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PETROGRAPHY AND ORIGIN OF ILLINOIS NODULAR CHERTS

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ABSTRACT

Seventy-eight samples of nodular chert from 18 Illinois limestone and dolomite formations, ranging from Cambrian through Mississippian age, were investigated to determine the petrography and mode of origin of the nodules. Regardless of geologic age or type of host rock, the nodules were similar in mode of occurrence and in principal textural characteristics.

The cherts are dominantly microcrystalline or cryptocrystalline quartz with a lesser amount of fibrous quartz. No opal or hydrated silica was detected. Almost all the cherts contain residual masses of their host rock. Field relationships and a variety of evidence for replacement leads to the conclusion that the cherts are epigenetic concretions formed by metasomatic processes operating during diagenesis and involving the aggregation of silica that originally had been deposited syngenetically with, and dispersed through, the host rocks.

INTRODUCTION

Chert is found in many limestones that crop out in Illinois and range in age from Cambrian to Mississippian. The chert may appear as nodules, lenses, or beds, and some Devonian rocks in extreme southern Illinois are entirely chert.

The investigation reported here was limited to a study of the nodular cherts and deals with the petrography and mode of origin of the nodules.

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SAMPLES

Chert samples were obtained from 35 outcrops whose geological identity and number of outcrops sampled are given below:

Mississippian System: Kinkaid limestone (1); Vienna limestone (1); Ste. Genevieve limestone (2); St. Louis limestone (4); Salem limestone (3); Burlington limestone (4).

Devonian System: Clear Creek chert (1); Backbone limestone (1); Bailey limestone (2).

Silurian System: Racine dolomite (1); Kankakee dolomite (2); Sexton Creek limestone (2); Girardeau limestone (1).

Ordovician System: Galena dolomite through Platteville dolomite (1); Shakopee dolomite (2); Oneota dolomite (2).

Cambrian System: Trempealeau dolomite (1).

The outcrops were sampled by taking several nodules of chert from each conspicuous zone or type of chert present. The samples were marked so that thin sections could be cut normal to the bedding.

OCCURRENCE AND COLOR OF CHERT NODULES

Chert nodules in Illinois limestones and dolomites generally are less than three feet in diameter and are commonly more or less flattened parallel to their horizontal axes and to the bedding of the enclosing rock. Nodules of almost perfect spherical shape have been observed but are rare. All nodules have a megascopically smooth, though commonly irregular, outline. Knobs and other types of protuberances are common. The contact of the nodules with the enclosing rock is in all cases smoothly curving. Some nodules have a knife-sharp contact, in others the contact is diffuse or transitional.

Oolites, bedding, and other structures that are common in the surrounding limestone or dolomite are observed in many chert nodules. In some places in extreme southern Illinois, nodules contain small crystals and thin veins of fluorspar that cross linear structures in the chert without offset and terminate at the boundary of the nodules.

The nodules in Illinois limestones and dolomites do not appear to be associated with a solution channel network or with specific bedding planes. The chert is randomly distributed at most exposures. The nodules are restricted to certain zones in some formations, but within the zones they do not occur at any particular horizon. Where nodules are concentrated in a particular zone they are commonly smaller than at nearby localities where they are farther apart.

The cherts sampled were black, brown, gray, blue, and white (discounting surface staining due to weathering). Chert of the Kinkaid, Ste. Genevieve, and St. Louis limestones generally was dark gray to black; light brown, light gray, and white cherts predominate in the Salem, Burlington, Clear Creek, Backbone, Bailey, Racine, Kankakee, Sexton Creek, Galena, Platteville, Shakopee, Oneota, and Trempealeau formations. Dark brown chert was found only in the Girardeau and Vienna limestones, and one sample from the Vienna formation was light blue.

METHODS OF LABORATORY STUDY

All chert samples were studied by means of thin sections. Seventeen samples were further studied by infrared absorption and eleven by x-ray diffraction to check on the accuracy of thin-section determinations and to investigate the possible occurrence of opal and amorphous silica.

All samples investigated by infrared absorption contain quartz, eleven calcite, and twelve dolomite. Six samples contain both calcite and dolomite; of the remaining eleven, six cherts contain dolomite as the only carbonate whereas five have calcite but no dolomite.

