of the forms, textures, and structures of minerals, rocks, and organisms and of the conditions under which these *normally* develop. Pseudomorphism results from the dissolving of one mineral or group of minerals and the depositing of another mineral or group of minerals in its place. Usually the processes of solution and deposition go on concurrently through the medium of the same solution and without the development of open spaces—that is, by the process known as replacement or *metasomatism*. The most delicate and perfect pseudomorphs are developed in this way. In other instances, however, minerals are dissolved by one solution, and after an interval, during which open spaces exist, new minerals are deposited in these spaces by solutions of a different composition. This is *not* replacement. In all but exceptional instances it preserves only the outward form—not the internal structure. The value of pseudomorphism as a criterion of replacement is greatly enhanced by the finding of unreplaced remnants of the host.

Features that are on record as preserved pseudomorphically in ores include:

In Minerals
- Crystal outlines
- Cleavage
- Twinning planes

In Igneous rocks
- Flow lines
- Form, size and pattern of grains
  - Porphyritic texture

In Sedimentary rocks
- Bedding
- Cross-bedding
- Stylolitic structure
- Form, size and pattern of grains
  - Oölites
- Organic structures

In Metamorphic rocks
- Schistosity

Interesting examples of the preservation of crystal outlines after replacement are found in the Black Hills of South Dakota where, in the Tertiary mineralization, dolomites have been extensively silicified. Under the microscope in polarized light the rock is seen to be an aggregate of interlocking quartz grains, but in ordinary light it has the appearance shown in Figure 1 of Plate 8; the crystal outlines of the original dolomite rhombs are clearly recognizable in spite of the complete silicification.

An example of preservation of crystal faces of calcite after replacement by the silver haloid iodobromite is shown in Figure 2 of Plate 8.

The preservation of mineral cleavage after replacement though not common is occasionally noted. Figure 12 is a camera-lucida drawing of ore from Neihart, Montana. This shows a polished surface of galena with the familiar triangular pits formed during grinding and polishing by the tearing out of small inverted pyramids of galena between the cleavage planes. The sides of these pits define the three directions of galena cleavage. Parallel to these cleavage directions the galena has been replaced by polybasite, and the position of the galena cleavage plane from which the replacement started is still recognizable under the microscope as a faint line running from end to end of the polybasite lenses. Similar relations are shown in another specimen from the same locality (Fig. 14).

Textures characteristic of igneous rocks are sometimes perserved in ores developed by replacement. Figure 13 shows a rich silver ore from Leadville, Colorado, in which the sulphosalt of silver, stephanite, has extensively replaced the matrix of a quartz porphyry while leaving the quartz phenocrysts intact and only partially replacing the sanidine.

Textures characteristic of sedimentary rocks are rather common in replacement ores. An excellent illustration of preservation of bedding is shown in Figure 3 of Plate 8, taken in the Union mine in the Black Hills, South Dakota. The entire face shown is ore, mainly quartz with minor amounts of pyrite, fluorite, and gold. The bedding of the original shaly limestone is still beautifully preserved in the ore.

Preservation of cross-bedding after replacement, though occasionally observed, is much less common than preservation of bedding, probably because the kinds of rocks in which cross-bedding is best developed—namely sandstones—are usually not easily replaced.

Stylolites results from uneven solution under mild pressure in sedimentary rocks, and except in rare instances they are formed only in calcareous sediments. Stylolites found in ores, therefore suggest that the ores are replacements of limestones or dolomites. Stylolites are common in the "blanket" deposits of fluorite in Southern Illinois

that are known to be replacements of limestone. The field shown in ordinary light in Figure 5 of Plate 8 is essentially all fluorite.

As indicated in a later chapter oölitic textures are common in many iron ores; the oölites usually are hematite but sometimes chalmersite. These ores were formerly considered replacements of oölitic limestones, but it is now known that in most of

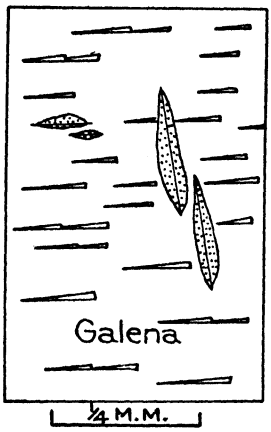

FIGURE 12

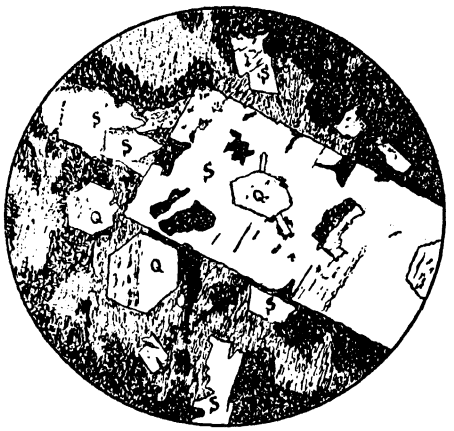

FIGURE 13

FIGURE 12.—*Replacement of galena along cleavages*

From Neihart, Montana; replacement along cleavage planes by polybasite. The position of galena cleavage is still recognizable in the polybasite as a faint line. After Bastin.

FIGURE 13.—*Replacement of an igneous rock by stephanite (St)*

In silver ore from Leadville, Colorado, porphyritic texture still conspicuous. S = Sanidine, Q = Quartz. × 36. After Fenner.

these ores the oölites are original sea-bottom deposits. Preservation of oölitic textures by the replacement of limestones by *silica* is, however, common. In the Baraboo District, Wisconsin, chert nodules have been developed in an oölitic limestone; within the nodules are oölites identical with those in the bordering limestone except that they are silica rather than calcite (Pl. 8, fig. 6). Similar evidence would be necessary to establish the replacement origin of oölitic hematite or chalmersite ores. The forms and to some extent the concentric structure of oölites are preserved in quartz and pyrite in certain gold-silver ores of the Black Hills, South Dakota (Pl. 9, fig. 1). It is known on other grounds that these ores are replacements of dolomitic limestones. The preservation of oölitic textures is corroborative evidence.

The forms and structures of organisms are not uncommon in ores that have replaced sediments. Replacements of wood by sulphides are particularly well shown in sulphide nodules in disseminated copper deposits in sandstones. In such occurrences it is important to distinguish between the filling of the cell interior and replacement of the cell walls. Figure 3 of Plate 5 shows ore from the Sacramento District, New Mexico, in which pyrite (white) has filled the cells, and chalcocite (light gray) has re-

placed the cell walls and filled small cracks. It has been inferred that the cells in the upper left of this figure were rotted and crushed at the time of mineralization, while the bulk of the specimen was not crushed. This suggests mineralization prior to deep burial.

One of the most striking instances of replacement of animal remains is afforded by the pyritic fossils of the Tully formation in New York. In the normal phases of this formation the fossils are calcareous, but over restricted areas the same species are pyritic. As no organisms are known to secrete pyritic shells pyrite must have replaced originally calcareous shells. The pyritic fossils average about one fifteenth the size of the same species in the normal Tully limestone. This dwarfing suggests that local high iron and sulphur content on the sea bottom in Tully times stunted the growth of the shells and also led shortly thereafter to their replacement by pyrite in the muds of the Devonian sea bottom. Obviously the interpretation of these pyritic fossils as due to replacement is based *not* on textural evidence but upon a knowledge of the composition of shells and the geologic conditions under which they are deposited.

Spurr observed an extraordinary instance of replacement of a gastropod shell by silver at Aspen, Colorado.

Edwards (1947, p. 28, Figs. 29, 30) has described and figured banded lead-zinc ores formed by replacement of schistose porphyry from Captain's Flat, New South Wales, and also by replacement of contorted slates from Mount Isa, Queensland. Microscopic study is often necessary to determine whether the mineralization preceded or followed the metamorphism. If it followed the metamorphism the metallic and other minerals introduced by replacement should show no deformation.

TRANSECTING OR CROSS-CUTTING TEXTURES

In the pseudomorphic replacements just described the forms, structures, or textures of the host are preserved and constitute evidence of replacement. In other instances no traces can be found in the guest mineral of the forms or textures of the host. If, however, the guest minerals *transect* features characteristic of the host, replacement can still be demonstrated. Figure 4 of Plate 8 shows a schist in which three pyrite cubes are developed. If they had antedated the development of the schistosity they should show fracturing and other evidence of deformation, and the schistosity of the rock should wrap around them. Instead, the pyrite crystals, unfractured, transect the schistosity. It is clear that schist once occupied the spaces now occupied by pyrite. The pyrite must therefore have developed by replacement after the development of the schistosity. Similar relations in gold ore from Timmins, Ontario, are shown in Figure 2 of Plate 9.

Structures transected by replacing minerals include typical grain patterns of igneous rocks, bedding, flow structure, schistosity, and organic structures.

In most cases of transecting textures the guest mineral develops its own characteristic crystal form as in the pyrite cubes of Figure 4 of Plate 8 and Figure 2 of Plate 9. These replacements are therefore *automorphic* in contrast to the pseudomorphic replacements described in the preceding section. Depending on the circumstances the same mineral species may replace the host pseudomorphically or automorphically.

(*Contrast* Figure 4 of Plate 8 with pyritized fossils.) The reasons for this contrast are not fully known. Some minerals, such as quartz, barite, and pyrite, have a much stronger tendency than others to develop their own crystal outlines in replacement.

GUIDED PENETRATION TEXTURES

GENERAL DESCRIPTION

Replacing minerals are often found along structural or textural features that would naturally be expected to guide penetrating solutions, such as fractures, bedding planes, schistosity, contacts between mineral grains of the same or different species, mineral cleavages, and mineral partings and twinning planes.

Although much replacement is initiated from visible openings, one of the most striking features of the process is the ability of the solutions to penetrate along directions of schistosity or contacts of mineral grains, where the highest powers of the microscope reveal no open channels. From such guiding features the solutions penetrate and replace the bordering minerals.

Figure 3 of Plate 9 shows anglesite (dark) associated with galena (light) from the Maury mine, Patagonia district, Arizona. Triangular pits in the galena indicate the three cleavage directions. The vein form of the anglesite and the fact that the veinlets follow cleavage directions of the galena show that it either replaces the galena or fills fractures in it. The ragged outlines of the veinlets and the failure of their walls to match is evidence of replacement. Figure 14 shows replacement of galena by polybasite, guided by galena cleavages and by the contacts of mineral grains. Commonly, as in Figure 14, only a single guest mineral is involved, but in other instances guided replacement involves a *succession* of replacing minerals as shown in Figure 15, in which galena was first replaced along quartz contacts and along cleavages by an unidentified mineral—probably a sulphide of lead and silver—and this in turn was replaced by argentite.

In other replacements one or more host minerals have been replaced by an aggregate of from two to many guest minerals. In the ore, shown in Figure 1 of Plate 10 galena has been replaced by argentite along contacts of quartz with galena and along contacts of differently oriented galena grains. In two places, however, segments of electrum (stippled) take the place of argentite. The replacements are believed to be hypogene since electrum is not known as a supergene mineral. Alternative explanations are possible. Electrum and argentite may have simultaneously replaced galena through the agency of the same solution. If so, it seems strange that the two very different minerals should have replaced the galena at the same rate, as shown by the fact that the galena border remains smooth in passing from contact with argentite to contact with electrum. A perhaps more acceptable interpretation is that the galena was first replaced peripherally only by argentite and that through the agency of later solutions argentite was replaced selectively and only locally by electrum.

In other ores (Fig. 16), comparable textures were developed by supergene enrichment. A zone of argentite and covellite has developed along the contacts between quartz and sphalerite. Although the replacement rim varies in width its width does not change abruptly in passing from argentite to covellite. We seem here to have the

alternative possibilities of (1) simultaneous replacement of sphalerite by covellite and argentite, or (2) replacement of sphalerite by covellite alone followed by later and localized selective replacement of covellite by argentite, or (3) peripheral replacement of sphalerite by argentite beginning along quartz contacts and followed

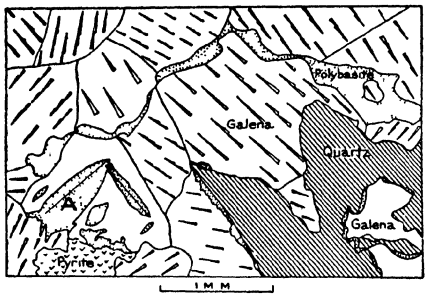

FIGURE 14

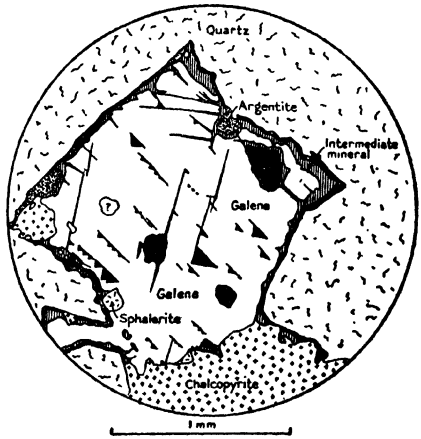

FIGURE 15

FIGURE 14.—*Guided replacement of galena*

Penetration of replacement along galena cleavages (at A), along contacts between galena grains, and along contacts between quartz and galena. Triangular pits developed in grinding and polishing define the cleavage directions in each galena crystal. At A replacement by polybasite has followed cleavages in the galena, and a faint line visible in the polybasite marks the position of the galena cleavage from which replacement started. Camera lucida drawing from lead-silver ore from Neihart, Montana. After Bastin.

FIGURE 15.—*Successive replacement of galena by two minerals*

Guided replacement of galena along quartz contacts and along cleavages by an unidentified mineral, probably a sulphide of lead and silver (intermediate mineral). This in turn has been replaced by argentite. Camera lucida drawing of ore from Liberty Bell Mine, Telluride, Colorado. After Bastin.

by selective, localized replacement of argentite by covellite. Nothing in the textures shown in Figure 16 justifies a choice among these three possibilities. However, in the laboratory sphalerite is not readily attacked by silver sulphate solution whereas covellite is. This favors the second as against the third explanation. In another ore (Fig. 17) galena was first replaced by an undetermined mineral x; this in turn has been replaced by a somewhat irregular association of argentite and chalcopyrite either simtaneously or in sequence. Most of the chalcopyrite follows the quartz-argentite conultacts, favoring the interpretation that the chalcopyrite is the youngest mineral in the section and is a replacement mainly of argentite and subordinately of mineral x.

While it is uncertain whether the relations illustrated in Figures 16 and 17 are the result of contemporaneous replacement by two guest minerals, simultaneous replacements by a group of several minerals are common, particularly in hypogene ore deposition. In the ore shown in Figure 2 of Plate 10 there are two groups of minerals, a relatively coarse group consisting of galena and quartz and a much finer-grained

group consisting of carbonate, argentite and chalcopyrite. The minerals of this finer group replace the coarse quartz and galena.

A very characteristic replacement texture is produced when a brittle mineral without cleavage, such as pyrite, is irregularly fractured and then partially replaced along

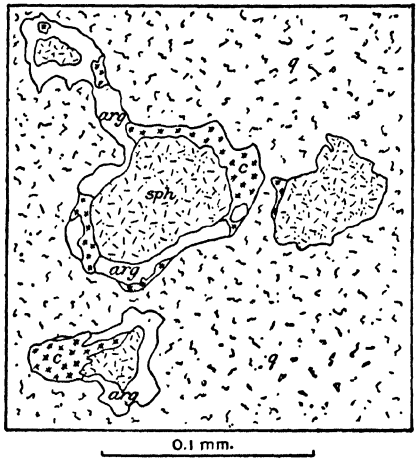

FIGURE 16

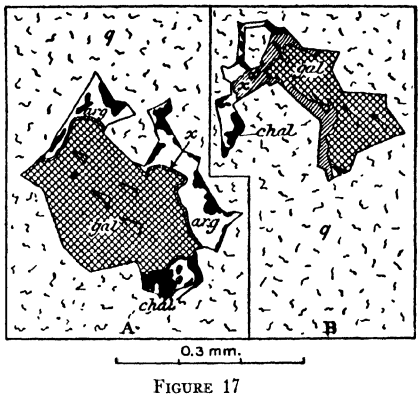

FIGURE 17

FIGURE 16.—*Replacement of sphalerite (sph) by covellite (c) and argentite (arg)*
q is quartz 1000 ft level of Tonopah—Belmont Mine, Tonopah, Nevada. × 350 After Bastin

FIGURE 17.—*Complex replacement of galena (gal)*
Along quartz (q) contacts Unidentified replacing mineral (x) is probably a lead-bearing silver sulphide This in turn has been replaced by an irregular association of argentite (arg) and chalcopyrite (black). Camera lucida drawing of polished surface. Jim Butler Mine, Tonopah, Nevada. After Bastin.

the fractures. On a small scale this texture is shown in Figure 5 of Plate 9 in which all stages from incipient to nearly complete replacement of Pyrite are shown. On a larger scale similar phenomena are shown in Figure 6 of Plate 9. None of the numerous pyrite areas are in contact with each other. Also, the walls of the veinlets between the pyrite areas do not "match", and groups of pyrite areas *cannot* be fitted together like the fragments of a jig-saw puzzle. All these features indicate replacement and are not harmonious with the alternative explanation of a simple filling of the interspaces of a breccia, although, as might be expected, in some instances both filling and replacement have taken place. In some instances faint lines along the middle of the veinlets mark the position of the crack along which replacement started. In still other instances replacement has proceeded further, and a second replacement mineral appears about midway between the areas of host minerals—*i.e.*, guest mineral No. 2 begins to develop where guest mineral No. 1 began to develop. *Replacement contacts are tight or sealed contacts and later replacements seldom if ever follow along them.*

PARALLEL ORIENTATION PATTERNS

The fractured and partially replaced pyrite shown in Figures 5 and 6 of Plate 9 has no microscopic character that defines the crystallographic orientation of the several pyrite areas. If such a character were present it would probably reveal that entire groups of pyrite areas were parallelly oriented. In galena, however, triangular pits show the crystallographic orientation of each grain, and frequently neighboring areas of galena now separated by such minerals as anglesite are all parallelly oriented (Pl. 9, fig. 4). It is inferred that the galena was originally one grain and that replacement encroaching along cleavage directions has subdivided it. Such parallel orientation patterns become a valuable criterion of replacement when supported by other evidences.

On a much larger scale, parallel orientation of bedding in isolated rock fragments may indicate replacement.

DIFFUSE PENETRATION AND AUTOMORPHIC REPLACEMENT TEXTURES

GENERAL DESCRIPTION

Nature draws no sharp boundaries between the types of replacement just described that have been conspicuously guided by structural and textural features of the host and other replacements in which control by fractures, mineral cleavage, contacts of mineral grains, etc., has *not* been conspicuous. Nevertheless, in many replacements the replacing solutions have evidently been able to penetrate the host with about equal facility in many or at least several directions, and it is sometimes difficult to determine why the guest mineral is where we find it with respect to the host. If fractures and grain contacts have guided replacement, these, as in the case of some fine-grained rocks, were so small or so numerous and diversely oriented that replacement proceeded about as readily in one direction as in another. Or the composition or porosity of the host or the composition of the replacing solutions were more important than structural or textural controls. Such replacements show characteristic textures quite different from the guided replacements already described, and it may be helpful to group them under a separate caption of "Diffuse penetration textures." As in many instances the guest minerals show their own characteristic crystal outlines, such *automorphic* replacements will also be considered.

Unquestionably the physical character of the host is often responsible for diffuse penetration. It is common in ores that have replaced fine-grained sediments as in those of Mount Isa (Pl. 11, figs. 1, 3). Occasionally the host is a fine-grained igneous rock such as felsite. The physical character of individual mineral grains of the host is sometimes important for Schouten (1946, p. 373, pl. 5) has shown that in the ores of Meggen, Germany, marcasite is more readily replaced than its associated pyrite.

In other cases fair-sized crystal grains have been penetrated by the replacing solutions in diffuse fashion because they lacked both fractures and mineral cleavage; a good example is seen in Figure 18 where silver replaces pearcite, which lacks cleavage. The pearcite has characteristic curved outlines concave outward as if the silver had "bitten" into it. Conversely, the rounded silver outlines, convex outward, seem to indicate movement of the silver-bearing solutions outward from numerous centers.

The chemical constitution of the host rather than its physical properties seems in many instances to have favored diffuse penetration, for some of the best examples of diffuse penetration are in ores in which the metallic minerals have replaced a carbonate gangue (Pl. 11, fig. 4; Pl. 12, fig. 4). In such cases the host was apparently so

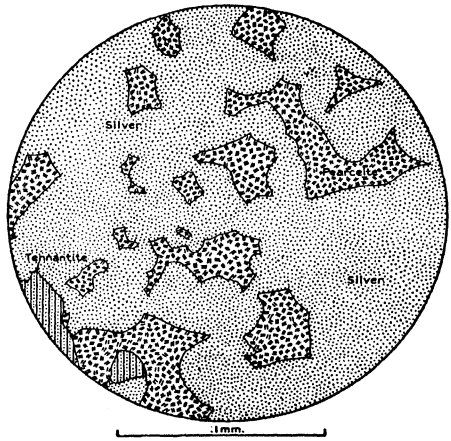

FIGURE 18.—*Native silver replacing pearceite*

Aspen, Colorado. Camera lucida drawing. × 360. After Bastin. The convex curved outlines of the silver against pearceite are noteworthy

easily replaced in all directions that the guidance of cleavage and grain contacts became trifling advantages. It is not uncommon, however, to find even in the same ore specimen both diffuse penetration and guided penetrations in the replacement of carbonates.

Finally, in many instances the nature of the *guest* determines whether the replacement is of the diffuse-penetration or guided-penetration type. Most *automorphic* replacements are partly independent of the structural or textural features of the host, and certain minerals such as pyrite, quartz, and barite have a stronger tendency than others to assume their own crystal outlines in replacement. These crystal outlines transgress the textural features of the host (Pl. 9, fig. 2). Some automorphic replacements are guided by *major* structural features of the host, such as bedding, but are independent of *minor* features.

One of the simplest forms of diffuse-penetration texture is exemplified in Figure 1 of Plate 11. Both automorphic crystals and spheroidal pellets of pyrite have replaced shale. Etching with acid reveals that some of the pellets are concentric.

Slightly more complex is the so-called "Atoll Structure" well displayed in the ores of Mount Isa, Queensland (Pl. 11 fig. 3). Here spheroidal masses are made up of a shell of pyrite within which occurs another sulphide, either galena, sphalerite, or chalcopyrite. All these sulphides are replacements. There are often breaks in the pyrite shell, and through such breaks the galena or sphalerite *within* the pyrite shell are often continuous with areas of the same minerals *outside* the shell. The other sul-

phides are considered younger than the pyrite. As mineralization proceeds they replace the pyrite until only remnants of the pyrite rims remain. They may also replace the shale. Rarely, at Mount Isa the "Atolls" occur in small groups somewhat resembling "tubercle" textures.

"Pellets" may be clustered as is beautifully shown in a Swedish silver ore (Pl. 11, fig. 2) in which clusters of spheroids of native silver (white) are surrounded by rammelsbergite, $NiAs_2$. Both have replaced calcite (black).

Next in complexity among the diffuse-penetration textures are those the writer has termed "tubercle" textures (Pl. 11, figs. 4–7). These textures are particularly well developed in the cobalt-nickel-silver ores of the Cobalt district and neighboring camps in Ontario, but they are also characteristic of the cobalt-nickel-silver ore type in the Erzgebirge of Saxony and of similar ore occurrences in several other widely separated localities.

To interpret properly the tubercle textures one must know the general order of deposition among the minerals as shown by *other* textural features. In the Cobalt district, Ontario, the order of deposition of the dominant ore minerals as shown by the relationships of veinlets, by the arrangement of minerals in small vugs or "nests" and by unquestionable replacement relationships is: older calcite, lollingite, niccolite and breithauptite, and native silver with younger calcite and minor polybasite. In the neighboring South Lorraine district the sequence as determined by the same criteria is: older calcite and a little quartz, safflorite with some arsenopyrite and cobaltite, nicolite with some tetrahedrite, and native silver and younger calcite. In both districts there is some overlap among the members of the sequence.

The simplest tubercle texture is illustrated by Figure 4 of Plate 11 showing an ore from South Lorraine in which the tubercles are composed of a single metallic mineral —arsenopyrite—in a calcite matrix. The arsenopyrite (at right) shows its own characteristic crystal faces against calcite both on the inside and on the outside of the tubercles. At the extreme right a few small isolated arsenopyrite crystals occur in the calcite. The form of these "tubercles" is not what one would expect from deposition in open spaces. They are interpreted as automorphic replacements of the calcite by arsenopyrite through the agency of solutions penetrating the calcite in diffuse fashion from several feeding centers.

Slightly later in the mineralization process in the same mine tetrahedrite was deposited by continued diffuse penetration as shown in Figure 6 of Plate 11. This is more highly magnified than Figure 4 of Plate 11; it shows clearly the automorphic outlines of the arsenopyrite on both the inside and outside of the tubercles and shows the deposition of tetrahedrite through the replacement of calcite both inside and outside the arsenopyrite tubercles. Tetrahedrite within the arsenopyrite tubercles is in places continuous with that outside. It conforms to the automorphic outlines of the arsenopyrite and is therefore somewhat younger. In other ores of Cobalt and South Lorraine that show tubercle textures, lollingite or safflorite rather than arsenopyrite formed the original tubercles, and niccolite and breithauptite rather than tetrahedrite formed the cores. In many instances native silver, niccolite, and breithauptite seem to have simultaneously replaced the original calcite cores of the tubercles. Much native silver is, however, somewhat younger than niccolite and

breithauptite (Pl. 11, fig. 5). Why tubercle textures should be confined mainly to ores of the cobalt-nickel-silver type is not clear; the presence of an easily replaceable calcite gangue is probably one factor, and the strong tendency of certain arsenides such as arsenopyrite, smaltite, and safflorite to form automorphic crystals may be another.

Keil (1931) has described from the cobalt-nickel-silver ores of Annaberg, Saxony, the textures shown in Figure 2 of Plate 12. Rather irregularly distributed crystals of native bismuth are enveloped in safflorite, and both lie in a quartz gangue. Beyond the main areas of safflorite in the upper part of this figure are several star-shaped twins of safflorite isolated in the quartz. It is difficult to see how isolated crystals of bismuth suspended in a tenuous mineralizing solution could have served as nuclei about which safflorite crystallized, and how isolated symmetrical star twins of safflorite could have remained suspended in a tenuous mineralizing solution and become enclosed in quartz. These textures resemble the tubercle textures found in the same ores and like them are interpreted as replacement textures. As a result of diffuse penetration of silica by the metal-bearing solutions scattered bismuth crystals were first deposited and safflorite was then deposited around them.

Ödman (1945) has described in a cobalt-nickel-silver ore from Sweden dendritic crystals of native silver surrounded by calcite. In Figure 4 of Plate 12 the boundaries of the calcite grains can be seen, and the silver crystals bear no relation to either the cleavage or the grain contacts of the calcite. Figure 5 of Plate 12, taken from ore from the same mine, clearly pictures replacement of calcite by diffuse penetration of silver-bearing solutions. The rudely tuberclelike outlines of some of the silver is noteworthy.

SKELETAL CRYSTALS OR DENDRITES

Skeletal crystals, as delicate as many that can be formed artificially by crystallization from concentrated solutions, form naturally in ores by replacement. Such forms are particularly common in the ores of the cobalt-nickel-native silver type both in North America and in Europe. Most of these skeleton crystals are either native silver or native bismuth rimmed by cobalt or nickel arsenides; the matrix is usually a mixed carbonate such as dolomite. Figures 1 and 3 of Plate 12 show the typical forms of the skeleton crystals of native silver enclosed by lollingite from Cobalt, Canada. As native silver is isometric the skeleton crystals usually branch at right angles and show octahedral faces. Native bismuth is rhombohedral and shows skeleton crystals quite unlike those of native silver.

Some misconceptions concerning these skeletal crystals can be corrected as the number of described occurrences has increased. In spite of the rather obvious enclosure of automorphic silver crystals by the arsenides some observers interpreted the silver as a later filling of the cores of arsenide dendrites. The arsenides however, have clearly replaced the silver of the dendrites in places. In Figure 3 of Plate 14 some of the silver contacts are ragged replacement contacts eating into the elsewhere straight boundaries of the silver. Furthermore some of the arsenides enveloping silver such as rammelsbergite and safflorite are orthorhombic and thus could not have determined the isometric form of the enclosed silver. In the cobalt-nickel-silver

ores of the Laver mine in northern Sweden some dendrites of native silver have arsenide envelopes (Pl. 14, fig. 3), but others (Pl. 12, fig. 4) are without any envelope of metallic minerals.

Van der Veen (1925), who studied many occurrences of skeletal crystals of silver and of bismuth, noted that in some occurrences in the Erzgebirge and at Cobalt, Canada, the cores within the arsenide envelopes were calcite rather than silver. Even in the same specimen some cores were calcite and others were silver. However, the cores of calcite were identical in form with those of silver. Van der Veen believed that the original dendrites were calcite formed by the alteration of vein dolomite early in the mineralization. The native silver he regarded as a replacement of original calcite dendrites. How isometric dendrites could be produced by rhombohedral calcite is not explained. Furthermore the accessory minerals in some of the calcite dendrites are chalcopyrite, galena, argentite, sphalerite, and rarely proustite, all known to be *late* minerals in these ores. Their testimony points to the caclite of the dendrites as one of the latest rather than one of the earliest minerals of the ores. Also Keil (1931) has shown that at Marienberg, Germany, and at Cobalt silver dendrites are partially replaced by calcite. In Figure 1 of Plate 14 within the arsenide envelope the calcite is usually peripheral, enclosing replacement remnants of silver. Similar relations in ores from Cobalt are shown in Figure 2 of Plate 14.

To check these age relations further Bastin re-examined a specimen of silver dendrites from Cobalt, Ontario, now in the University of Chicago collections and shown in Figure 3 of Plate 12. Megascopic examination of this hand specimen in reflected polarized light (through polaroid spectacles) shows that the matrix of the dendrites is an aggregate of diversely oriented carbonate grains and that the orientation of the dendrites is independent of either the cleavage or the grain boundaries of the matrix. The silver exhibits skeletal crystal forms of isometric pattern. The microscope shows that most of the borders of the silver are sharp, straight octahedral faces. Irregular borders are usually due to intergrowth with niccolite of a smaller order of grain size. Niccolite may occur even near the centers of the silver crystals but is most abundant just inside or just outside the silver border. The outlines of the very small niccolite inclusions are rotund, but the larger ones are automorphic against the silver. Clear evidences of replacement of niccolite by silver are very rare.

The thin envelope of arsenides around the silver is mainly lollingite (Pl. 12, fig. 3). The silver usually shows sharp crystal boundaries against lollingite and any overlap in age between silver and lollingite is slight. The inner part of the lollingite is much intergrown with niccolite, but the outer part is free from niccolite. Peripherally the lollingite is automorphic against carbonate.

In a few places the cores within the arsenide envelope are carbonate although they show the same forms elsewhere shown by the silver. There has thus either been a replacement of silver (and some niccolite) by carbonate or the reverse is true as Van der Veen concluded. In choosing an explanation it should be noted first that silver cores are the rule, and carbonate cores the local exception. Every gradation may be found from cores almost completely silver (with some niccolite) to cores completely carbonate. Within a single dendrite calcite may supplant silver for a short distance; then silver may come in again, all without change in the isometric

outlines of the dendrite. In places, areas of silver with the "ragged" outlines so often characteristic of replacement are enclosed by calcite. The writer agrees with Keil (1931) that carbonate has replaced silver; usually the small niccolite areas originally enclosed by the silver have also been replaced by the carbonate, but in places a few niccolite remnants remain.

Perhaps the clearest evidence that the isometric dendrites of the ores of the cobalt-nickel-silver type were originally silver and were later partly replaced by calcite is found in the ores of the Miller Lake-O'Brien mine at Gowganda, Ontario. The writer collected and studied representative silver ores from this mine in 1948. The wallrock is exclusively Nipissing diabase much of which has been extensively replaced by calcite. In this rock native silver occurs as fracture fillings and as replacements; and dendrites or skeletal crystals are well developed among the replacements. In cross sections of the dendrites (Pl. 13) three minerals are shown—native silver, lollingite, and calcite. Most of the silver dendrites have spider-like outlines, four or five irregular arms diverging from a center. In Figure 1 of Plate 13 the core of the dendrite is wholly silver. In Figure 2 of Plate 13 the core is mainly silver, but replacement (in part automorphic) by calcite has begun at the tips of two arms. In Figure 5 of Plate 13 more than half of the original core has been replaced by calcite. In Figure 3 of Plate 13 there has been spotty replacement of the silver core at six points. In Figure 6 of Plate 13 calcite has completely replaced the silver core.

Commonly the replacement of silver by calcite begins in the peripheral portions of the silver cores, often at or near the tips of the arms. Commonly small veinlets of calcite connect the calcite of the matrix with the calcite patches that are contiguous to the silver. These veinlets presumably represent the channels along which calcite-bearing solutions penetrated the lollingite to reach the easily replaceable silver. These veinlets are particularly well shown in Figure 3 of Plate 13 which also shows that the calcite usually replaces the silver automorphically.

SELECTED REFERENCES

Table 1 lists the described occurrences of dendrites in ores of the cobalt-nickel-silver type and shows that (1) Bismuth dendrites are enveloped by some of the same minerals that envelop silver. (2) Among the 22 silver ores listed, 15 have cores dominantly silver which is usually free from accessories. Five are dominantly calcite, and two are dominantly proustite. Where calcite is the dominant mineral of the core there are usually present as accessories one or more of the minerals galena, sphalerite, argentite, chalcopyrite, and rarely proustite, all of which are characteristically minor and late minerals in this type of ore; this suggests that calcite is also a late mineral. (3) A great variety of minerals may form the envelope of the dendrites. Of these, lollingite and rammelsbergite are orthorhombic, niccolite is hexagonal, and pitchblende is amorphous; none of these could possibly be responsible for the isometric forms of the silver dendrites. These relations support the interpretation that the original minerals of the dendrites were silver and more rarely bismuth and that the silver dendrites were not uncommonly replaced by calcite which in some instances was accompanied by galena, chalcopyrite, and proustite.

Some of the more important references on dendritic and other textures in ores of

the cobalt-nickel-silver type as listed in the bibliography are: Bastin (1922; 1925; 1949), Keil (1931), Montgomery (1948), Ödman (1945), Van der Veen (1925), Zuckert (1925).

RHYTHMIC OR DIFFUSION BANDING

Diffuse penetration sometimes results in the deposition of minerals in concentric bands with a more or less regular spacing that seems to imply a rhythm in the pre-

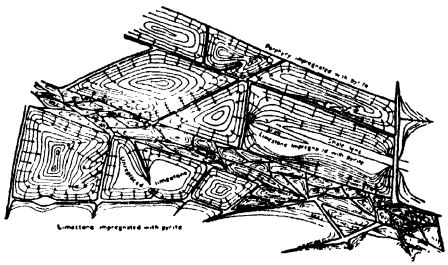

FIGURE 19.—*Diagram illustrating relation of diffusion banding to fractures* Leadville, Colorado. Arrows show directions of diffusion. After Loughlin.

cipitation process. Such banding has been termed rhythmic or diffusion banding. It is usually related to fractures in the host but is little influenced by lesser textural features. Figure 5 of Plate 14 shows diffusion banding developed during oxidation in an even-grained felsite from Tonopah, Nevada (Bastin and Laney, 1918 p. 45–46). The microscope reveals the cubical outlines of former pyrite grains that were distributed rather evenly through the rock. The pyrite grains have been completely oxidized, but the product of their oxidation, hydrous iron oxide, has not remained where the pyrite grains were but has become distributed in bands. No banding was produced in the *coarse* fragments enclosed in the felsite. The general directions of flow of the solutions that produced the bands can easily be inferred.

Diffusion banding is usually restricted to fairly fine-grained rocks. At Tonapah, oxidation banding occurs not only in the felsite shown in Figure 5 of Plate 14 but also in trachyte porphyry, a notably coarser rock. It is common elsewhere in sandstones and is found in some quartzites. It may also be developed in great delicacy in colloidal silica as familiarly exemplified by agates.

Most oxidation bands are yellowish brown; the coloring material is finely divided hydrous iron oxide. In the Precambrian quartzites of Baraboo, Wisconsin, the bands were presumably originally hydrous iron oxide, but this was dehydrated during the metamorphism of sandstone to quartzite to form fine red hematite.

Although little influenced by the textural features of the rock in which it occurs, oxidation banding is usually related to joints and other fractures much as is hypogene diffusion banding (Fig. 19).

Oxidation banding cannot be the product of a single episode of diffusion but represents the composite effects of many episodes of wetting and drying, solution and deposition, during prolonged periods.

TABLE 1.—FIGURED OCCURRENCES OF DENDRITES IN ORES OF THE COBALT-NICKEL-SILVER TYPE

Author and number of figure	Dominant mineral of dendrite	Accessory minerals of dendrite*	Rim or Rims of dendrite	General matrix	Locality
Van der Veen 31	Silver		Smaltite	Dolomite, Some Quartz	Freiberg
Van der Veen 32	Calcite	Chpy., Gal., Arg., Sph.	Safflorite		Same Mine
Van der Veen 33	Calcite	Gal., Arg.	Smaltite	Dolomite	Freiberg
Van der Veen 34	Calcite	Gal., Arg.	Smaltite		Same Specimen
Van der Veen 35	Proustite	Gal., Arg.	Ni, Ra, Saf.	Dolomite	Joachimsthal
Van der Veen 36	Calcite	Arg., Pr., Chpy.	Intergrown Ni and Dol.	Dolomite	Joachimsthal
Van der Veen 37	Silver		Feathery Arsenic		Annaberg
Van der Veen 60	Proustite		Feathery Arsenic		Joachimsthal
Van der Veen 109, 110	Silver	Few dendrites are calcite	Saf., Sm., Ni, Br.	Dolomite	Cobalt
Ödmæn 4 and 5	Silver		Rammelsbergite	Calcite	Laver Mine, Sweden
Bastin (1925) VII-C, D	Silver		Lollingite	Calcite	Cobalt, Canada
Keil 49	Calcite		Safflorite, Cobaltite	Calcite	Cobalt, Canada
Guild XIX-A, B	Silver		Smaltite	Calcite	Cobalt, Canada
Keil 20	Silver		Ra, Saf.	Calcite	Marienberg
Keil 46	Silver	Calcite replacing silver	Ra, Saf.		Marienberg
Neumann VII-14	Silver		Chloanthite	Calcite	Kongsberg, Norway
Krieger 7, 8	Silver		Nickel Skutterrudite	Qtz. Cal. Sid.	Bullard's Peak, New Mexico

TABLE 1.—*Continued*

Author and number of figure	Dominant mineral of dendrite	Accessory minerals of dendrite*	Rim or Rims of dendrite	General matrix	Locality
Bastin (1922) 35	Silver		Chloanthite	Quartz, Barite	Wickenberg, Arizona
Zuckert, Table 4, Fig. 14	Silver	Argentite	Safflorite	Quartz, Dol.	Joachimsthal
Montgomery, 3, 4	Silver	Calcite	Safflorite	Calcite	Gowganda
Kidd and Haycock, 74—1	Silver		Safflorite-Rammelsbergite	Carbonates	Great Bear Lake
Kidd and Haycock 74—3 and 4	Silver		Colloform Pitchblende, some Chpy.	Quartz, Some Carbonate	Great Bear Lake
Van der Veen 40	Bismuth		Safflorite	Dolomite	Schneeberg
Van der Veen 77	Bismuth		Saf., Ra.	Jasper	Schneeberg
Van der Veen 78	Bismuth		Rammelsbergite		Joachimsthal
Van der Veen 79, 80	Bismuth		Saf., Ra.	Dolomite, Some quartz	Schneeberg

* Abbreviations:

Arg.	Argentite	Ni.	Niccolite
Br.	Breithauptite	Pr.	Proustite
Cal.	Calcite	Ra.	Rammelsbergite
Co.	Cobaltite	Saf.	Safflorite
Chpy.	Chalcopyrite	Sm.	Smaltite
Dol.	Dolomite	Sph.	Sphalerite
Gal.	Galena		

Although diffusion banding is commonest as an oxidation phenomenon it is occasionally developed in hypogene mineralization. The hypogene examples are most common in limestones and more rare in the carbonate gangue of ores. An example from the coablt-nickel-silver ores of Cobalt, Ontario (Pl. 14, fig. 4) shows concentric bands of smaltite developed by the replacement of calcite. Miller (1913, Fig. 11) illustrates a specimen showing 25 such bands in a calcite gangue.

The rhythmic banding thus far described consists of bands of a single mineral alternating with remnants of the matrix which it has replaced. In other instances both bands consist of introduced minerals, and little if any of the original matrix remains. Such ores have been well described by Loughlin (Emmons, Irving, and Loughlin, 1927, p. 202–205). In this ore (Pl. 15, fig. 1) limestone has been replaced

by alternate bands of pyrite (light) and sphalerite (dark). Width and spacing of the bands are strikingly uniform. In the Leadville ores pyrite and sphalerite are the dominant minerals of the bands, but minor amounts of chalcopyrite or galena are in places present. Since, in some ores in other regions all the bands are of one mineral, possibly in the Leadville ores the pyrite bands were first formed and the sphalerite deposited later by replacement of the limestone remnants between the pyrite bands. Because the successive bands of sphalerite are occasionally connected across the intervening pyrite band by tiny sphalerite veinlets, Loughlin (Emmons, Irving, and Loughlin, 1927, p. 202–205) believes that at Leadville the sphalerite bands were formed slightly later than the pyrite bands. Whether such age diversity characterizes all double-banded ores is not yet certain. Double-banded ores almost identical with those of Figure 1 of Plate 15 form a minor feature of the fluorspar veins of Kentucky (Bastin, 1931). The bands are sphalerite and fluorite that have replaced vein calcite.

While diffusion banding is largely independent of the minor textures of the rocks in which it is formed it may be markedly controlled by joints and other fractures and sometimes by bedding. Fractures, indeed, usually represent the channels through which the replacing solutions entered the rock as is diagrammatically shown by the arrows in Figure 19.

Diffusion banding that partially simulates the natural occurrences can easily be produced in the laboratory. Liesegang (1913) allowed silver nitrate solution to diffuse into gelatine containing potassium bichromate and obtained a series of bands of silver bichromate spaced at progressively greater intervals away from the center of diffusion. Ostwald termed these bands "Liesegang rings" and believed that as the slilver nitrate solution diffuses outward a supersaturated solution of silver bichromate is formed; when the metastable limit of this solution is exceeded the silver bichromate is precipitated, forming the first ring. Continued outward diffusion of the silver nitrate causes renewed formation of silver bichromate solution in the zone surrounding the first ring until precipitation again ensues, and so on. As the source of silver nitrate becomes depleted, the solutions must diffuse farther before supersaturation is reached and the successive rings are deposited at increasing intervals.

These and other experimental productions of diffusion banding are instructive, but the conditions under which diffusion banding forms in nature may involve many other factors. Most artificial diffusion banding has been formed in gels that were inert media not reacting chemically with the diffusing solutions. The color banding in agates may not have involved chemical interaction with silica gel, but the development of diffusion bands in limestones or in carbonate gangues involved solution of the carbonates. The natural banding is not the product of a single episode of diffusion but the combined product of many episodes during which the diffusing solutions in many cases changed in concentration, composition, and temperature.

Double banding due to diffusion (Pl. 15, fig. 1) would once have been interpreted as due to successive deposition in open spaces and hence as implying sudden and frequent alternations in composition of the mineralizing solutions. Under the diffusion explanation no such unusual changes need be postulated.

Lindgren described very delicate diffusion banding in certain ores of Tintic, Utah.

The intervals between these bands are 1 millimeter or less (Pl. 15, fig. 2). The bands are often curved in beautiful concentric arrangement. The light bands consist of spherules of chalcedony cemented by granular quartz with undulatory extinction. The dark bands contain irregular grains of galena, rounded grains of sphalerite, and a few small cubes of pyrite. According to Lindgren colloidal silica first replaced limestone; sulphide-bearing solutions diffusing through the gelatinous silica then deposited sphalerite and galena in successive bands.