Quartz is present in all eleven cherts examined by x-ray diffraction. Calcite is present in ten of the cherts, whereas dolomite is present in eight. The only chert that has no calcite contains dolomite.

X-ray and infrared absorption data for the cherts examined by both methods are in agreement. All the cherts contain carbonate minerals. Quartz is the most abundant and essential mineral in the nodules. None of the cherts examined contain opal or amorphous silica.

PETROGRAPHY

The character of the cherts as shown in thin sections is indicated in table 1 and summarized in table 2. No consistent differences related to geologic age are evident. The quartz in cherts from dolomites is somewhat finer grained than that in limestone cherts, but not distinctively so.

More fossils and coquina or coquinoid material is found in cherts from limestones than in dolomite cherts, which is to be expected because Illinois limestones contain more fossils and fossil materials than do Illinois dolomites. Because of the general similarity of the cherts, their character and origin is discussed without reference to formation or kind of host rock.

Quartz

Thin-section examination of the cherts shows that their essential and most abundant mineral is quartz. The quartz occurs in three varieties, microcrystalline (crystals less than 0.1 but greater than 0.01 mm. in diameter), cryptocrystalline (less than 0.01 but greater than 0.002 mm. in diameter), and fibrous quartz, sometimes called chalcedony. Some specimens show quartz euhedra. The length of fibers in the fibrous quartz is highly variable, depending upon the width of separation of centers of crystallization.

The terms microcrystalline and cryptocrystalline quartz have been here adopted for convenience in textural classification of the cherts. The two types, which differ only in grain size, make up more than 95 percent of the silica content of the chert.

Keller (1941), using the term microcrystalline quartz to include the material for which both terms mentioned above have been used, showed that the fine-grained silica has the optical and x-ray properties of quartz, except that the extinction is undulatory. He attributed the undulatory extinction to straining of the quartz developed in transition from colloidal silica to microcrystalline quartz. There seems little evidence that the quartz crystals are greatly strained. It is known that if crystals are smaller than the thickness of the section, undulatory extinction results from the passage of light through several misorientated crystals.

It appears that such small quartz crystals grow by replacing carbonate minerals. Individual crystals of carbonate are found that have microcrystalline and cryptocrystalline quartz traversing them, outlining crystal rims, or growing in irregular masses within them.

The microcrystalline quartz commonly is clouded by opaque dust. This dust may be iron, carbonaceous matter, or other impurities that could be accepted and hidden in the structure of the carbonate minerals but that finds no place in quartz and so forms an oxide or other phase, either crystalline or amorphous, that is less subject to replacement by the quartz. Some darkening of the quartz is due to finely disseminated carbonate material that has not been

Table 1. - Texture of Chert

Formation and kind of rock ls.= limestone dol.= dolomite	Texture of chert						Evidences of replacement			
							Pseudomorphs of quartz after			
							Crystals		Primary rock structures	
	No. samples	Coquina or coquinoid	Cryptocrystalline mosaic	Microcrystalline mosaic	Oolitic	Siliceous *	Unit	Aggre-gates	Oolites	Fossils
Kinkaid ls.	1	-	-	1	-	-	-	1	-	1
Vienna ls.	3	1	1	1	-	-	-	3	-	3
Ste. Genevieve oolite	13	4	3	5	5	-	1	13	5	12
St. Louis ls.	10	1	1	7	-	1	1	10	-	9
Warsaw-Salem ls.	4	2	2	-	-	-	-	4	-	4
Burlington ls.	13	7	2	4	-	-	-	12	-	12
Clear Creek ls.	2	-	-	2	-	-	2	2	-	1
Backbone ls.	3	-	1	1	-	1	1	3	-	3
Bailey ls.	3	-	1	2	-	-	2	3	-	1
Racine dol.	3	-	3	-	-	-	1	3	-	3
Kankakee dol.	3	-	1	1	-	1	1	3	-	1
Sexton Creek ls.	2	-	-	2	-	-	-	2	-	2
Girardeau ls.	1	-	1	-	-	-	-	1	-	-
Galena dol.	5	-	4	1	-	-	-	5	-	4
Platteville dol.	1	-	1	-	-	-	-	1	-	1
Shakopee dol.	6	-	-	-	1	4	5	6	4	-
Oneota dol.	4	-	1	1	-	2	-	3	-	1
Trempealeau dol.	1	-	-	-	-	1	-	1	-	-

* Not strictly textural but included here for convenience.