SELECTIVE AND NONSELECTIVE REPLACEMENT

As already indicated certain rock types, notably carbonate rocks, are particularly susceptible to replacement, while others, notably shales and slates, are usually resistant. Obviously, too, replacement is as much dependent on the composition of the mineralizing solution as upon the composition of the rock; for example, relatively pure limestones are usually particularly susceptible to contact-metamorphic replacement, but at San Jose (Bastin, 1937), Tamaulipas, Mexico, the intrusive rock and chert nodules in the limestone rather than the pure limestone have been metamorphosed. As the diorite is essentially free from quartz, solutions from the diorite magma were low in silica and probably rich in carbonates. They were thus incapable of reacting with the limestone.

In monomineralic rocks such as limestones there is, except for scattered impurities, no opportunity for selective replacement; such rocks are usually replaced en masse up to a rather definite boundary. In rocks made up of many minerals on the other hand replacement is likely to be notably selective. For example, pyrite developed by replacement in granite is often confined to biotite grains, the iron of the biotite apparently entering into the pyrite. The potash mica, sericite, selectively replaces the potash feldspar orthoclase. Replacement that is at first highly selective may be less so and spread to other minerals as the process goes on. In the ore shown in Figure 2 of Plate 10 replacement was at first confined to galena, but later, as at *A*, quartz began to be replaced.

In automorphic replacements in which the replacing mineral assumes its own crystal outlines notably different relations exist. The crystals of pyrite shown in Figures 4 of Plate 8 and Figure 2 of Plate 9 may have begun their growth as selective replacements of some iron-bearing silicate, but as the crystal grows its faces come in contact with many rock-forming silicates, possibly a half dozen or more. As the pyrite crystal grows, always maintaining its cubical form, all these rock silicates are replaced simultaneously and at approximately the same rate. The full explanation of such replacements is not yet clear.

GRAPHIC REPLACEMENT TEXTURES

Graphic intergrowths may be formed in several ways. They may be formed in igneous rocks as a result of simultaneous crystallization (Fig. 1), rarely they seem to have formed during contact metamorphism (Schwartz, 1931, p. 755, Fig. 18), and there is ample evidence that they are developed in ores by replacement. In an excellent example described by Lindgren from the Tintic district, Utah (Pl. 15, fig. 4), galena is traversed by bladelike crystals of barite and is in "graphic" association

with tennantite and pearcite. The barite has *not* been replaced by the galena, but it has been replaced, in transecting fashion, by the tennantite and pearcite. Therefore these minerals must also have replaced the galena.

Another excellent example of graphic texture developed by replacement (Pl. 15, fig. 3) pictures ore from the Engels Mine, Plumas County, California. Here the chalcocite in intergrowth with bornite can be traced into continuity with chalcocite that has replaced bornite along quartz contacts.

In general, the evidences cited for a replacement origin of graphic associations of the type just described are (1) boundaries that are serrate under high magnification and from which tongues of the guest project into the host, (2) veinlets of the guest traversing the host (Pl. 15, fig. 3), (3) occasional presence in the guest of a median line marking the position from which replacement started, (4) transitions from the graphic associations into unquestioned peripheral replacements.

Schwartz (1930) has described a graphic association of cerussite and covellite that is a replacement of an original graphic association of galena and stromeyerite.

Little or no system or regularity can be found in the graphic intergrowths formed by replacement.

LARGE DISPARITY IN GRAIN SIZE BETWEEN GUEST AND HOST

Under constant environment a group of crystallizing minerals tends toward moderate uniformity in grain size. Notable departure from such uniformity by certain minerals would, therefore, imply changed conditions, and hence age difference. The interpretation of phenocrysts of porphyries as differing in age from their matrix is a familiar application of this principle. There are notable exceptions to this rule, but it is valuable if used with caution.

In replacement in ores the guest minerals may be notably smaller in grain than those of the host (Pl. 10, fig. 2), or they may be notably larger (Pl. 9, fig. 2).

FORM OF THE GUEST-HOST CONTACT

In automorphic replacements the guest mineral develops its characteristic crystal faces, and its contacts with the host are smooth, straight crystal faces (Pl. 8, fig. 4). In pseudomorphic replacements the guest-host contact is likely to be curved or sinuous but may be either ragged or smooth, and the reason for the contrast is not always clear. In the ore shown in Figure 2 of Plate 10 very ragged contacts are developed in the replacement of galena, whereas in the graphic replacements (Pl. 15, fig. 4) the galena contacts are smooth. Both are believed to be hypogene replacements.

In some replacements the host shows scalloped outlines and appears "bitten-into" by the guest mineral. This texture is well shown in ore from Aspen, Colorado in which native silver contains abundant replacement remnants of pearcite (Fig. 18).

HYPOGENE VERSUS SUPERGENE REPLACEMENT

It is usually important in the study of ore deposits to determine whether replacements were accomplished by deep-seated (hypogene) or surficial (supergene) solutions. The criteria thus far used are few and are not textural for replacement patterns are identical in hypogene and supergene replacements.

The most readily available criterion is based upon knowledge of the geological and geochemical conditions under which various minerals form. This knowledge rests mainly on field observation and subordinately on laboratory experimentation. Certain minerals form only under the near-surface conditions of the oxidized zone. Replacements involving these minerals may safely be interpreted, therefore, as supergene. A common example is the development of anglesite and cerussite along galena cleavages (Pl. 9, fig. 3). Replacements by limonite and malachite are also supergene.

On the other hand, certain associations of lime silicates such as garnet, epidote, and wollastonite, which replace limestones, form only under conditions of hypogene mineralization. In certain rich ores of Tonopah, Nevada, electrum occurs as a replacing mineral. It is believed to be hypogene because in the zone of oxidation gold and silver tend to part company. Consequently, if they are redeposited from descending solutions they are deposited, not in alloy with each other, but as separate minerals such as argentite and relatively pure gold.

Many minerals can form under either hypogene or supergene conditions, and these include native silver, silver sulphides, and probably the sulphosalts of silver, whose origin it is often of great practical importance to diagnose.

In supergene replacement within the zones of oxidation and of downward enrichment two minerals seldom simultaneously replace the same host. They are usually in sequence such, for example, as galena first replaced by argentite and the latter by native silver. A few questionable instances of simultaneous supergene replacement by two minerals have, however, been described (Fig. 16).

In hypogene mineralization apparently simultaneous replacement of one or more host minerals by an aggregate of many guest minerals is common. Certainly the several minerals of the replacing aggregate show no evidences of any considerable age diversity. The ore from Tonopah, Nevada (Pl. 10, fig. 2), is an excellent example. Many great replacement ore bodies in limestone like those of Leadville, Colorado, afford larger-scale examples. It must not be inferred that sequences may not develop in hypogene replacement, but they appear much less characteristic than in supergene replacement.

Finally, depth relationships may be a criterion of hypogene vs supergene replacement. Replacements on an important scale at depths well below the effective circulation of meteoric waters suggest hypogene origin. On the other hand, the limiting in certain districts of some kinds of replacements to very moderate depths suggests supergene origin. At Aspen, Colorado, replacement of argentite and some other silver minerals by native silver is common near the surface but diminishes with depth and has not been noted below 1200 feet. The frequent limitation of chalcocite to depths of a few hundred feet is another example.

SYNTHETIC REPLACEMENTS

Replacements have been produced in the laboratory by a number of experimenters. Some replacements were incidental to experiments to elucidate the chemistry of natural replacement processes (Young and Moore, 1916). The most extensive were those of Ray (1930) carried on primarily to aid in understanding the natural processes of copper enrichment. Polished ore fragments composed of associations of

pyrite, bornite, chalcopyrite, chalcocite and sparse enargite were placed in a steel bomb with distilled water and heated for 3 to 5 days at temperatures of 100° to 175°C. All the specimens showed complete dispersion of small bornite masses into the enclosing chalcocite presumably as a result of solid solution; there was also incipient replacement of bornite by chalcopyrite and of pyrite by chalcocite. Next, similar specimens were placed in pyrex bombs with distilled water and heated in a water bath at 90°–100°C. for from 7 to 20 days. Changes in the specimens were identical with those obtained at higher temperatures. Finally in similar experiments with pyrex bombs one or more of the substances copper sulphate, sulphur, sodium carbonate, and ferric sulphate were added to the water and heated at 90° to 100°C. With these materials replacements were also readily obtained involving, in addition to those obtained with pure water, a replacement of bornite and chalcocite by covellite, of sphalerite by chalcocite, and of galena by covellite. In these experiments a number of natural replacement textures were imitated; delicate micrographic intergrowths of bornite and chalcocite and of sphalerite and chalcocite were produced.

The most elaborate and instructive experiments were conducted by the Dutch geologist Schouten (1934). He made no attempt to imitate closely in the laboratory the complex natural conditions under which replacement proceeds either as regards composition of the replaced substances, or the composition, temperatures, pressures, etc., of the replacing solutions. The purpose was to obtain clearly displayed replacement phenomena by *any* rapid and effective means, to study the textures produced, and to compare them with natural textures. Some of the textures produced have not been observed in nature.

Several of the principles of replacement deduced from the study of ores are confirmed by these experimental results. For example, the form of many of the synthetic replacements shows they were guided by fractures (Pl. 15, fig. 5) even where these fractures are too small to be detected by the microscope in the unaltered portions of the host mineral. Some small cracks present in the host were preserved throughout replacement, indicating that fractures though usually younger than their hosts may rarely be older.

In some experiments large automorphic crystals proved more resistant to replacement than fine aggregates of the same mineral (Pl. 15, fig. 6). Similar relations are not uncommon in natural replacements. Such relations show that automorphic outlines do not invariably indicate complete exemption from replacement. These observations are of great aid in interpreting the relations of silver to niccolite in ores from Cobalt, Ontario (Pl. 11, fig. 5). In these ores silver has clearly replaced niccolite extensively in the central portions of many of the "tubercles", and many of the contacts between silver and niccolite are typically ragged replacement boundaries. Yet some of the larger niccolite areas (smooth gray) retain their automorphic outlines.

While many synthetic replacements are guided by textural features of the host, others seem not to be so controlled. Figure 7 of Plate 15 shows the replacement of a phosgenite ($PbCO_3 \cdot PbCl_2$) crystal by galena without visible relation to the cleavage of the host. Somewhat analagous relations in natural replacements are illustrated in Figure 4 of Plate 12.

Pseudographic textures were developed by synthetic replacement of chalcopyrite

by bornite, and textures closely resembling some exsolution textures were produced by the synthetic replacement of bornite by chalcopyrite.

Many synthetic replacements were notably selective, and, according to Schouten, in general the attack on different minerals becomes more selective as the solutions are more dilute. Chalcopyrite was replaced by chalcocite synthetically much more rapidly than was pyrite—a fact commonly observed also in natural occurrences.

Volume relations in synthetic replacements are significant. Most natural replacements seem to have taken place with little if any change in volume. For the most part these replacements have occurred below the ground-water level under a fair amount of pressure due to overload. Schouten's experiments were carried out in covered dishes at atmospheric pressures and temperatures below 100°C. In some of the experiments notable increases in volume produced disintegration, and notable decreases in volume produced porosity. Figures 1 and 2 of Plate 16 show development of porosity in the synthetic replacement of cuprite by native copper obtained by treatment by a reducing solution of $Sn(OK)_2$ in KOH. The physical conditions of Schouten's experiments are more closely analogous to those of the zone of oxidation than of the deeper zones, and close study of the natural replacements of the oxidized zone may disclose that volume changes there are more common than at greater depths. In nature moving solutions and open chemical systems contrast with the quiescent closed systems in most laboratory experiments. Anderson and Merritt (1937), in studying the shallow replacements involved in the alteration of anhydrite to gypsum, in dolomitization, in serpentinization, and in the formation of bauxite, found no evidences of volume changes; the replacements were largely pseudomorphic.

Finally, synthetic replacement may offer a valuable new technique as an aid in the interpretation of microscopic textures, particularly those of supergene mineralization. In some cases hardly perceptible textures may be made clearly visible by synthetic replacement as an alternative to the customary etching methods.

CHAPTER VI. MICROSCOPIC INCLUSIONS

SOLID INCLUSIONS

Solid inclusions may be either younger, older, or the same age as the mineral that encloses them but it may be difficult to determine the age relation. The inclusions may be irregularly rounded or bladelike, or may have well-developed crystal outlines. Their sizes may be remarkably uniform or highly variable, their distribution even or erratic, regular or irrregular. Regularity in distribution may be related to the boundaries of the enclosing mineral or to its internal features such as cleavage, parting, and twinning planes. In the present state of our knowledge the interpretation of inclusions may best be presented through examples whose interpretation is fairly clear.

Certain inclusions have been formed by replacement and are therefore younger than the mineral that encloses them. Figure 3 of Plate 16 pictures an ore from Mount Isa, Australia, in which inclusions of galena occur in calcite. The white dotlike inclusions occur at the intersections of twinning planes and cleavages in the calcite. Proof of origin by replacement is afforded not only by this fact but by transitions from round inclusions to typical replacement veinlets that follow either cleavage or twinning directions.

Small inclusions of chalcopyrite in sphalerite are common in ores. Van der Veen (1925) has figured an occurrence in which these inclusions are distributed along the cleavage directions and the twinning planes of the sphalerite (Pl. 16, fig. 4). Quite possibly this, like the inclusions of galena in calcite just described, is a replacement. However, in this instance we have *not* a sulphide and a carbonate but two sulphides capable of forming a solid solution (Buerger, 1934). In the unmixing of a solid solution the minor component often develops along crystallographic directions of the major mineral. Possibly, therefore, the texture shown in Figure 4 of Plate 16 is due to unmixing, and therefore the chalcopyrite and sphalerite are contemporaneous. Similarly arranged inclusions of pyrrhotite in sphalerite have been described from the Malanas district in Sweden (Gavelin, 1939). The general criteria of unmixing have been discussed and apply to microscopic inclusions as well as to the textural patterns usually referred to as intergrowths. Unmixing is apparently limited to closely related minerals within the same chemical group such as oxides, sulphides, or elements. Solid solutions of oxides and sulphides are not known. Uniformity also characterizes *most* patterns due to unmixing. However, by reheating, Schwartz (1931) has produced unmixing textures in ores from Bisbee, Arizona, in which chalcopyrite is very irregularly distributed through bornite.

Inclusions may help interpret the age relations of ore minerals. Figure 20 is a camera-lucida drawing of an ore from the San Jesus mine, Zacatecas, Mexico. In the absence of any evidences of replacement the idiomorphic quartz is interpreted as older than the freibergite which conforms to its outlines. Unmixing appears to be ruled out because of the chemical dissimilarity of the three minerals. The localization of sphalerite inclusions in the peripheral portions of the quartz and in the Freibergite suggests that sphalerite began to crystallize after quartz crystallization was well

advanced and continued to deposit with freibergite after quartz crystallization had ceased. The paragenesis is interpreted as follows:

Quartz ————————
Sphalerite ————————
Freibergite ———

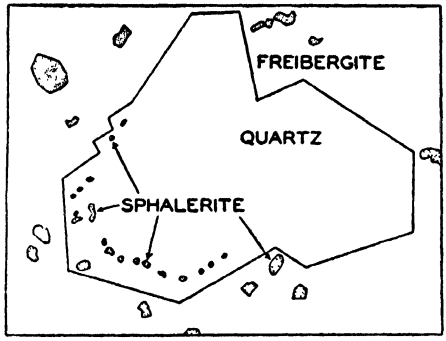

FIGURE 20.—*Inclusions of sphalerite in quartz and freibergite*
San Jesuo Mine, Zacatecas, Mexico. × 53. After Bastin.

Loughlin (Loughlin and Koschmann, 1942) has described from the Magdalena District, New Mexico, blebs of chalcopyrite in sphalerite that are confined to the outer portions of the sphalerite crystals and are roughly aligned parallel to the crystal borders. These inclusions probably had a *different* origin from the chalcopyrite inclusions of Figure 4 of Plate 16. Loughlin was inclined to ascribe them to replacement, but the evidence is inconclusive. They may well be of the same origin as the sphalerite inclusions in quartz shown in Figure 20.

FLUID INCLUSIONS

Fluid inclusions in ores have received less study than those in normal igneous rocks and in pegmatites. Usually they are detectable only in transparent gangue minerals and a very few metallic minerals such as light-colored sphalerite that are transparent in thin sections. In a few instances they have been recognized in opaque metallic minerals such as dark sphalerite and galena. Fluid inclusions in ores; however, deserve much more study because some of them undoubtedly represent minute samples of the solutions from which the ores were deposited. Even qualitative studies of such samples are of value.

Newhouse (1932) has reviewed the data on the occurrence of fluid inclusions in ordinary rocks and in ores. Most such inclusions are irregular but crystal outlines have been observed in a few fluid inclusions in quartz and galena. Buerger (1932) has described tiny cavities in galena bounded by octahedral and cubical crystal faces and filled with a sodium chloride brine. He believes "negative" crystal faces may have developed as normal crystal faces lining a slight depression in a galena surface; the depression then became boxed in or covered by rapid growth of an adjacent galena

crystal. A few of these cavities are up to 1 centimeter in maximum dimension. Newhouse (1932, p. 430) has observed similar cavities in galena from many other localities and has tested the fluids they contained.

The principal substances identified in fluid inclusions are:

Gases	*Liquids*
	H_2O
SO_2	CO_2
H_2S	Petroleum (rare)

Radicles in Solution		
Basic	Acid	*Solids*
Na	Cl	NaCl (in cubes)
K	SO_3	Bitumen
Ca		
Mg		

Newhouse studied 15 samples of galena and 4 of sphalerite from lead and zinc deposits in Europe and North America. In all of these the abundant basic radicles in the

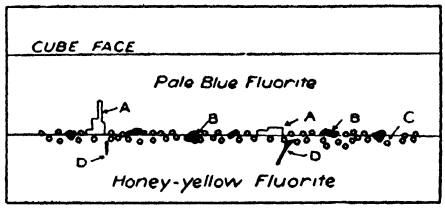

FIGURE 21.—*Inclusions of chalcopyrite, petroleum and marcasite in fluorite*

Negative crystal faces of fluorite (A) and inclusions of chalcopyrite (B), petroleum (C), and marcasite (D) along transition plane from yellow to blue fluorite. Cleveland Mine, Spar Mountain, Illinois. × 5. After Bastin.

fluids of the inclusions were sodium and calcium, and the acid radicle was chlorine. Calcium is decidedly subordinate to sodium. The concentrations are estimated at 12–25 grams of NaCl per 100 cc. or 4 to 9 times the percentage of NaCl in sea water. Even these roughly quantitative determinations of the composition of the fluid inclusions indicate that, if these are valid samples of the solutions from which the ores were deposited, their composition and high concentration rule out both meteoric and connate waters as agencies for the deposition of these lead and zinc ores. The alternative is a magmatic source.

Study of fluid inclusions seems to furnish a new method for determining deposition temperatures of certain ores. Newhouse (1933) in studying cleavage flakes and polished thin sections of light-colored sphalerite noted in the fluid inclusions an approximate constancy between the size of the inclusion and the size of the enclosed bubble. This suggested that the combined composition of gas and liquid was the same in all inclusions from any one specimen. Probably when the inclusions formed only one phase was present—the gas was completely dissolved in the liquid. On the stage of the microscope, Newhouse heated plates of sphalerite from 13 districts and found that, for any one district, the gas bubbles all disappeared at very nearly the same

temperature—for the Joplin, Missouri, district, for example, at from 125°–135°C., and for Southeastern Missouri at 105°–135°C. Newhouse believes these represent the temperatures of ore deposition. They support a hydrothermal origin for the Mississippi Valley lead and zinc ores. In applying this method polished surfaces are essential because of the high index of refraction of sphalerite.

An apparently contemporaneous association of fluid inclusions of petroleum and solid inclusions of marcasite and chalcopyrite are shown in Figure 21. They occur in the outer portion of a fluorite crystal at the contact of a layer of yellow fluorite with an overlying blue layer. They are approximately contemporaneous with the enclosing fluorite.

Primary and secondary inclusions must be carefully distinguished for only the primary inclusions give information on the composition or temperature of the mineralizing solutions. If planes of fluid inclusions cross grain boundaries without effect or change of direction they are almost certainly secondary.

Ingerson (1947) has pointed out that if there is reason to believe that the fluid inclusions being studied were formed under considerable pressures (as for example the fluid inclusions in the quartz of certain pegmatites) it is necessary to estimate these pressures and to apply a pressure correction before any reasonable estimate of the temperature of mineral deposition is possible. The composition of the liquid of the inclusion must also be considered.

CHAPTER VII. PARAGENESIS OR ORDER OF DEPOSITION

INDIRECT GEOLOGICAL AND GEOCHEMICAL EVIDENCE

The writer and others (1931, p. 576–577) have summarized the nature and value of this class of evidence as follows:

"The methods of reasoning employed to prove successive deposition may be conveniently classified under two types:

I. Indirect geological or geochemical evidence.
II. Direct textural or structural evidence.

"This division is not without overlap, however, since geologic concepts usually underlie even the textural and structural evidences. The first type of evidence rests upon the *a priori* postulate that for the formation of certain minerals and rocks markedly different environmental conditions (*e.g.*, temperature, pressure, solution composition, etc.) are necessary, than for the formation of others. This postulate is amply justified by a great body of observational data on mineral and rock occurrence, checked and supplemented by a lesser amount of laboratory data on mineral and rock synthesis. From such data we know definitely the conditions under which certain rocks have formed and it has also become possible to formulate a classification of minerals in accordance with their conditions of formation as first emphasized by Lindgren [1907] and later by W. H. Emmons [1908]. Many minerals may form through a considerable range of environmental conditions but it is clear that others form only through a comparatively narrow range.

"When minerals or rocks that require for their formation notably different environments occur associated, the inference is justified that a change in environmental conditions ensued after one and before the other formed, and hence that they are of diverse ages.

"An example of the use of this type of evidence in the case of ore minerals is afforded by associations of galena and anglesite. Although galena is deposited in nature under a wide range of conditions, it is not known to form under oxidizing conditions such as characterize the zone above the ground water level. Anglesite, on the other hand, is know to form only under oxidizing conditions. Furthermore, anglesite is commonly a product of the oxidation of galena. The association of these two minerals implies, therefore, a shift to highly oxidizing conditions after the deposition of galena and before the deposition of the anglesite—and therefore implies age diversity. Other familiar examples are the presence of garnet crystals in limestone and of calcite in the vesicles of basalt.

"Evidences of this type when used alone depend entirely for their validity upon the correctness of the geological or geochemical postulates. Fortunately in most cases textural or structural evidences of age diversity are also present."

AUTOMORPHIC OUTLINES

If the contacts of two crystalline minerals are in the main sharp and *straight* they probably represent crystal faces of one mineral or the other. Two alternatives then exist: (1) one mineral may be an automorphic replacement of the other, or (2) the minerals may have been deposited in succession, the younger mineral conforming to the crystal outlines of the older. The replacement alternative should be avoided unless evidence of replacement is definite, for even automorphic replacement usually betrays itself by spatial relations to fractures, by selecting certain minerals as host, or in other ways. In the absence of evidence of replacement, successive deposition must be postulated, and the question then is which mineral is the older. If mineral *A* shows crystal outlines characteristic of itself but not of mineral *B*, the problem is very simple. In Figure 1 of Plate 17 older pyrite (white) shows its characteristic cubical outlines against both quartz (dark) and chalcopyrite (gray). The quartz, in turn, shows its characteristic hexagonal outlines against chalcopyrite.

In other cases the problem is less simple. In Figure 22, mineral *A* may not show certainly identifiable outlines but its general outlines are *convex* or protruding outward, and its crystal faces join to enclose an isolated grain of *A* in *B*. Very probably

the crystal faces are those of mineral *A* rather than *B* and therefore mineral *A* antedates *B*. In Figure 23, however, crystal boundaries of *A* are in general *concave* outward, and probably these are crystal faces of mineral *B* rather than *A*. Examples of "concave" areas of a younger mineral in an older are shown in Figure 5 of Plate 16. Such areas although described for brevity as "concave" are bounded wholly or in

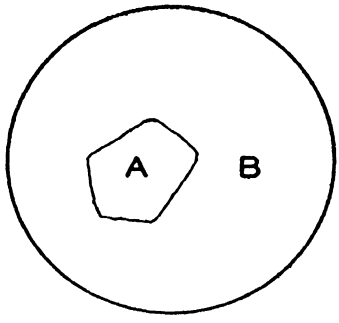

FIGURE 22

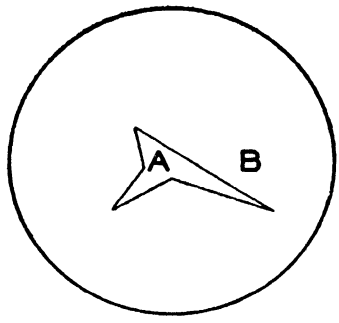

FIGURE 23

FIGURE 22.—*Diagram illustrating significance of protruding crystal boundaries*

FIGURE 23.—*Diagram illustrating significance of depressed crystal boundaries*

In Figure 22 (left) the outlines of mineral A, which are "convex" or protruding toward mineral B, indicate that A is probably older than B. In Figure 23 (right) the outlines of mineral A, which are "concave" toward mineral B, indicate that its boundaries are crystal faces of B and that A is the younger mineral.

part by crystal faces and must not be confused with the caries developed by replacement (Fig. 18).

PARTIAL AUTOMORPHISM IN TWO INTERGROWN MINERALS

Even in the same field of the microscope mineral *A* may in places show its own characteristic crystal outlines against mineral *B* and a short distance away the relations may be reversed. In most such ores there is evidence of only one generation of *A* or of *B*. Such cases usually indicate simultaneous crystallization of *A* and *B*. Such an interpretation is often supported by chemcial evidence, as for example in the case of certain intercrystallizations of cuprite and native copper formed by the oxidation of chalcocite according to some such reaction as $3\ Cu_2S + 4\ O_2 \rightarrow 2\ Cu + 2\ Cu_2O + 3\ SO_2$.

GRAVITATIVE CONTROL IN ORE DEPOSITION

Aside from its obvious control in the formation of sedimentary ore deposits gravity may be a factor in the deposition of other ore types. In some ore deposits the last minerals to be precipitated in vugs or other cavities were clearly deposited under the influence of gravity. Stalactites in the zone of oxidation are the most familiar examples of such deposits. In the fluorite deposits of southern Illinois layers of calcite or of barite crystals coat the ***upper*** surfaces of fluorite crystals in vugs but do not coat their flanks or downward facing surfaces (Pl. 16, fig. 6). The phenomenon is comparable to

the heaping up of snow on the roof of a house and its absence from the walls and under the eaves. Such deposits are obviously younger than the minerals which they coat and which show no relationship to the direction of the force of gravity.

SIMULTANEOUS DEPOSITION

The term simultaneous deposition has been used by geologists in somewhat diverse senses as indicated by such terms as "partial simultaneity", "essentially simultaneous" and "complete simultaneity." In the interest of clarity the term is here used in the description of ores in its narrow sense; *substances are considered to be simultaneous only if their precipitation from solution began concurrently and ended concurrently.* Under natural conditions simultaneous deposition, as so defined, is probably comparatively rare. Overlap is partial simultaneity, and evidence of overlap consists, in the last analysis, of evidence of simultaneous deposition in one part of the specimen and of successive deposition in an adjacent part. Any overlap should preferably be described under the caption of "overlap" rather than "partial simultaneity." It is a common relationship in contrast to the rarity of simultaneous deposition.

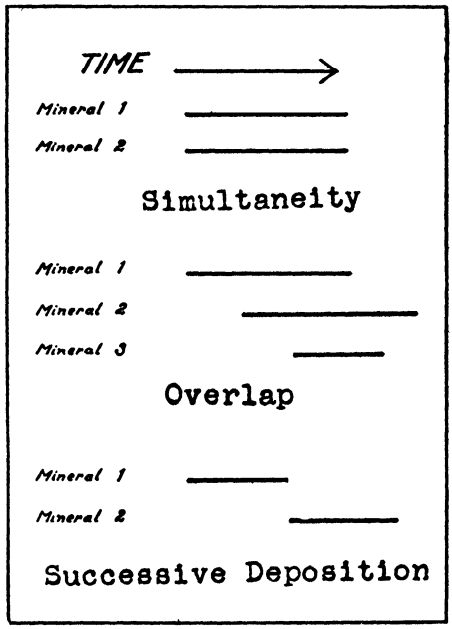

FIGURE 24.—*Paragenetic diagram illustrating preferred usage of the terms Simultaneity, Overlap, and Successive Deposition*

The age relations implied by the terms (a) simultaneity, (b) overlap, and (c) successive deposition, as used in this volume, are expressed diagrammatically (Fig. 24).

PARAGENETIC DIAGRAMS

In the simple form of paragenetic diagram (Fig. 25) the minerals of the ore are shown from top to bottom in the order in which their deposition appears to have *begun*; minerals that are simultaneous are obviously interchangeable in position. The width of the diagram is adjusted to the width of the printed page, and therefore in some diagrams involving many minerals the lines will be short while if few min-

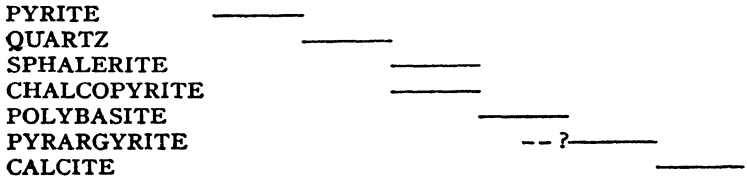

FIGURE 25.—*Paragenetic diagram, El Bote Mine, Zacatecas, Mexico*
Ore from 800-foot level. After Bastin.

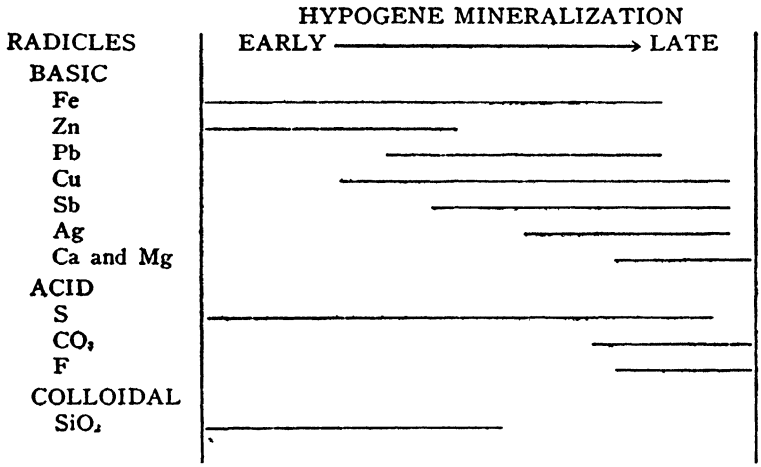

FIGURE 26.—*Composite paragenetic diagram for the Zacatecas district, Mexico*
In terms of the radicals, basic and acid, that make up the minerals of the ores. After Bastin

erals are involved they may be longer. Only the relative positions of the lines are significant.

In the interpretation of the microscopic textures of ores it is good practice to prepare a diagram of this sort for each polished specimen studied. From a series of such diagrams a composite paragenetic diagram for the ore body as a whole can be constructed for seldom are all significant mineral relations displayed in a single polished specimen. Comparison of many diagrams may bring out similarities and significant variations in the order of deposition.

Where mineralization has been fairly simple and uniform it may be possible to construct a composite paragenetic diagram for the whole district. Obviously many details are lost in such a graphic generalization, which represents only the more common relations for the district.

As an aid to understanding the chemistry of the mineralizing processes it is often helpful to translate a paragenetic diagram from terms of minerals into terms of the radicles which make up the minerals (Fig 26).

As there will be no overlap between (1) hypogene minerals, (2) minerals of the zone of downward enrichment, and (3) minerals of the oxidized zone, one can show in one composite diagram for an entire ore body or an entire district the paragenesis for all three depth zones, each zone being appropriately labelled.

Some geologists introduce the period of maximum deposition into the paragenetic diagram by substituting for all or part of the straight line a bulbous enlargement whose widest part represents the period of maximum deposition. So much uncertainty enters into such attempts that they seem to be of doubtful value.

CHAPTER VIII. TEXTURES OF SEDIMENTARY ORE DEPOSITS

GENERAL

Sedimentary ore deposits characteristically are associated with sediments of the more usual types commonly as beds or parts of beds; they usually show such characters as bedding, cross-bedding, fossils, ripple marks, etc. These features are familiar to all students in geology and are not described here. In using such features as evidences of sedimentary origin one must remember that they are sometimes preserved in ores formed by hydrothermal replacement of sediments—particularly limestones (Pl. 8, fig. 3).

Genetically the sedimentary ores may be mechanical deposits, as placers, or chemical deposits as many iron and manganese ores; biochemical processes sometimes play a role. To be properly classed as sedimentary they must have been syngenetic deposits in at least their early stages, but their metal content may have undergone subsequent redistribution by circulating solutions as well illustrated by the sedimentary "Red Beds" type of copper deposits in the Southwestern United States.

The oölitic textures common in many iron ores are described because they throw much light on the origin of such ores—more than has been generally recognized. Certain ore textures attributed to bacteria and algae are also briefly considered.

OÖLITIC TEXTURES IN ORES

GENERAL

Oölitic limestones afford the most familiar examples of oölitic textures and in early studies, oölitic iron ores were often interpreted as ferruginous replacements of oölitic limestones. As the oölitic iron ores have become better known it has become clear that the oölitic texture is an original character. Characteristic features of oölitic limestones have been described by Bucher (1918) and E. Cloos (1947). The present discussion is limited to the characteristics of oölitic iron ores.

Oölitic textures are fairly common among the bedded iron ores, particularly those composed dominantly of limonite or hematite. In most instances the typical well-rounded forms were developed on the sea bottom (rarely lake bottom) by aggregation of fine ferruginous material around nuclei under gentle wave agitation. In a few instances the forms have been modified by dynanometamorphism, or replacement has altered the original forms and composition.

The best illustrations and descriptions of the textures of the oölitic iron ores are those of Cayeux (1909; 1922) in his monumental volumes on the iron ores of France. The textural features of the French ores will therefore be described as a standard of comparison for those of other regions. From them much may be learned of the conditions of genesis of the ores.

OÖLITIC IRON ORES OF FRANCE

Oölites Defined. Cayeux defines oölites as small spherical or ovoid bodies of mineral matter of concentric structure and usually consisting of a core or nucleus enclosed by

a cortical envelope made up of concentric layers (Pl. 17, fig. 3). Bodies of similar size and shape often associated with oölites but lacking concentric structure Cayeux terms *false oölites*. The latter are mineral fragments or fragments of the hard parts of organisms that have been rounded by wave action but have not been appreciably enlarged by accretion (Pl. 17, fig. 4). Oölites are usually simple, composed of a single nucleus and its envelope. Rarely they are composite, two, three or even six simple oölites being surrounded by a common envelope. In some deposits composite oölites are absent, in most they are rare; in a few nearly all are composite.

Size of Oölites. Most oölites are 0.1–1.0 mm in maximum diameter; rarely a composite oölite or even a simple one may reach 2 mm.

Packing of Oölites. Oölites may be so closely packed that there is almost no matrix (Pl. 17, fig. 3) or very few oölites may be in contact (Pl. 17, fig. 5). Usually matrix is subordinate to oölites in volume.

Materials of the Oölitic Iron Ores. The same materials are found in oölites and matrix, but they are in different proportions, and their history has been different. They include siderite, chlorite, hematite, limonite, glauconite, magnetite and pyrite, and, as nonferruginous materials, silica, calcite, phosphate of lime, and a few others. Detrital minerals other than quartz and siderite are absent from most of the ores and are rare in others. The hard parts of organisms are often abundant.

The nucleus of the oölite may be angular or rounded. It is sometimes ferruginous material which may or may not be similar to that of the envelope. In other cases it is a foreign body such as a quartz or calcite grain or a fragment of bone or shell or a fragment of an older oölite. Only exceptionally is it impossible to differentiate nucleus and envelope. The size of the nucleus is extremely variable with respect to the entire oölite (Pl. 17, figs. 3, 6).

The envelope may be composted of one or more of the minerals siderite, chlorite, hematite, or limonite. Very rarely it may contain magnetite or pyrite. The concentric structure may be extremely fine and delicate or coarse and vague. It is usually most delicate in the chloritic oölites (Pl. 17, fig. 3). The iron oxides tend to mask the concentric structure.

The mineralogy of the oölitic iron ores of any one district is usually simpler than might be inferred from the comprehensive list of components given above for certain components are dominant and others rare or absent. For the French oölitic ores hematite is generally the dominant iron mineral in the Paleozoic ores and limonite in the later ores including the great deposits of Lorraine. Siderite, though subordinate to limonite, is of much importance in the ores of Lorraine (mainly as clastic grains) and not important in other French oölitic ores. In general carbonate-rich ores are also chloritic. Magnetite is of fortuitous occurrence in the Lorraine ores, and pyritic oölites are merely curiosities in the Oxfordian of Normandy.

Solidity of Oölites. Cayeux's illustrations show that where the ferruginous oölites are in contact each oölite maintains its smooth regularity of outline and is tangent to other oölites without indenting them. Evidence is lacking of mutual accommodations of form such as would be expected in plastic materials—as for example such accommodations as may be observed between biscuits in a pan. Therefore the oölites were

rigidly solid when deposited in contrast to the opinions of some writers (Bucher, 1918) that the oölitic envelopes were plastic, gelatinous, and colloidal. Much of the material that fed the growth of the oölites was probably very fine-grained but it seems to have been aggregated into solid oölites.

In a very few ores shallow cup-shaped depressions in the surface of one oölite may mark its places of contact with other oölites. This is clearly due to solution at points of contact subsequent to deposition and confirms the solid nature of the oölites since they have undergone no general deformation.

False oölites also testify to the mode of origin of the true oölites. The false oölites as shown at *A* and *B* in Figure 4 of Plate 17 are fragments of fossils or of minerals that have unquestionably been rounded mechanically, presumably by wave and current action. They lack the envelope of concentric layers that characterize the true oölites. They are, however, frequently associated with the true oölites (Pl. 17, fig. 5) and are similar to them in size and shape. Only in thin section are they readily distinguished. Indeed the distinction is purely one of definition for a false oölite may be regarded as an oölite composed wholly of a fossil or a mineral fragment. Oölites in a particular deposit tend apparently to build up by accretion to a certain order of size, while the larger shells and mineral fragments are worn down to the same order of size. Both are accomplished by wave action. There are transition forms—oölites with very thin concentric envelopes (*C* in Pl. 17, fig. 5).

False Oölites. Grains of various materials associated with the oölites and worn by wave action to the size and shape of oölites but lacking the concentric layers Cayeux terms "false oölites." In the formation of false oölites abrasion has been the dominant process. The distinction is no doubt justified as otherwise there would be no distinction between an oölitic deposit and, for example, a sanstone of rounded grains. In quantity false oölites are usually very subordinate to the oölites, but some Devonian iron ores contain *only* false oölites.

Most false oölites in iron ores show the organic structures of mollusks or of crinoids but may be partly or wholly ferruginous; there is every gradation from nonmineralized to completely mineralized false oölites. First there is usually a selective replacement of certain parts of the cellular material which tends to emphasize its structure; then there is a mass replacement penetrating inward from the periphery of the false oölite, that completely destroys the organic structure as shown at *A* and *D* in Figure 4 of Plate 17. If the false oölite were partly or wholly coarse calcite it may be partly replaced by hematite with the development of very ragged boundaries as shown at *F* in Figure 4 of Plate 17. Finally the entire false oölite may be converted to hematite, the process usually being completed first in the smaller false oölites as at *G*. Usually in such ores the original surface outlines of a shell or other organic material has been destroyed or greatly modified by abrasion due to wave action. Figure 5 of Plate 17 pictures on a smaller scale oölitic ore from another French iron ore deposit. It shows some of the same features exhibited in Figure 4 of Plate 17. There are remnants of crinoids in the interior of some of the oölites and false oölites at (*A* and *C*). Unlike Figure 4 of Plate 17, however, are abundant crinoidal remains in the matrix that have been partly replaced by hematite.

In Figure 6 of Plate 17 the oölites are nearly wholly hematite so dense that no concentric structure is visible. Oölites with nuclei of irregular quartz grains are shown at *A*. The cement is siderite.

Organisms. The hard parts of organisms found in the iron ores of France include fish remains, mollusks, crinoids, and Foraminifera. Others occur but only subordinately. Fish remains are present in all the ores but are particularly abundant in those of Lorraine. Mollusks and crinoids are in general the most abundant organisms. Foraminifera though widespread are in general rare and are mostly sea-bottom forms. All these fossils have been replaced in diverse degrees by various iron-bearing minerals and by silica. Boring algae represented by tubes and casts are not known to have been important in the formation of the ores.

Matrix or Cement. The matrix is usually subordinate in volume to the oölites. Its materials are in general the same as those composing the oölites but in different proportions. Calcite, often rhombohedral, is frequently abundant instead of rare as in the oölites. Clay also is a component. Cements composed largely of organic debris are extremely rare.

Conditions of Deposition. The smoothly rounded outlines of the oölites and false oölites clearly indicate prolonged agitation by wave action. The nature of the fossils testifies to marine waters. Wave action was severe enough to comminute the remains of mollusks and crinoids and occasionally to break oölites, the fragments becoming the nuclei of new oölites. These features as well as evidences of contemporaneous erosion point to deposition in moderately shallow seas, and the great lateral extent of many of the deposits shows that these seas were extensive.

Changes Subsequent to Deposition. Cayeux figures (Vol. II, Fig. 24) an ore from Lorraine in which wholly chloritic oolites have been deformed by unilateral pressure into irregularly spindle-like forms while the mechanically more competent grains of clastic siderite and phosphate of lime have not been distorted. Such pressure phenomena are, however, very rare among the French oölitic iron ores, whereas deformation of oölites in limestones is fairly common as described by E. Cloos (1947).

Replacement subsequent to the emplacement of the oölites has produced minor changes in some of the ores of Lorraine. In Figure 6 of Plate 17 secondary chlorite forms thin layers between the hematite of the oölites and the siderite of the matrix. Many of the hematite oölites are in contact, and at such contacts no chlorite is present. The chlorite therefore is not a part of the oölites but has been developed by replacement of the siderite of the matrix subsequent to the deposition of hematite oölites in their present position.

In a Silurian oölitic ore the originally smooth outlines of chloritic oölites have been rendered irregular through replacement by siderite beginning at the original periphery of the oölites and penetrating inward to varying depths (Pl. 18, fig. 1). The irregular inner border of the siderite in places cuts across the banding of the chlorite layers, and therefore it formed after the banding was developed. Another picture of the same ore confirms the originally chloritic nature of the oölite envelope, for where oölites are in contact there is a smooth contact of chlorite with chlorite with no intervening siderite.

The originally smooth outlines of oölites may also become irregular through

secondary enlargement. Hematite may develop in the matrix in continuity with that of the oölite and destroy its regularity of outline.

CLEVELAND IRON ORES OF ENGLAND

The most important iron ores of England, the Jurassic ores of the Cleveland Hills in North Yorkshire, have been comprehensively described by Hallimond (1925) and are characteristically oölitic. Evidence is decisive that they were deposited essentially in their present condition. They differ from most other oölitic iron ores in that the iron is largely in the ferrous state as *chamosite* (a ferrous aluminous silicate) and *siderite*; the ferric oxides are rare or absent except as products of recent weathering.

In the Cleveland ores the oölites average about one-sixtieth of an inch in diameter and are composed of chamosite and siderite. The chamosite is very fine-grained and is concentrically banded, but the siderite is granular and has usually replaced the chamosite beginning at the smooth periphery of the oölite. The inner border of the siderite zone is irregular. Some oölites are wholly chamosite, but oölites wholly siderite are rare. Some oölites have been replaced by calcite showing parallel orientation throughout a group of contiguous oölites. Many oölites have no recognizable nucleus, but others contain nuclei of shell fragments or broken oölites or crystals of chamosite. The matrix consists of siderite rhombs in a groundmass of fine chamosite.