† The numbers indicate the number of specimens showing a given phenomenon.

and Evidences of Replacement†

Evidences of replacement				
Embayed crystals or masses of host rock in chert	Residual islands of host rock in chert	Islands of quartz in carbonate host rock	Growth of new carbonate crystals	Miscellaneous
1	-	1	1	Relict bedding - 1
3	3	-	3	
11	10	9	8	
8	5	4	5	Relict bedding - 1
1	-	-	3	Relict limestone texture - 1
11	8	7	3	Pelitic - 2
2	1	1	1	
3	2	2	1	
3	2	2	2	Much carbonate - 2; relict bedding - 1
2	-	2	-	Much dolomite - 1
3	3	2	-	Relict bedding - 2
2	1	-	1	Relict bedding - 1
1	-	-	1	
3	3	3	-	Large masses of dolomite - 2
-	-	-	1	Large masses of dolomite - 1
4	2	5	2	Detrital quartz - 4
1	1	1	-	Relict bedding - 1
1	1	1	-	

Table 2. - Summary of Data in Table 1 by Host Rock Types

	Frequency (%)	
	Limestone	Dolomite
Texture of chert		
Coquina or coquinoid	27	0
Cryptocrystalline mosaic	22	43
Microcrystalline mosaic	45	13
Oolitic	9	4
Evidence of replacement		
Pseudomorphs of quartz after:		
Unit crystals	13	13
Aggregates of crystals	98	96
Oolites	9	13
Fossils	69	43
Embayed crystals or masses of host rock in chert	84	61
Residual islands of host rock in chert	58	43
Islands of quartz in host rock	47	61
Growth of new carbonate crystals	53	13

replaced. Thin sections of some cherts display no carbonate matter but are clouded by a brown film. X-ray studies of these same cherts show the presence of a carbonate that is in many cases dolomite.

In some of the carbonate-clouded thin sections, replacement veins of calcite have developed. Around the veins the slide is clear in plane polarized light. In polarized, analyzed light, however, the quartz crystals show up as the usual steel gray mosaic. Farther away from the vein, the quartz is cloudy and brown. It appears that the calcite vein has served as a depository for the brownish carbonate material removed from the cleared area.

Fibrous quartz of the variety often called chalcedony is subordinate to the other kinds of quartz in the cherts. It normally comprises less than three percent of the quartz present in any nodule. Folk and Weaver (1952) observed that chalcedony was restricted to sites where a free surface was present for its nucleation. Their observation is confirmed in the cherts studied for this report. Fibrous quartz is found in vugs, fissures, and inside of fossils or other minute openings in the rock, but was not observed to replace the carbonate minerals as do microcrystalline and cryptocrystalline quartz.

Pelto (1956) has shown that chalcedony consists of quartz crystallites oriented with their C-axes normal to the length of the fibers. Crystallite C-axes do not lie in a common plane; rather, they tend to rotate about the fibers, which are not perfectly straight. The structural misfit between crystallites is accommodated by elastic straining of the quartz structure on either side of the misfit boundary in the event of small angle deviations. Where the angular divergence is large, the misfit is accommodated by dislocations between the quartz crystallites and by less elastic straining. Instances of large-angle misfit (greater than about 20°) are characterized by dislocations increasing in number as the angle increases. Misorientations of very large angle may have dislocations so closely spaced that the material takes on the character of a liquid.

In view of Pelto's discussion there seems no justification for the use of the term "chalcedony," so "fibrous quartz" is used here as a term that is at once more descriptive of the material and less likely to lead to confusion. The properties of fibrous quartz that differ from those of macroscopic quartz crystals are probably due to the poor organization and the adsorption of ions in the zones of bad fit.