Concerning the British oölitic iron ores as a whole Hallimond says: (p. 10)

"The oöliths themselves are, in grading and distribution, practically equivalent to sand-grains. They are frequently *polished* and accompany polished sands. A striking feature is their uniform grading, examples of small and incompletely developed oolites being very rare, and this fact seems most consistent with the view that growth in the later stages was extremely slow and tended to reach a limit when the increased weight of the particle resulted in a degree of mechanical erosion that balanced the chemical precipitation."

OÖLITIC IRON ORES OF NEWFOUNDLAND

Among the best examples of oölitic iron ores on this continent are the Ordovician ores of Wabana, Newfoundland (Hayes, 1915). The ores are principally hematite and chamosite—an aluminous ferrous silicate. Siderite, abundant locally, is mainly confined to the matrix. The hematite is very finely divided and occurs principally in the oölites and subordinately in the matrix. In the oölites it is concentrically banded and is invariably associated with chamosite. When abundant it may mask the chamosite, but if the hematite is dissolved by acid the chamosite is always present. Usually the two minerals form alternate layers. In the leaner ores the envelopes of some oölites may be wholly concentric concentric layers of chamosite. In general chamosite is next to hematite in abundance.

Most of the oölites have a nucleus which may be an angular grain of quartz (shown to be detrital by its fluid inclusions and its apatite prisms), shell fragments, or chamosite or siderite. The matrix is predominantly siderite but may contain also hematite and chamosite as well as shell fragments. Siderite sometimes has replaced hematite and chamosite, but the reverse relation was never observed. The shell fragments are largely inarticulate brachiopods and hence are phosphatic. Fossil tubules of boring algae occur in oölites and in shell fragments.

Ripple-marked surfaces in all the workable ore beds and the occasional presence of

cross-bedding, indicate shallow water deposition, and graptolites and brachiopods indicate that the waters were marine. The common alternation of chamosite and hematite layers in the envelopes indicates that the two minerals are in the main contemporaneous though in places hematite has clearly replaced chamosite. Such replacements were diagenetic, taking place while the sediments were still unconsolidated. No important amounts of iron seem to have been added since sedimentation ceased.

Oölitic Pyrite Deposits. Minor deposits of pyrite in the Wabana district, although of no economic importance are one of the few described occurrences of oölitic pyrite. Oölitic pyrite (Pl. 18, fig. 3) occurs in one to three beds lying 1 to 10 feet stratigraphically above the highest hematite ores. There are no gradations from pyritic to hematitic ores, and the two resemble each other only in texture. In some of the oölites layers of pyrite alternate with what appears to be calcium phosphate. Fragments of brachiopods have been replaced to varying degrees by the pyrite. In contrast to the hematite ores graptolite remains, some replaced by pyrite, are common, indicating access to the open sea and its pelagic faunas. Unpyritized fossil fagments are often in contact with completely pyritized fragments indicating that these constituents were brought together mechanically on the sea bottom after the formation of the pyrite.

The pyritic beds alternate with fissile black graptolite-bearing shales. No pyrite masses transgress the bedding. There is no evidence that the pyrite beds are replacements of hematite-chamosite ores; Hayes considers that they are as truly original sediments as are the oölitic hematites, but deposited in deeper water where the development of hydrogen sulphide by organic reactions on the sea bottom led to the formation of iron sulphide and produced acid conditions inhibiting development of carbonates.

OÖLITIC IRON ORES OF THE UNITED STATES

Newland and Hartnagel (1908) have described the oölitic iron ores of the Clinton formation of New York, which are dominantly earthy hematite and are in part oölitic. The oölitic grains are spherical or somewhat flattened and seldom over 1 mm in diameter. There is usually a nucleus—commonly a quartz grain—and an envelope of concentric hematitic layers. If the hematite of the oölites is dissolved by HCl a gelatinous mass of transparent silica remains having the form of the original oölite The matrix is usually granular calcite. The oölites are usually closely packed. The rounded quartz nuclei, rarely over half the diameter of the oölite, contain fluid inclusions and small crystals of rutile and hematite. Their source is probably the Precambrian crystalline rocks. In some of the Clinton ores the hematite is mainly a replacement of the hard parts of bryozoans, crinoids and brachiopods and false oölites may be more abundant than oölites (Pl. 18, fig. 2).

Some of the Clinton ores are directly overlain by limestone which shows no evidence of replacement by hematite and is not oölitic. Newland and Hartnagel interpret the ores as original sea-bottom deposits.

The most detailed studies of the microscopic characters of the Clinton iron ores of New York are those of Alling (1947). The Clinton ores are lenticular and not confined to a single horizon within the formation. The textures are characteristically oölitic;

in the western part of the ore zone the nuclei are almost exclusively fossil fragments; in the eastern part nuclei of rounded quartz grains become predominant. The microscope shows that the fossil fragments were rounded, apparently by wave action, and their small openings were filled by calcite, clay, and phosphatic material. Later the

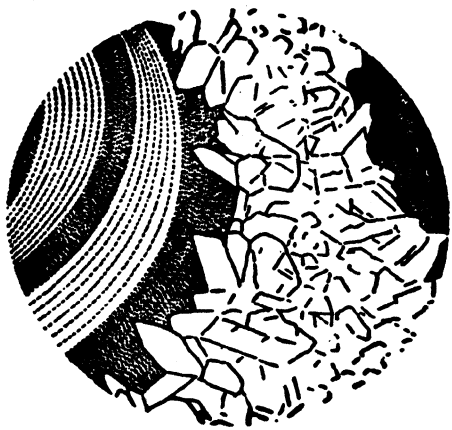

FIGURE 27.—*Automorphic replacement of iron silicate oölites by siderite*

From near Priston, Kentucky. After Bucher. These relations are interpreted by Bastin as indicating automorphic replacement of solid iron silicate oölites by siderite as indicated by the transecting of the iron silicate laminae by the siderite crystals.

fossiliferous material was replaced in part by hematite and coated by concentric layers of chamosite-hematite often several hundred in number. All these processes are believed to have taken place on the sea bottom before the consolidation of the deposit and therefore were *diagenetic* changes. The source of the iron still remains uncertain.

In the Birmingham district of Alabama as in New York both oölitic ores and ores that are largely replacements of fossil fragments occur (Burchard, 1910). The oölitic ores are aggregates of somewhat flattened grains, $\frac{1}{16}$ to $\frac{1}{10}$ inch in diameter, lying with their flatter dimensions parallel to the bedding. The oölites usually consist of a nucleus of rounded quartz and an envelope of successive layers of hematite associated with some siliceous and aluminous material. The matrix consists of hematite with some calcite. The so-called "fossil ores" of Birmingham consist of the fossil remains of bryozoans, crinoids, corals, and brachiopods many of which are broken and water worn. Usually either fossil or oölitic ore predominates in a bed, but in places they are mixed in nearly equal proportions.

In a few of the American oölitic ores replacements subsequent to deposition appear to have modified the ores. In an oölitic iron ore from Kentucky (Bucher, 1918, p. 598–601), in which oölites of iron silicate lie in a siderite matrix (Fig. 27), automorphic crystals of siderite project into the oölites. Bucher interprets this as evidence that the oölites were gelatinous and yielding, but the laminae of the oölites are not deformed

but are sharply transected by the siderite crystals probably indicating automorphic replacement of solid iron silicate.

EVIDENCES OF ALGAL AND BACTERIAL REMAINS

The causes of the deposition of the great sedimentary deposits of iron ores that constitute the World's greatest reserves have been a persistent and vexing problem that is made more difficult because comparable ores seemingly are not being deposited today. Because some living algae and Bacteria can deposit hydrous iron oxides from ferruginous solutions (Harder, 1919) there has been much interest in the occasional occurrence of seemingly fossil algae and bacteria in or closely associated with some sedimentary iron ores. Conclusive evidence is still lacking that they played a major role in ore deposition.

Cayeux (1909, Figs. 33–35) figures excellent examples of boring algae in both hematite oölites and matrix in a Silurian iron ore. Hayes (1915, Pl. XXI-B) describes and figures the tubules of boring algae from the oölitic hematites of Wabana, Newfoundland.

Gruner (1924) has described and figured what appear to be bacterial and algaforms in cherts from beds stratigraphically below the Huronian Biwabik iron-bearing formation of the Mesabi range in Minnesota (Pl. 17, fig. 2)

CHAPTER IX. ORE TEXTURES FORMED BY DYNAMOMETAMORPHISM

Regional metamorphism is unfavorable for the *formation* of ore deposits but has in certain instances deformed pre-existing deposits and developed pronounced foliation in the ore. The changes produced are largely textural and mineralogic; there is little change in gross chemical composition, and transfer of material has taken place only through very limited distances. Ores deformed by dynamometamorphism may look much like ores that have been formed by the replacement of metamorphic rocks, but in the latter only the wall rocks and not the ore minerals show evidence of deformation.

In Europe the ores of Rammelsberg are an excellent example of regionally metamorphosed ore deposits. These ore bodies are enclosed in slates of Devonian age that have undergone considerable distortion. The ores consist, in approximate order of abundance, of sphalerite, chalcopyrite, galena, pyrite, and arsenopyrite with a barite gangue, and in places are much contorted. (Fig. 28). Lindgren and Irving (1911) have shown that the ores have been regionally deformed with the development of a gneissoid structure. Pyrite and barite—the components most resistant to deformation—form lenticles or nodules around which wrap bands of the weaker minerals galena, sphalerite, and chalcopyrite. Figure 4 of Plate 18 shows nodules of a relatively resistant coarse intergrowth of barite and sphalerite around which the weaker metallic minerals are moulded as thin laminae.

Among the common ore minerals galena is particularly susceptible to deformation. In the initial stages such deformation may manifest itself in hand specimens by curved cleavage faces, and under the microscope in polished specimens by curvature in the lines of the characteristic triangular pits. More intense deformation reduces the galena to an aggregate of minute grains and develops planes of flowage forming a schistose mass commonly termed "steel" galena (Pl. 18, fig. 5). Where more resistant minerals are associated with the galena the latter wraps around them in the fashion similar to that shown in Figure 4 of Plate 18.

In some cases the harder minerals of the ore are broken by the differential pressure, and the fragments are dragged apart in the flowage of the steel galena.

Movements within the ore minerals under differential pressures are not random adjustments but have been shown (Buerger, 1928) to be in part due to slipping or translation along definite crystallographic planes and in part to reorientation tending to bring certain crystallographic directions into parallelism with the direction of maximum pressure. Galena was found particularly susceptible to deformation, and pyrite particularly resistant. Sphalerite, chalcopyrite, and pyrrhotite were intermediate.

In experiments on the deformation of ores and ore minerals under pressure (Buerger, 1928; Newhouse and Flaherty, 1930; Edwards, 1947, Fig. 33) certain metallic minerals that seldom show original or primary twinning may develop conspicuous *secondary* twinning under very mild external pressures that are insufficient to shear, brecciate, or granulate them. Sphalerite and chalcopyrite not uncommonly exhibit such twinning. This twinning must be used with caution as evidence of stresses of

external origin since it *may* also develop as a result of mild internal stresses such as those incident to the unmixing of solid solutions (Buerger and Buerger, 1934). Etching with appropriate reagents and/or examination in reflected polarized light is usually necessary to recognize such secondary twinning.

FIGURE 28.—*Contorted banding in ore*
From Rammelsberg, Germany. a = pyrite, b = chalcopyrite, c = galena. × 1. After Wiechelt.

Finally, certain textures exhibited by some of the native metals—some original textures and some the result of mild degrees of thermal or of dynamic metamorphism—have been investigated by British metallurgists (Carpenter and Tamura, 1928; Carpenter and Fisher, 1930; 1932) in a valuable study applying to the naturally occurring metals the methods used in the study of manufactured metals and alloys. Such studies may be illustrated by those on silver which are of especial geologic interest.

Native silver recrystallizes so readily under even mild pressures that in the grinding of specimens preparatory to polishing for microscopic study the surface layers undergo a distortion which masks the true texture of the metal beneath. By repeated etching this surface layer can be removed, and further gentle polishing combined with etching reveals the original texture. Twenty-one specimens of native silver from 14 localities widely scattered over the world but mostly from Canada and the United

States were prepared and studied under the reflecting microscope and by heat treatment (annealing). Some of the specimens showed a homogeneous granular texture (Pl. 18, fig. 7) identical with that developed artificially in silver by annealing above 200°C. Silver from other localities had been partially recrystallized, and still other silver (Pl. 18, fig. 6) showed no recrystallization but was characterized by zonal structures. The silver that on geological evidence appeared to be supergene showed evidence of only partial recrystallization or none at all. In contrast, the silver from Cobalt, Ontario, and neighboring camps of the same type, now generally interpreted as hypogene, was all of the completely recrystallized type. Carpenter and Fisher (1932) conclude that this silver recrystallized in the presence of hot hypogene solutions above the normal atmospheric recrystallization temperature of silver (about 200°C.).

While temperature may well have been the dominant control for these contrasting textures, both time and pressure may have been collaborating factors, inasmuch as supergene native silver is invariably of shallow and recent origin, whereas the silver occurrences of Cobalt and neighboring camps are Precambrian and of more deep-seated origin.

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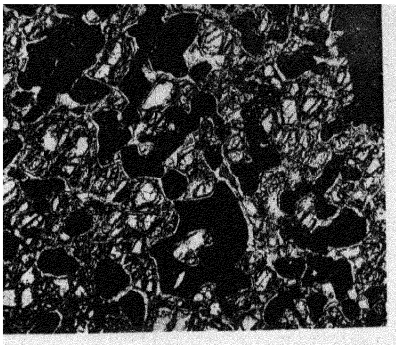
EXPLANATION OF PLATES

PLATE 1.—TEXTURES OF MAGMATIC ORES; PYRITE IN SANDSTONE

Figure

1. Texture approximating xenomorphic (allotriomorphic) granular in chromite ore from Castle Crag mine, Shasta County, California. × 15. Chromite (black) and olivine indent each other and are essentially contemporaneous. Crystal facets are rare, and rotund grain outlines prevail. After Diller.
2. Hypautomorphic granular texture in chromite ore, Dolbear mine, Siskiyou County, California. × 15. Chromite largely euhedral and enclosed in olivine which is partly serpentinized. After Diller.
3. Graphic intergrowth of platinum and osmium in a polished nugget from the Ural Mountains. Etched with Aqua Regia. Magnification not stated but considerable. After Duparc.
4. Layered chromite (dark) in serpentine (after olivine), Campo Formosa, Baia, Brazil. Fountain pen is 12 cm. long. After W. D. Johnston and De Souza.
5. Hand specimen of banded chromite ore from the Akaishi mine, Japan, showing fluxion structure. About half natural size. After Takeo Kato.
6. Pyrite, partly replaced by network of bornite and chalcocite, all forming the matrix of sandstone. Visingso, Sweden. × 111. After O. H. Ödman.



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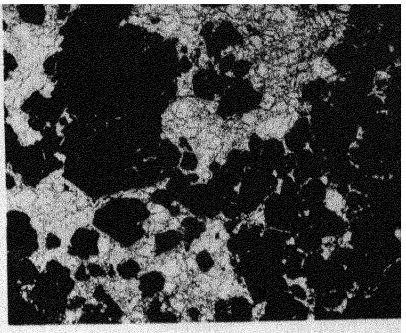

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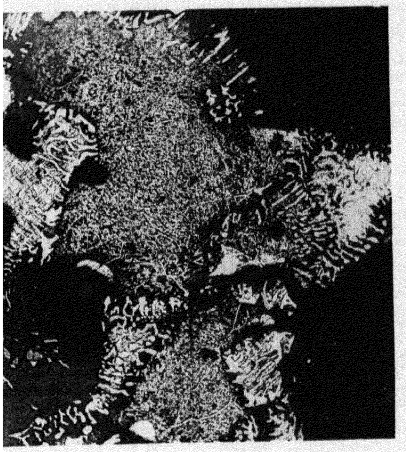

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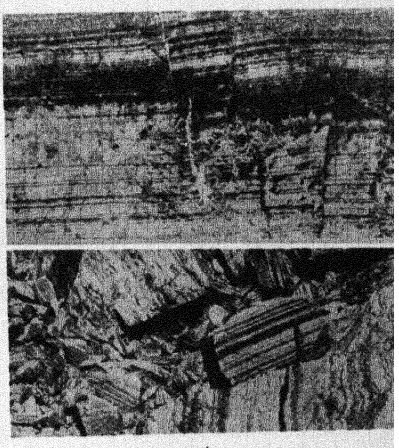

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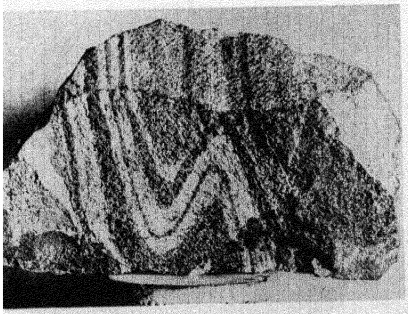

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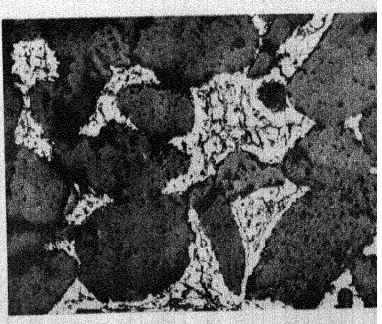

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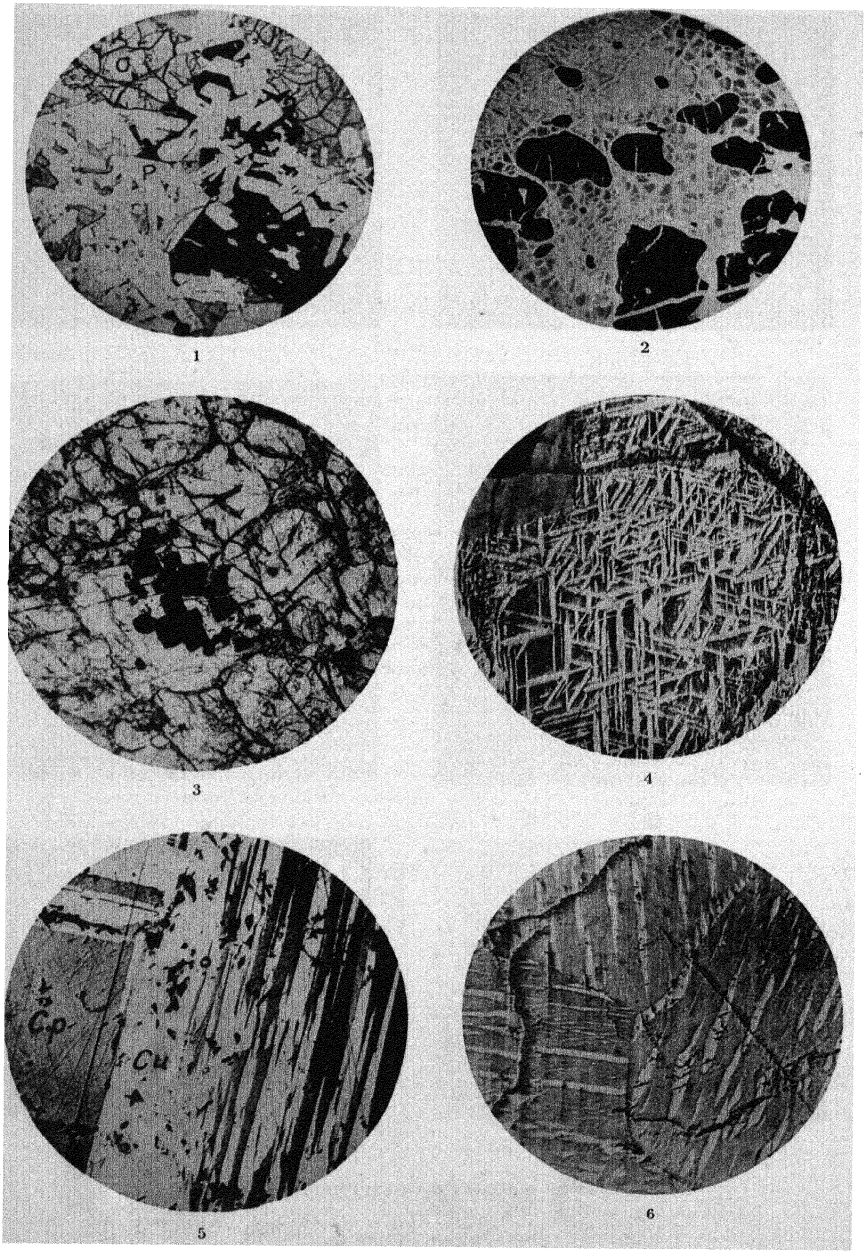
TEXTURES OF MAGMATIC ORES; PYRITE IN SANDSTONE



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TEXTURES OF MAGMATIC ORES

PLATE 2.—TEXTURES OF MAGMATIC ORES

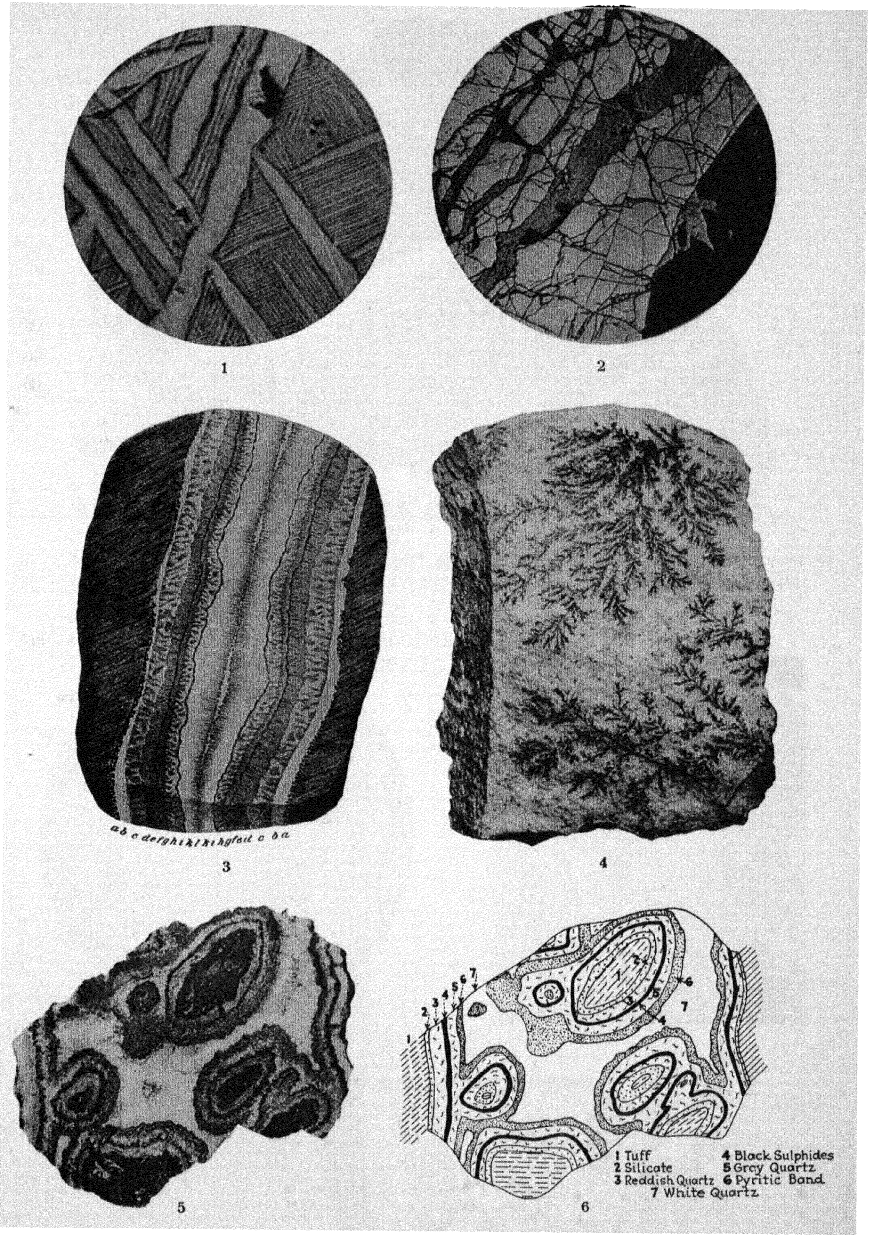
Figure

1. Hypautomorphic granular texture in sulphide diabase from Cook County, Minnesota. The rock consists of fresh olivine (O), augite (A), plagioclase (P), and sulphides (black). Euhedral crystals of feldspar are embedded in the sulphides. After Schwartz.
2. Corrosion of chromite (black) by Olivine that is now completely serpentinized. Feragen, Norway. × 35. After Donath.
3. Synneusis or "swimming-together" texture in chromite (black grains) embedded in olivine (gray). Dunite from Hestmando, Norway. × 19. After J. H. L. Vogt.
4. Unmixing (exsolution) texture in cumberlandite from Cumberland, Rhode Island. Magnetite (dark) and ilmenite (light). Dark spot at upper left is olivine. × 190. After C. H. Warren.
5. Exsolution or unmixing texture in cubanite (Cu) and chalcopyrite (Cp). Republic Mine, Fierro, New Mexico. × 90. After Schwartz.
6. Unmixing or exsolution texture in granular ilmenite rock from St. Urbain, Quebec. Ilmenite dark, hematite light. × 62. After C. H. Warren. The lack of relationship between the hematite bands and the boundaries of the ilmenite grains shows the the banding developed after the ilmenite grains had formed.

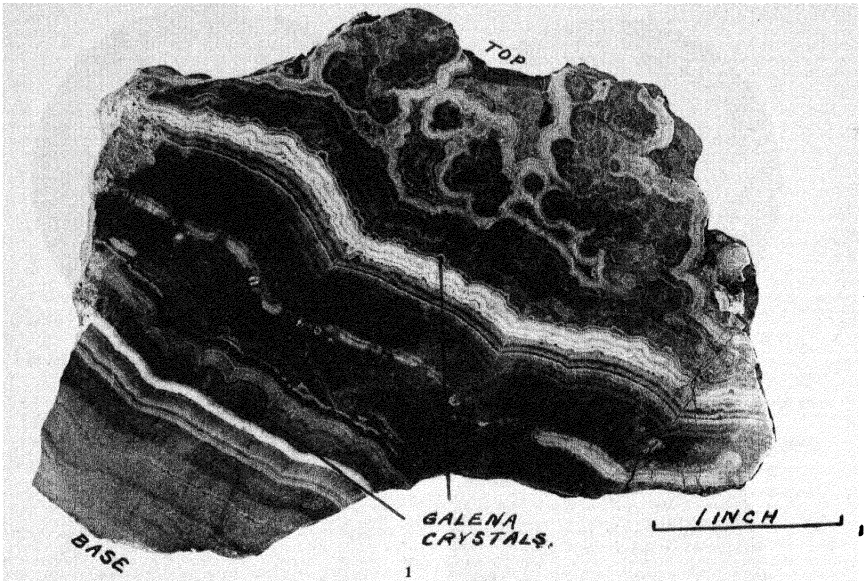
PLATE 3.—TEXTURES DUE TO DEPOSITION IN OPEN SPACES; EXSOLUTION TEXTURE

Figure

1. Unmixing (exsolution) texture in intergrowth of chalcoyprite (light) and bornite. × 560. Note narrowing of chalcopyrite bands near intersections. The texture was induced in ore from Globe, Arizona by artificial heating and quenching. After Schwartz.
2. Fractures in pyrite (light gray) filled with chalcopyrite (dark gray). Ruisenor mine, Cuba. × 64. After Graton. The matched walls of the veinlets indicate fracture filling.
3. Crustification in vein in Churprinz mine near Freiberg, Saxony. From walls toward the center the bands are (a) brown sphalerite, (b) white quartz, (c) green fluorite, (d) brown sphalerite, scattered, (e) barite, (f) marcasite, narrow, (g) barite, (h) fluorite, (i) marcasite, (j) white calcite, (k) wine yellow calcite. The repetitions of barite, fluorite, and marcasite are noteworthy. The vein traverses gneiss. After von Weissenbach.
4. Dendritic deposit of manganese oxide in joint plane in white porphyry. Leadville, Colorado. After Loughlin.
5. Cockade or ring structure in ore from Alacran Mine, Zacual, Mexico. × ½. After Spurr.
6. Diagrammatic sketch identifying the composition of the rings shown Figure 5.



TEXTURES DUE TO DEPOSITION IN OPEN SPACES; EXSOLUTION TEXTURE



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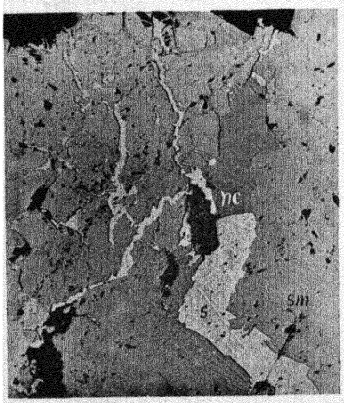

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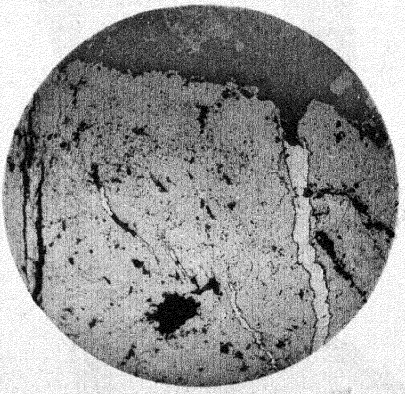

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TEXTURES DUE TO FRACTURE FILLING

PLATE 4.—TEXTURES DUE TO FRACTURE FILLING

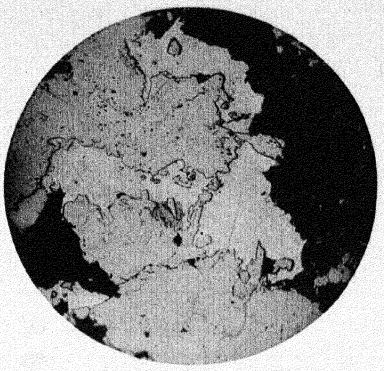
Figure

1. Delicate crustification in finely crystalline zinc ore or "Schalenblende" from Aachen, Germany. The specimen is mostly sphalerite with scattered galena crystals in certain bands. Courtesy of U. S. National Museum.
2. Segmented veinlet of niccolite (nc) and silver (s) in smaltite (sm) in ore from Cobalt, Ontario. Note that in many places automorphic crystals of smaltite form the walls of the veinlet. × 87. After F. N. Guild.
3. Segmented veinlets of native silver (white) and calcite (black) in safflorite (gray). Note the tendency toward matching of the walls in the larger veinlet. Frontier Mine, South Lorraine, Ontario, Canada. × 30. After Bastin.

PLATE 5.—FILLINGS OF RUPTURES AND OF PLANT CELLS; COLLOIDAL TEXTURES

Figure

1. Irregular segmented veinlets of native silver (light gray) and calcite (black) in safflorite (dark gray). Note the automorphic outlines of safflorite against the native silver. Frontier mine, South Lorraine, Ontario. × 40. After Bastin.
2. Central part of the field shown in Figure 1 magnified 150 diameters to show the automorphic crystals of safflorite (darker gray) projecting into silver (light gray). After Bastin.
3. Filling of cells of wood by pyrite (white). Most of the filling took place before any collapse of the cells, but the cells in the upper left corner appear to have been partly rotted and collapsed when the pyrite was deposited. Chalcocite (light gray) occupies the spaces between the pyrite and has largely replaced the cell walls. × 35. Ady Cupb mine, Sacramento District, New Mexico. In Permian beds. After R. P. Fischer.
4. Marcasite in botryoidal masses with spheroidal structure that has replaced sphalerite (gray) and gangue (black). Iron Queen mine, Bigbug District, Arizona. × 80. After Lindgren. The spheroidal banding is probably due to diffusion and not to surface tension.
5. Colloform texture on microscopic scale developed in siliceous gold ore from the Talisman mine, Karangahake, New Zealand. × 35. After S. F. Adams.
6. Same field as in Figure 5 but in polarized light.



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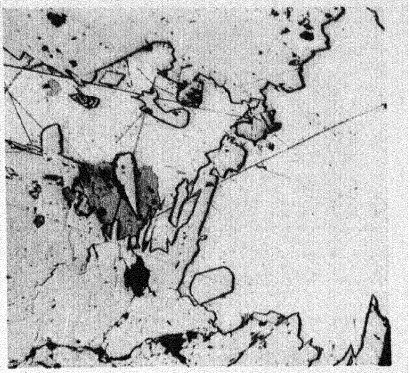

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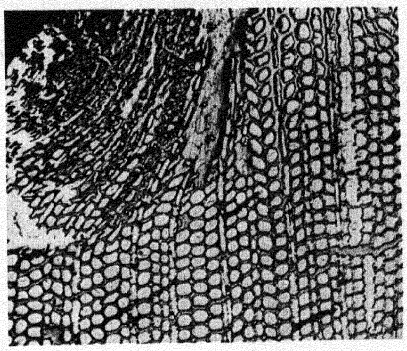

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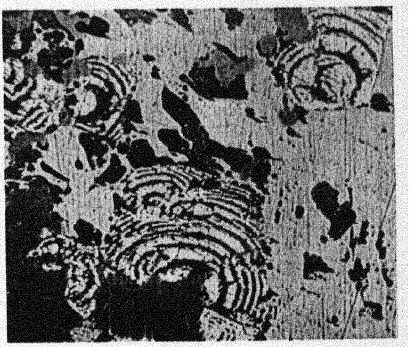

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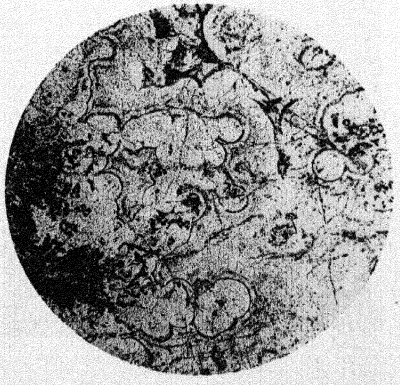

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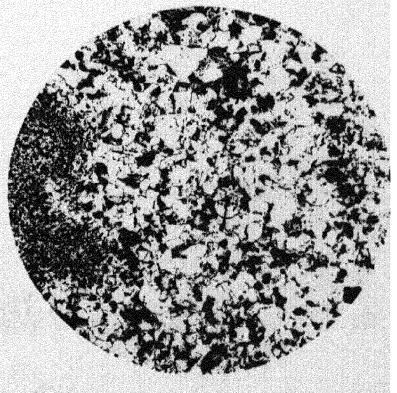

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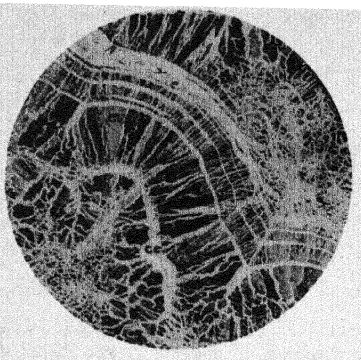
FILLINGS OF RUPTURES AND OF PLANT CELLS; COLLOIDAL TEXTURES



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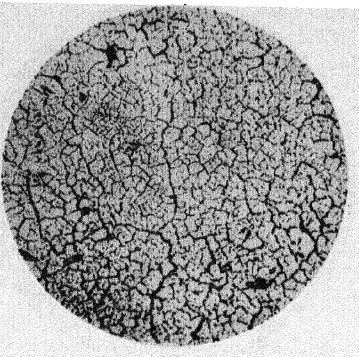

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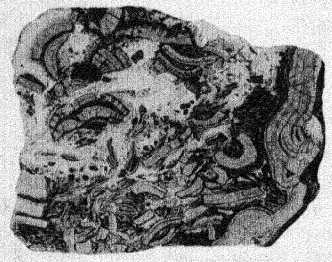

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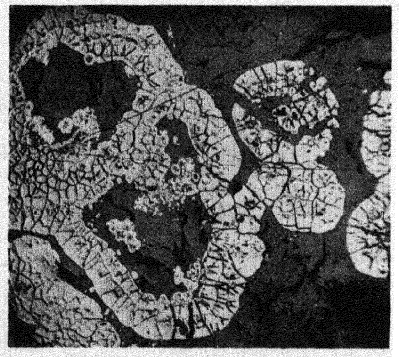

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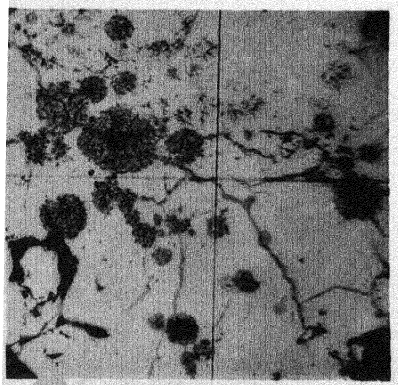

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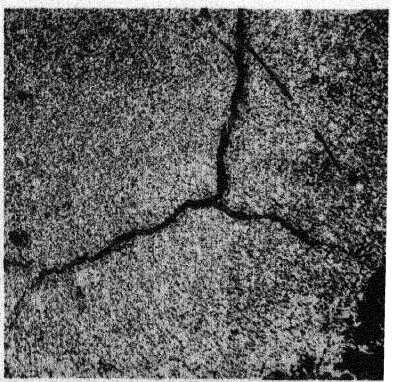

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COLLOIDAL TEXTURES

PLATE 6.—COLLOIDAL TEXTURES

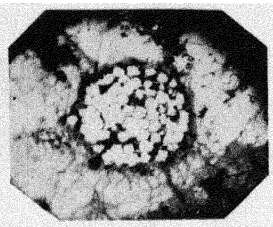
Figure

1. Covellite (cv) and chalcocite (cc) ore from Bonanza mine, Kennecutt, Alaska. The texture is interpreted as metacolloidal on the basis of the spheroidal forms and because the chalcocite in replacing covellite seems to have been guided by concentric and radial and network fractures presumably formed by shrinkage. × 32. After Bateman and McLaughlin.
2. Irregular fracture pattern or "crackled" texture brought out by etching of chalcocite with nitric acid. × 65. Interpreted as due to shrinkage of colloidal chalcocite. Kennecott mine, Alaska. After Bateman and McLaughlin.
3. Brecciated colloform pyrite, Red Cloud mine, Silverton, Colorado. The pyrite fragments lie in a matrix of quartz carrying a little sphalerite and galena. Hand specimen. After F. L. Ransome.
4. Colloform pitchblende in a gangue of barite. × 35. Radial shrinkage cracks are well developed. Joahimsthal, Bohemia. After Schneiderhorn and Ramdohr.
5. Forking shrinkage crack developed in and confined to globular masses of mixed chalcocite and covellite in a matrix of chalcopyrite. × 155. Cornwall mines, Missouri. After G. W. Rust.
6. Bifurcating shrinkage crack developed in globular area of sulphides that has replaced a dolomite nodule in sandstone. The sulphides are aggregates of tiny chalcopyrite spherules, often with pyrite cores, in a bornite matrix. Cornwall mine, Missouri. × 23. After G. W. Rust

PLATE 7.—COLLOIDAL TEXTURES

Figure

1. A single framboidal group of pyrite crystals from the same specimen as Figure 3 but magnified 700 diameters to show the crystalline outlines of the pyrite grains. After Rust.
2. Pyrite crystals disseminated and also in framboidal groups in a matrix of sphalerite. Meggen, Germany. × 480. After Schouten.
3. Framboidal texture in masses of pyrite (p) in the Cornwall mine, Missouri. The pyrite clusters lie in a matrix of chalcopyrite (cp) and are interpreted by Rust as due to the crystallization of a gel globule about many centers. × 122. After Rust.
4. Spheroidal outlines, interference surfaces, and shrinkage cracks in pitchblende (light) from Great Bear Lake, Canada. The dark-gray mineral surrounding the pitchblende is quartz. The shrinkage cracks are filled in part with covellite. × 60. After D. F. Kidd.
5. Brecciated smaltite (white) whose fractures are filled with colloform native arsenic (gray) that has been partially replaced by siderite (dark gray). × 22. After Keil, Freiberg, Germany.
6. Shrinkage cracks developed in successive bands of chalcopyrite deposited on the walls of a fracture in dolomite. Cracks in lower band were developed before upper band was deposited. × 10. Cornwall mine, Missouri. After G. W. Rust.



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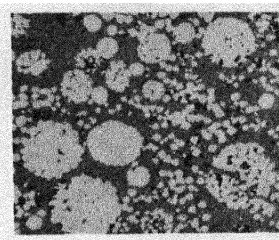

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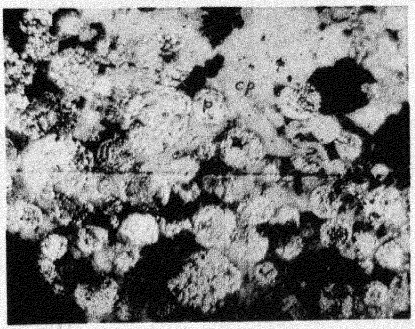

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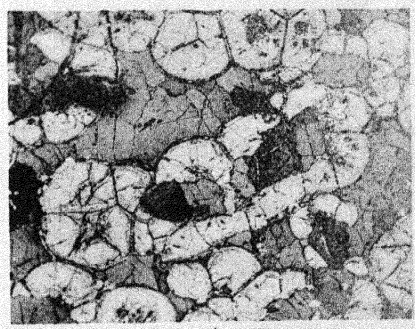

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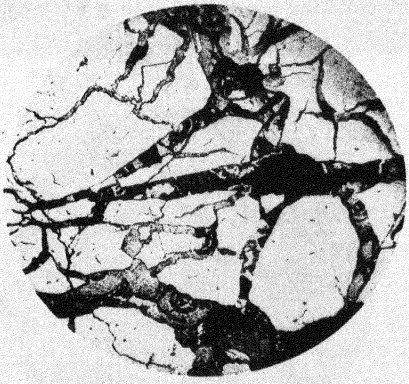

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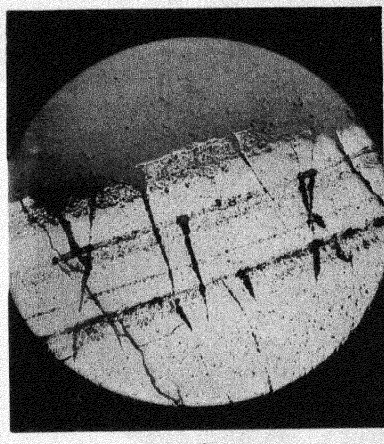

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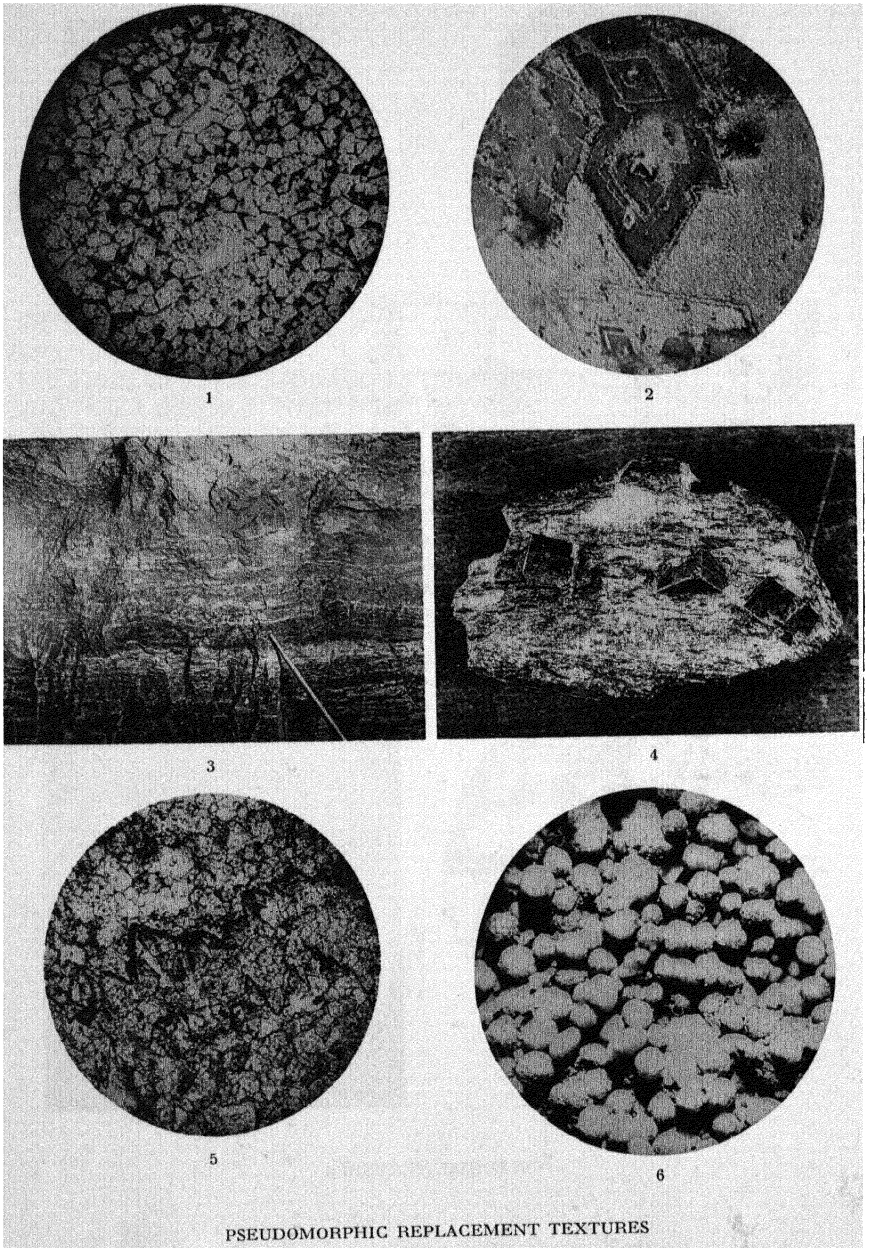
COLLOIDAL TEXTURES



PSEUDOMORPHIC REPLACEMENT TEXTURES

PLATE 8.—PSEUDOMORPHIC REPLACEMENT TEXTURES

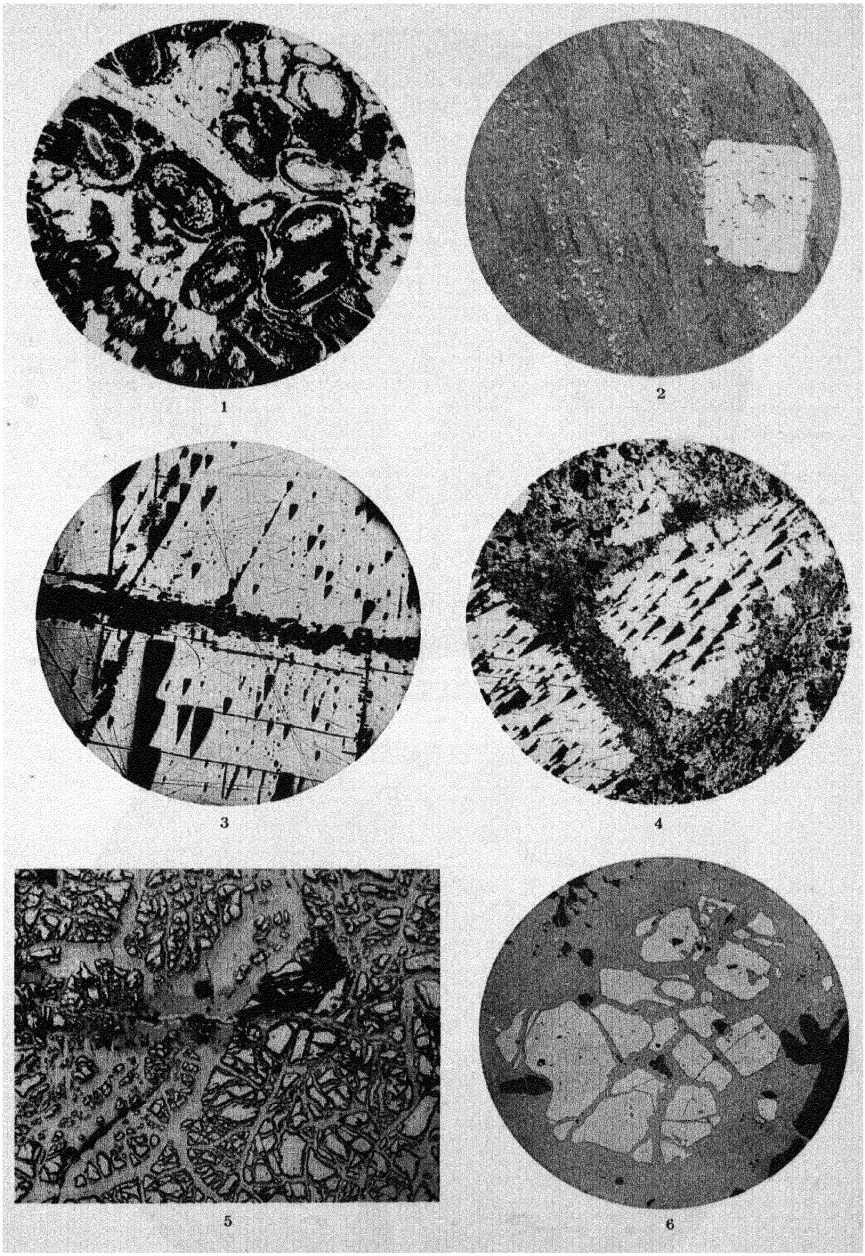
Figure

1. Silicified dolomite, Black Hills, South Dakota. In polarized light the rock is seen to be an aggregate of interlocking quartz grains. In ordinary light the crystal outlines of dolomite grains are clearly recognizable in spite of the complete silicification. After J. D. Irving.
2. Replacement of calcite by iodobromite (pebbly gray) with preservation of the crystal outlines of calcite. × 300. Chanarcillo, Chile. After Whitehead.
3. Preservation of bedding in ore that has replaced limestone, Union mine, near Terry, Black Hills, South Dakota. After J. D. Irving.
4. Transection of schistosity by pyrite cubes. About natural size. After Bastin. The perfect pyrite cubes show no distortion and sharply transect the straight schistosity.
5. Stylolites (jagged black bands) preserved in fluorite. Microphotograph, Spar Mountain, Hardin County, Illinois. After Bastin. These fluorite deposits are replacements of limestones.
6. Oölitic texture in weathered chert nodule in Prairie du Chien limestone near Baraboo, Wisconsin. Except for their siliceous character the oölites in the chert are identical with those in the enclosing limestone and have formed by replacement. × 5. After Bastin.

PLATE 9.—REPLACEMENT TEXTURES

Figure

1. Oölite outlines and concentric structure preserved in auriferous quartz and pyrite that have replaced dolomitic limestone. Siliceous gold-silver ores of the Bald Mountain area, Black Hills, South Dakota. × 22. After Connolly.
2. Transection of schistosity by a cube of pyrite. Hollinger mine, Porcupine district, Ontario, Canada. × 78. After Graton.
3. Replacement veinlets of anglesite developed along cleavage directions of galena (light). Maury mine, Patagonia district, Arizona. × 43. After Bastin.
4. Galena (with triangular pits) partly replaced by anglesite (dark). The parallelism of the triangular pits in the three galena areas indicates that they once formed a single galena area and have been isolated by replacement. Maury mine, Patagonia district, Arizona. × 38. After Bastin.
5. Pyrite (white) partly replaced by chalcocite along irregular fractures. Visingso, Sweden. × 100. After Ödman.
6. Pyrite (light gray) partly replaced by bornite (dark gray). Butte, Montana. × 65. After Graton. Note that none of the pyrite areas are in contact.



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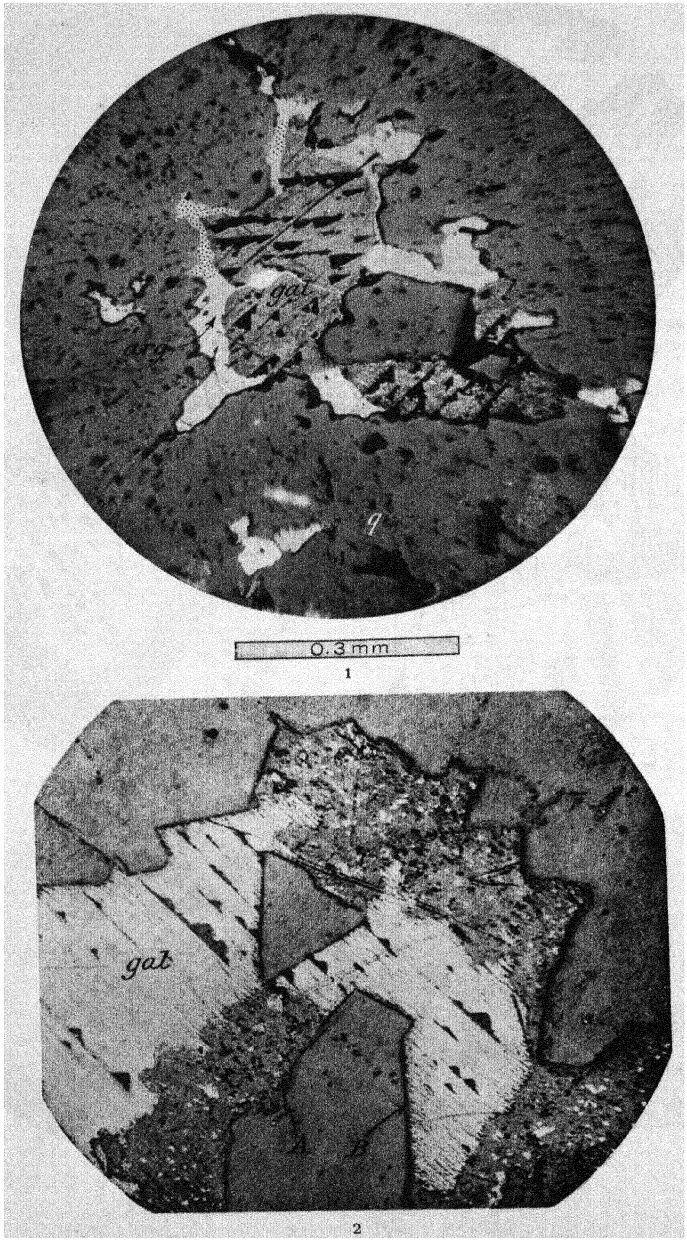
REPLACEMENT TEXTURES



REPLACEMENTS GUIDED BY MINERAL CONTACTS

PLATE 10.—REPLACEMENT GUIDED BY MINERAL CONTACTS

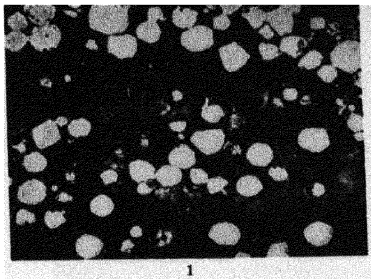
Figure

1. Guided replacement of galena (gal) along its boundaries with quartz (q) and along contacts between galena grains by argentite (arg) and electrum (stippled). The argentite and electrum may both have replaced galena contemporaneously or electrum may be a late local replacement of argentite. Extention mine, Tonopah, Nevada. After Bastin.
2. Partial replacement of galena (gal) by a fine aggregate of a light-colored carbonate, argentite, and some chalcopyrite. There has also been a slight local replacement of the quartz (q) as shown by the ragged boundary of the quartz at A as contrasted with the smooth boundary at B. Tonopah-Belmont mine, Tonopah, Nevada. × 100. After Bastin and Laney.

PLATE 11.—DIFFUSION TEXTURES, MAINLY SILVER ORES

Figure

1. Spherical pellets and crystal grains of pyrite developed by diffuse replacement of shale (black). Mount Isa, Queensland. × 460. After Grondijs and Schouten.
2. Clusters of spheroids of native silver (white) surrounded by rammelsbergite (gray). Both have replaced calcite (black). × 140. Laver mine, Sweden. After Ödman. The very unusual spheroidal form of the silver suggests that it may have been deposited as a colloid.
3. Atoll shapes of pyrite (white, high relief) with galena cores. Outside the atolls are a carbonate (black), sphalerite (gray), and galena (smooth white). All have developed by the replacement of shale. Mount Isa, Queensland. × 54. After Grondijs and Schouten.
4. Tuberclelike growths of arsenopyrite replacing calcite (black). Frontier mine, South Lorraine, Ontario. × 27. After Bastin.
5. Ore from the Townsite mine, Cobalt, Ontario, in which native silver (white) has partly replaced the arsenides of tubercles and the calcite (black) outside the tubercles. × 29. After Bastin.
6. Tuberclelike growths of arsenopyrite (white) and tetrehedrite (light gray) in calcite (dark gray). × 68. After Bastin. Note that tetrehedrite occurs both inside and outside the arsenopyrite tubercles.
7. Tubercle texture in rich silver ore from the Townsite mine, Cobalt, Ontario. ⅔ natural size. After Bastin.



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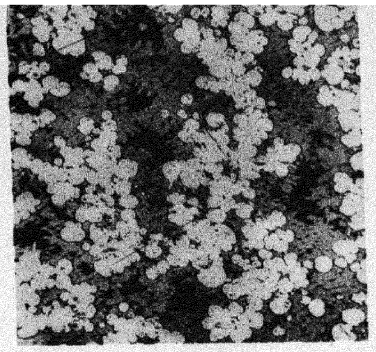

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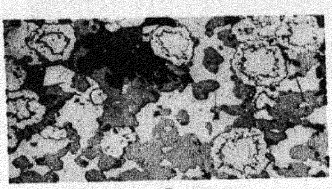

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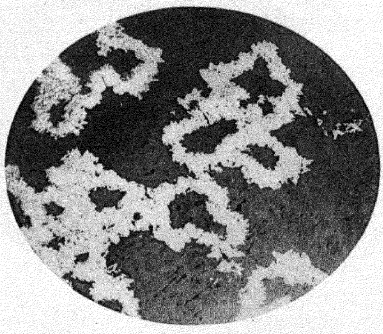

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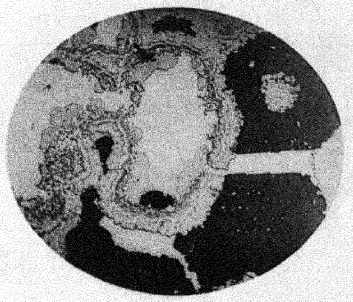

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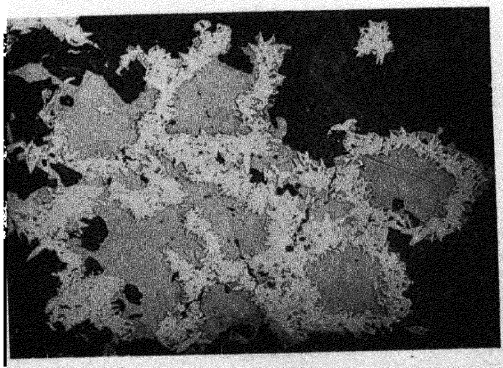

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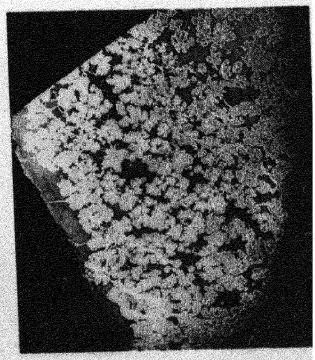

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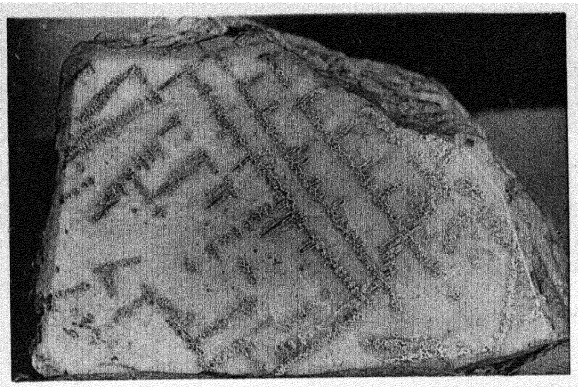
DIFFUSION TEXTURES, MAINLY IN SILVER ORES



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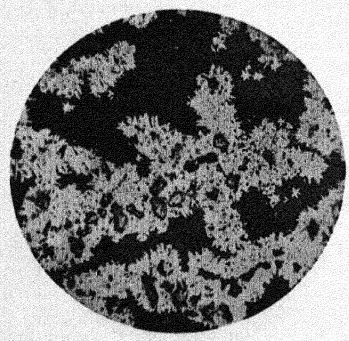

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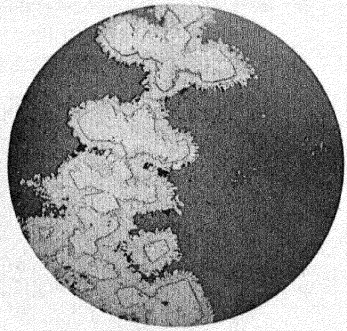

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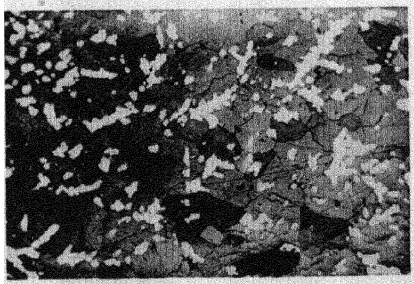

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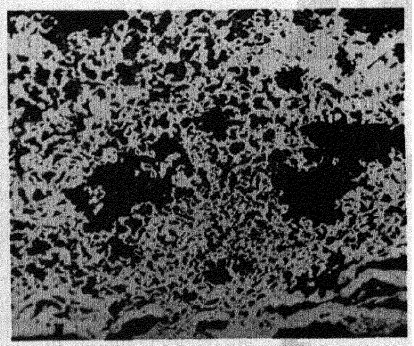

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DIFFUSION TEXTURES IN SILVER ORES

PLATE 12.—DIFFUSION TEXTURES IN SILVER ORES

Figure

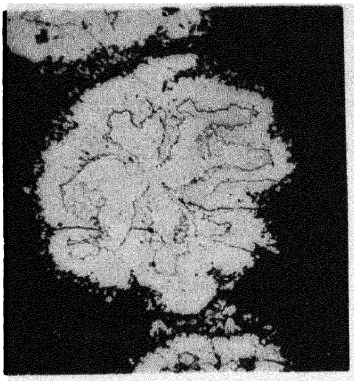
1. Skeleton crystals of native silver enveloped by lollingite in a matrix of calcite. × 1½. Casey mine, Cobalt, Ontario. After Bastin.
2. Native bismuth as crystallization centers for safflorite rosettes. These lie in a quartz matrix. Isolated in the quartz can be seen four small star-shaped triplets of safflorite. Annaberg, Saxony. × 58. After K. Keil.
3. Skeleton crystals of silver (white) enveloped in lollingite (light gray) in a calcite matrix (dark gray). × 25. Same specimen as Figure 1. Casey mine, Cobalt, Ontario. After Bastin.
4. Dentritic crystals of silver in calcite in cobalt-nickel-silver ore from Laver mine, Sweden. × 23. After Ödman. The grain boundaries of the calcite can be dimly seen, and the silver bears no relation either to these or to calcite cleavage.
5. Diffuse replacement of calcite by native silver (white). Laver mine, Sweden. × 77. After Ödman.

PLATE 13.—REPLACEMENT OF SILVER DENDRITES BY CALCITE

The six figures form a sequence of cross sections of silver dendrites enveloped by lollingite and lying in a matrix of calcite. The dendrites range from 1 to 3 mm. across. Miller Lake-O'Brien mine, Gowganda, Ontario, Canada. After Bastin.

Figure

1. Silver core (white) of the dendrite forms a five-armed cross enveloped in lollingite (gray) Two small veinlets of calcite (black) traverse the lollingite and enlarge slightly where they encounter the silver.
2. Dendrite showing a four-armed silver core in which the tips of two of the arms have been replaced by calcite (black).
3. Dendrite in which several calcite veinlets traverse the lollingite and enlarge abruptly where they reach and replace the silver.
4. Another dendrite showing that it is usually the outer portions of the silver core that are the first to be replaced by calcite.
5. Another dendrite in which portions of the five-armed silver core have been replaced by calcite. These are the portions adjacent to two calcite-filled fractures which probably served as feeders for the replacing solutions.
6. In this section no silver remains within the dendrite. The core is now wholly calcite but the cross-shaped form of the original silver core which it replaced is perfectly preserved.



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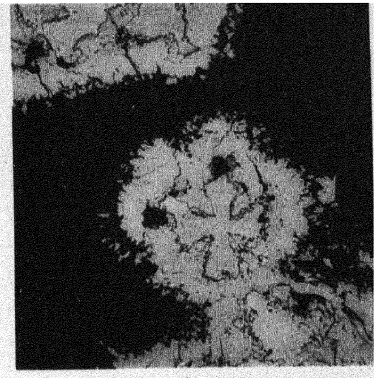

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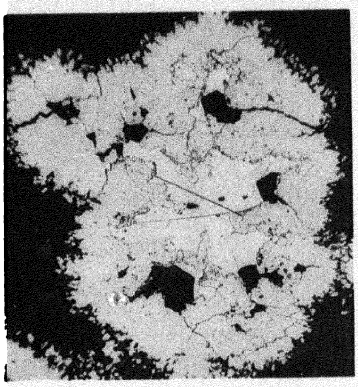

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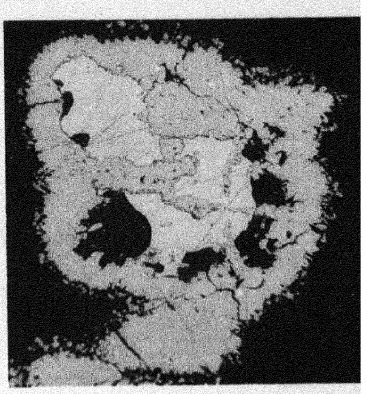

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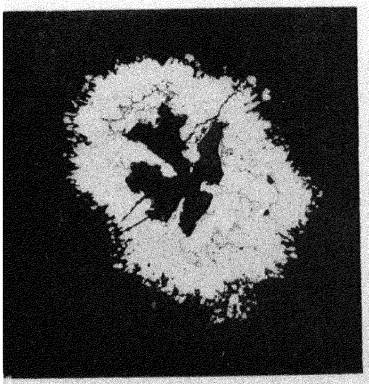

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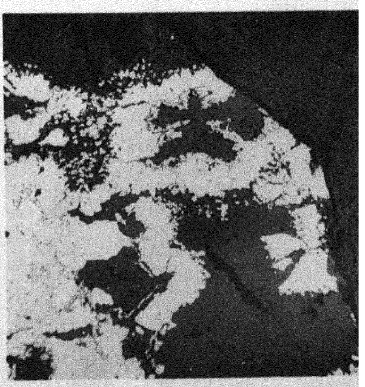

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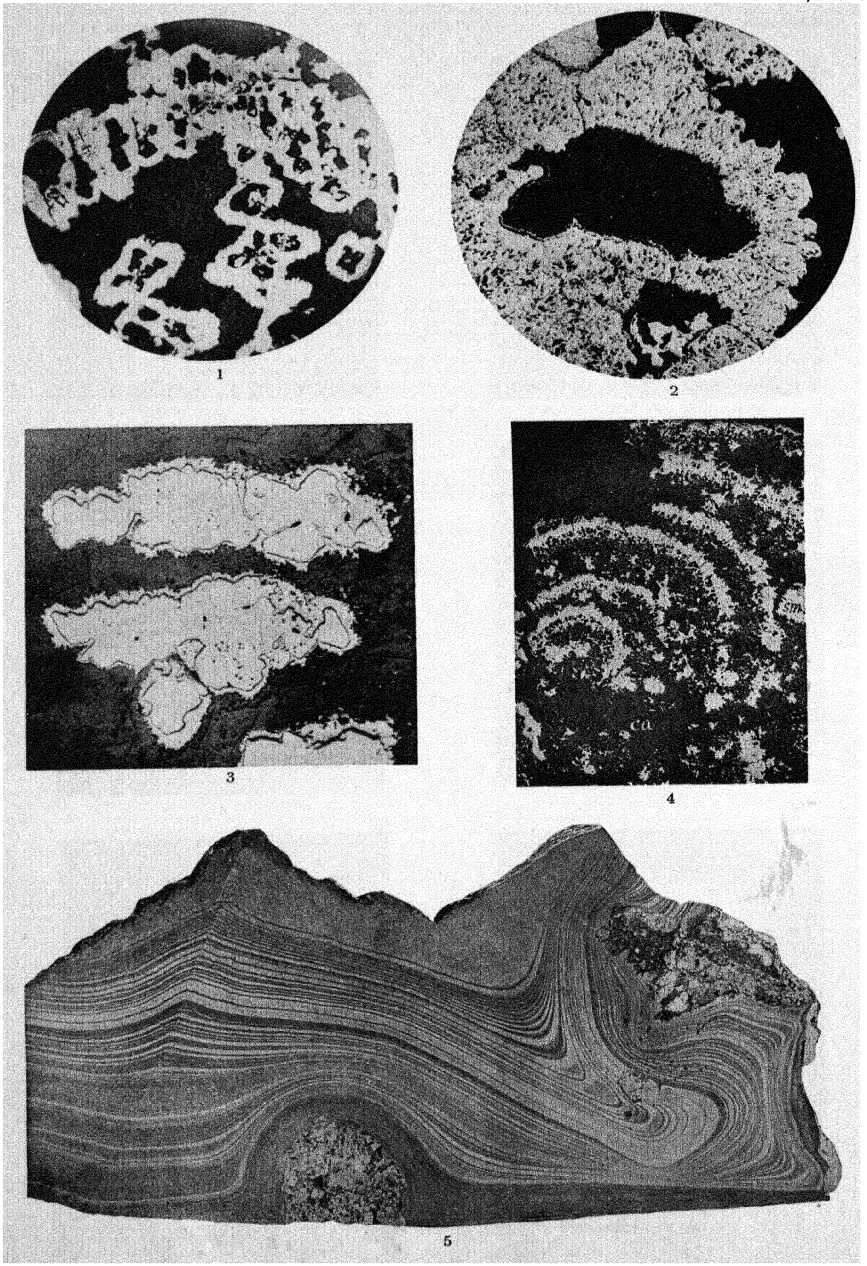
REPLACEMENT OF SILVER DENDRITES BY CALCITE



REPLACEMENT OF SILVER DENDRITES; DIFFUSION BANDING

PLATE 14.—REPLACEMENT OF SILVER DENDRITES; DIFFUSION BANDING

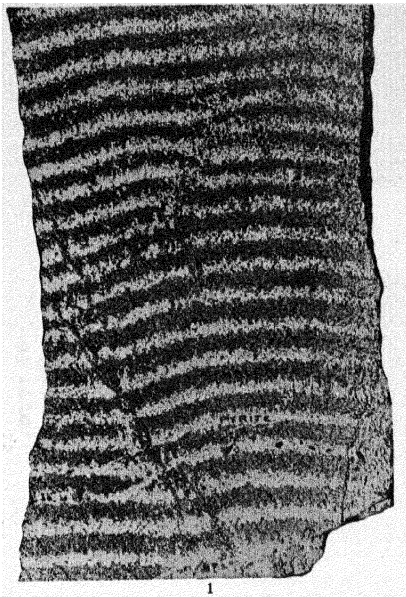
Figure

1. Silver dendrites (white) enveloped by rammelsbergite-safflorite all in a calcite matrix (black). The silver of the original dendrites has been partly replaced by calcite, usually peripherally. × 50. Marienberg, Erzgebirge, Germany. After Keil.
2. Skeleton crystals originally of silver (white) now largely replaced by calcite (black) are enveloped in safflorite and cobaltite. In parts of the larger dendrite a narrow rim of silver remains; in the smaller one an irregular remnant of silver remains. × 100. Cobalt, Ontario. After Keil.
3. Skeleton crystals of silver (white) enveloped by rammelsbergite (crusts) in a matrix of calcite. × 37. Laver mine, Sweden. After Ödman. The outlines of the silver are mostly automorphic but in places are very ragged due to replacement by rammelsbergite.
4. Rhythmic deposition of smaltite in calcite. Cobalt, Ontario. × 8. After Guild.
5. Diffusion banding due to oxidation in fine, even-grained felsite. Tonopah, Nevada. Nearly natural size. After Bastin and Laney.

PLATE 15.—DIFFUSION BANDING; GRAPHIC REPLACEMENT TEXTURES; SYNTHETIC REPLACEMENTS

Figure

1. Diffusion banding developed in limestone in South Moyer mine, Leadville, Colorado. Light bands are pyrite, and dark bands are sphalerite. × 2. After Loughlin.
2. Diffusion banding in jasperoid from Gemini mine, Tintic, Utah. Light bands are chalcedony and quartz. Dark bands show grains of galena, sphalerite, and some pyrite. Ordinary light. × 15. After Lindgren.
3. Graphic association of chalcocite (cc) and bornite (b). The chalcocite of the intergrowth is continuous with chalcocite that has replaced bornite along quartz contacts. Engel's mine, Plumas County, California. After A. F. Rogers.
4. Graphic association of galena (white) with tennantite and pearceite (both gray). Barite blades (black) have not been replaced by galena but by tennantite and pearceite. It follows that the latter must also have replaced the galena. Tintic, Utah. × 480. After Lindgren.
5. Synthetic replacement of pyromorphite (gray) by galena (white) along a multitude of small fractures which are unrecognizable even under high powers of the microscope in the unaltered pyromorphite. × 130. After Schouten.
6. Synthetic replacement of pyromorphite by galena (white) in which relatively large automorphic crystals of pyromorphite have been resistant to replacement whereas interstitial fine-grained pyromorphite has been completely replaced by galena. × 80. After Schouten.
7. Synthetic replacement of phosgenite (dark) by galena (light). The replacement seems to have been of the diffuse type uninfluenced by the well-defined cleavage of the phosgenite. × 33. After Schouten.



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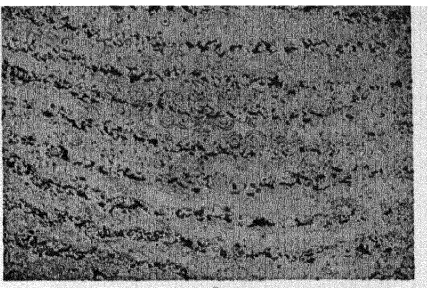

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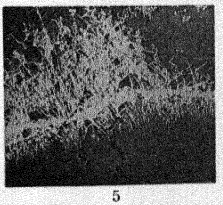

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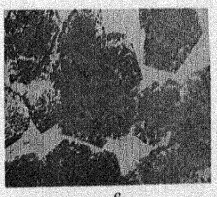

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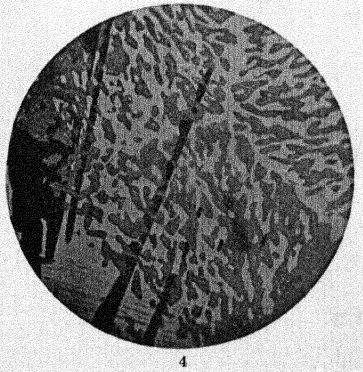

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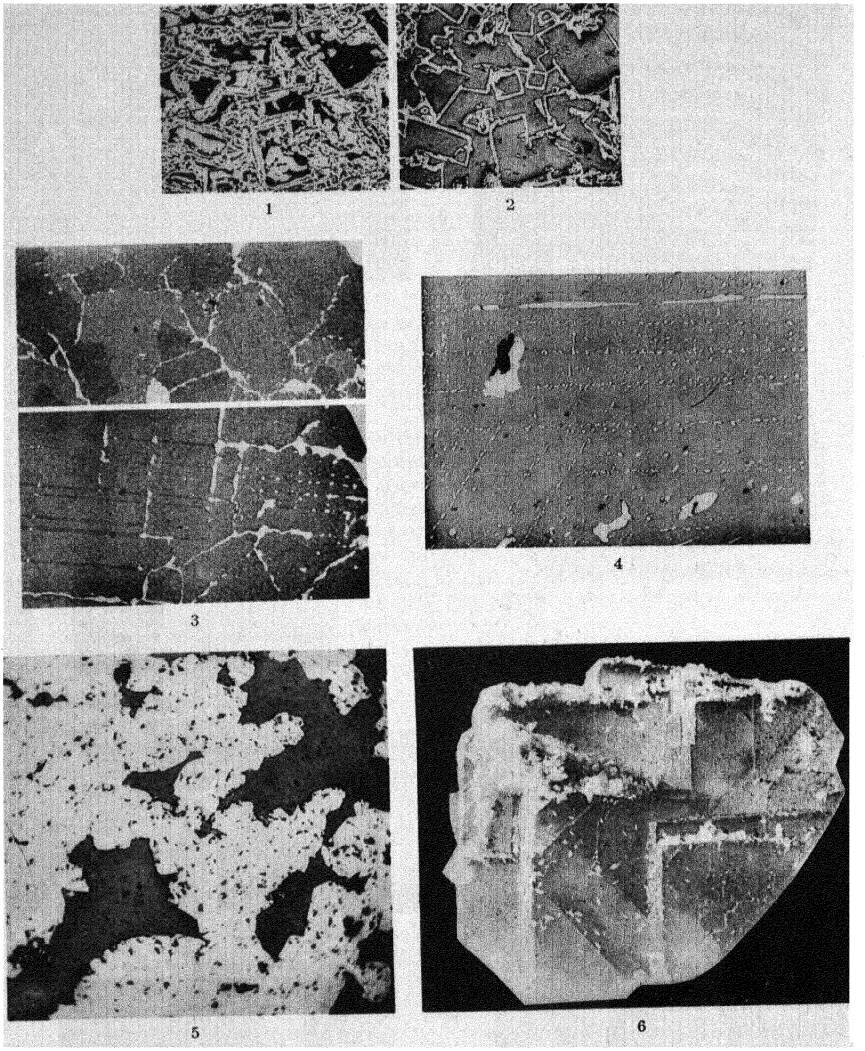
DIFFUSION BANDING; GRAPHIC REPLACEMENT TEXTURES; SYNTHETIC REPLACEMENTS



SYNTHETIC REPLACEMENTS; INCLUSIONS FORMED BY REPLACEMENT

PLATE 16.—SYNTHETIC REPLACEMENTS; INCLUSIONS FORMED BY REPLACEMENT

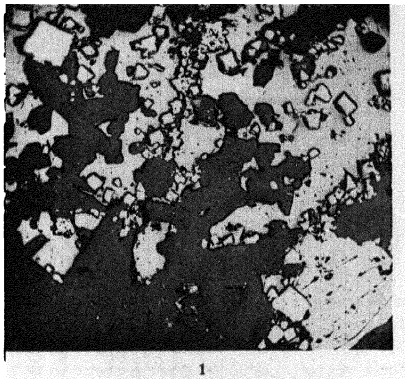
Figure

1. Complete synthetic replacement of cuprite by metallic copper (white) with preservation of cuprite structure and with loss of volume as shown by the numerous vugs (black) and porosity. × 39. After Schouten.
2. Same specimen as Figure 1 but in the initial stages of replacement. White is copper, and gray is cuprite. × 39. After Schouten.
3. Dotlike inclusions of galena (white) at intersections of cleavage and twinning planes in calcite (dark). Mount Isa, Australia. × 58. After Grondijs and Schouten. Note transitions from dotlike inclusions to replacement veinlets.
4. Inclusions of chalcopyrite in sphalerite distributed along twinning directions and cleavage. × 72. After Van der Veen.
5. Quartz (dark) conforming to the outlines of older sphalerite (light) which in part are straight crystal faces and in part rotund as is so common in sphalerite crystals. Carmen mine, Zacatecas, Mexico. × 25. After Bastin.
6. Calcite crystals (white) deposited on fluorite in Hillside vein, Rosiclare, Illinois. The calcite has been deposited like snow on a roof mainly on upward-facing surfaces of the fluorite indicating a gravitational control in its deposition. × $\frac{2}{3}$. After Bastin.

PLATE 17.—TEXTURES OF OÖLITIC IRON ORES; PYRITE; IRON BACTERIA

Figure

1. Cubical pyrite (white) showing automorphic outlines against quartz (dark) and chalcopyrite (gray). The quartz in turn shows automorphic hexagonal outlines against the chalcopyrite. In the absence of any evidences of replacement, the order of deposition is interpreted as pyrite, quartz, and chalcopyrite. San Jesus mine, Zacatecas, Mexico. × 28. After Bastin.
2. Supposed remains of Iron Bacteria from chert in Pokegama formation conformably underlying the Biwabik iron-bearing formation of the Mesabi district, Minnesota. × 1540. After Gruner.
3. Oölitic iron ore showing closely packed oölites of chlorite and hematite with concentric structure (A). A highly hematitic and nearly opaque oölite is shown at C. Nuclei of siderite are shown at B. The sparce matrix is mainly chlorite with some siderite. × 23. La Ferriere aux Etangs, Orne, France. After Cayeux.
4. Iron ore from Thoste,France, composed of false oölites in a matrix (E) of calcite. A to D are crinoid plates partly replaced by hematite. F = false oölites in which calcite has been extensively replaced by hematite. G = hematite oölites. × 32. After Cayeux.
5. Iron ore from Mazenay, France, showing both false oölites composed of crinoid remains (A) and oölites (Bto E) in a matrix largely of granular calcite but with some organic remains. × 32. After Cayeux.
6. Iron ore from Mexy, France, with oölites of hematite (black) in a matrix mainly siderite. At A are quartz nuclei. Around most of the oölites a narrow band of chlorite (B) has developed by the replacement of the siderite matrix. In places, as at C, only ragged replacement remnants of siderite remain. Where hematite oölites are in contact no chlorite band is present. × 29. After Cayeux.



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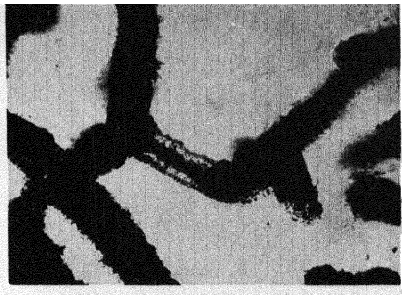

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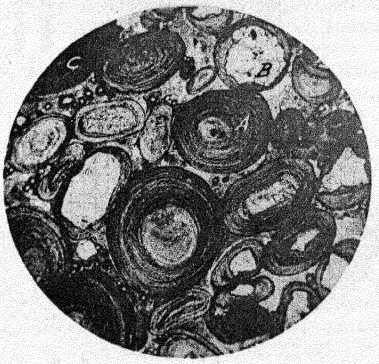

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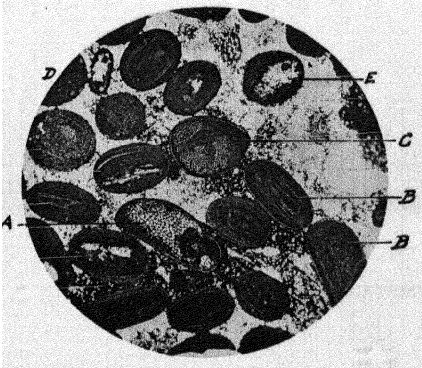

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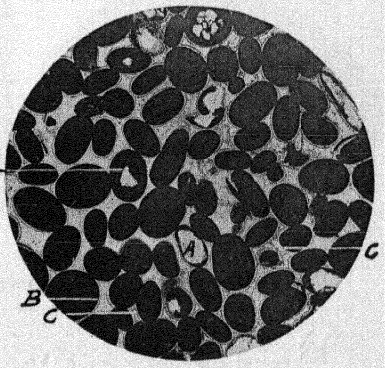

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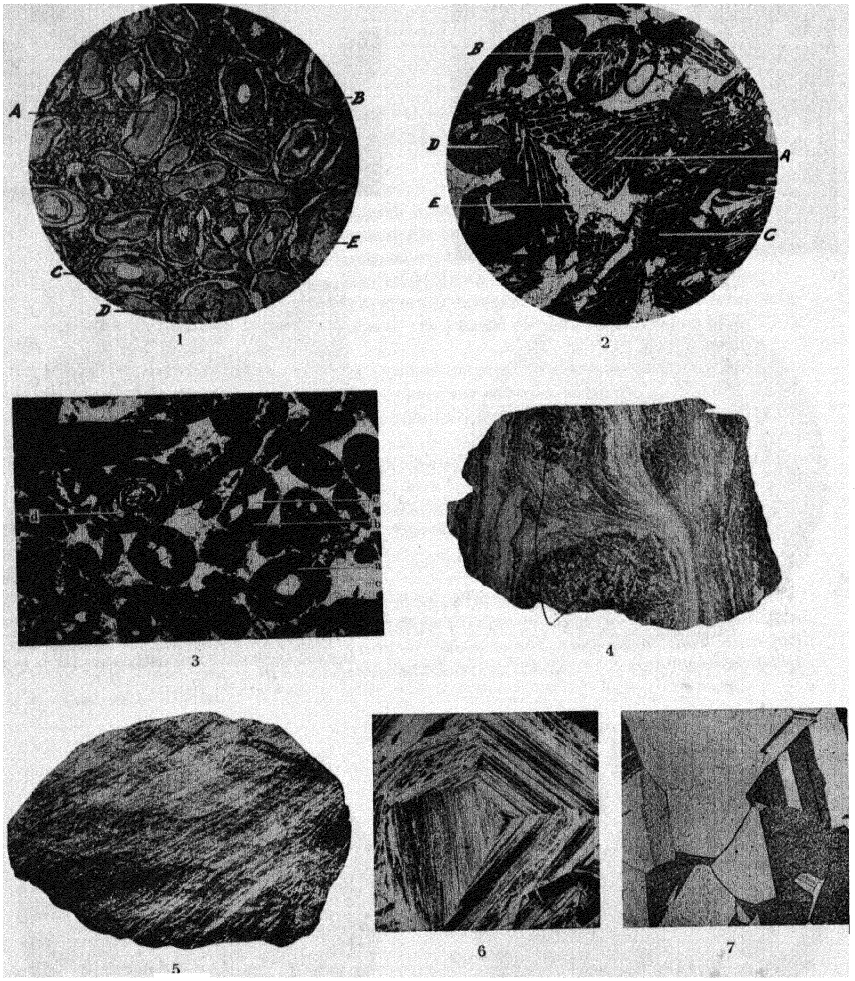
TEXTURES OF OÖLITIC IRON ORES; PYRITE; IRON BACTERIA



OÖLITIC IRON ORES; DYNAMOMETAMORPHIC TEXTURES

PLATE 18.—OÖLITIC IRON ORES; DYNAMOMETAMORPHIC TEXTURES

Figure

1. Silurian oölitic iron ore showing peripheral replacement of chloritic oölites (A) to various depths by white rims of siderite. The smooth outer border of the siderite marks the original border of the chlorite oölite. The inner border of the siderite is irregular and in many places transects the concentric banding of the chlorite. Nuclei of quartz are shown at B and C, and nuclei of siderite at D and E. The matrix consists of siderite rhombs embedded in chlorite and hematite. × 23. La Ferriere aux Etangs, Orne, France. After Cayeux.
2. Clinton oölitic iron ore from Wolcott, New York, composed of rounded and semiangular fossil remains partly replaced by hematite. These lie in a matrix of calcite (E). B = oölite composed of a nucleus of a rounded bryozoan fragment largely hematized and enveloped in a cortex of hematite. C = bryozoan remains almost completely filled and replaced by hematite. D = crinoid remains partly filled and replaced by hematite. × 14. After Cayeux.
3. Oölites of pyrite (black) in a matrix of crystalline quartz (c), Wabana, Newfoundland. a = pyrite oölite, b = pyritized shell fragment partly replaced by quartz (e), d = quartz in oölite. × 56. After Hayes.
4. Dynamometamorphosed ore from Rammelsberg, Germany, showing flowage of finely granulated sulphides around nodules of relatively coarse barite and sphalerite. Natural size. After Lindgren and Irving.
5. Hand specimen of "steel" galena from the Slocan district, British Columbia showing characteristic banded structure produced by crushing under differential pressure. × 1½. After Uglow.
6. Zonal texture brought out by etching in native silver from the Elkhorn mine, Montana. × 100. After Carpenter and Fisher.
7. Homogeneous recrystallized texture brought out by etching in native silver (of Precambrian age) from Cobalt, Ontario, Canada. × 100. After Carpenter and Fisher.

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