The fibrous quartz in Illinois cherts developed in openings, generally small, and the initial deposits consisted of very short fibers extending perpendicularly from many closely spaced sites of nucleation on the walls of the openings. Successive groups of fibrous crystals grew on the free ends of preceding crystals to form rough zones or bands, and in each new group the nuclei were wider and the fibers longer than in preceding zones. This continued until the openings were filled or crystals for some reason ceased to form. Some of the zones are brown, others clear; fibers in some zones have a positive elongation, others a negative elongation. Some of these phenomena are shown in plate 2, figure D.

The fibrous quartz in small openings commonly consists of a single zone of fibers and shows the foregoing features less perfectly than does the quartz in larger cavities. The number of zones is largely dependent upon the size of the opening. The length of the fibers in successive zones varies, but in one large fracture filling the initial fibers were 5 to 10 microns long and developed from centers of crystallization 3 to 5 microns apart. Successive zones of fibers started from centers 10 to 20, 40 to 60, and 100 to 130 microns apart and the fibers were respectively 8 to 10, 20 to 40, and 160 to 290 microns long.

Carbonate Minerals

Carbonate minerals, either as individual crystals or host-rock masses, occur in all thin sections studied. Many nodules found in limestone have a zone of dolomite crystals near the chert-limestone interface. Although dolomite is common in cherts from limestones, nodules from dolomite formations contain little calcite.

Isolated masses of host rock, ranging from large polycrystalline fragments down to individual crystals, are common in nodules from all formations examined. These carbonate minerals have the same optical properties and, subject to their retaining sufficient size to contain them, the same primary structures and textures as the host rock. Moreover, structures are observed that originally must have been calcite but are now in whole or in part composed of quartz. This is particularly true of fossils. Fossils found in the chert, whether contained within carbonate rock fragments and composed of carbonate minerals or now completely quartz, are those characteristic of the formation.

Host-rock masses have been more or less embayed by the quartz. Many large crystals have been embayed to such an extent that they appear surrounded by small islands in optical continuity with the main mass. The degree of corrosion by quartz generally increases from the peripheral regions to the center of the nodule.

Other isolated carbonate crystals within the chert and immediately on the host-rock side of the chert-limestone contact have optical properties that differ from the minerals of the host rock. These crystals, which are subhedral to euhedral in outline, are called euhedral carbonates in this discussion. Examination with the universal stage, using the technique described by Emmons (1943), shows the euhedral carbonates to be dominantly dolomite with a few crystals of ankeritic composition.

The euhedral carbonates in the chert are free of opaque impurities to a much greater extent than are the host-rock minerals. They have never been observed to contain fossil fragments, oolites, or any other primary rock structures. They form sharp boundaries against both quartz and host-rock carbonate crystals, truncating any primary structure present in either. Euhedral carbonates seem to be attacked by quartz much more slowly than are the carbonate crystals of the host rock. They persist as sharp euhedra until zones deep within the nodule are reached where the host-rock carbonates have been reduced to shred-like masses. A few euhedra in the central regions of nodules show attack by the quartz in that the obtuse corners are corroded; the rest of the crystals retain sharp outline.

Thin sections of the host rock a few inches away from the interface show few euhdral carbonates. Beginning about one-half inch on the host-rock side of the interface, carbonate euhedra become common and continue to increase in abundance until a maximum concentration is reached at the interface of host rock and chert interface. Skeletal crystals are common near the chert and host-rock contact. Although found in both host rock and quartz, they are more common in the former. Some skeletal euhedral carbonates disregard crystal boundaries and incorporate portions of several crystals. Others are compositionally zoned as shown in plate 2, figure F. Universal stage measurements indicate that some of the zones are ankeritic and others dolomitic.

To summarize, the euhedral carbonates are marked by the following characteristics. 1) They are rare or absent in the host rocks except near the boundaries of chert nodules. 2) Their chemical composition differs from that of the host rock. 3) Even in coquinoid or oolitic rocks, the euhedral rhombs have not been observed to contain fossils or oolites. 4) They are always euhedral to subhedral and form sharp boundaries against the matrix except where corroded by quartz. 5) They are free of opaque impurities to a much greater extent than are the host-rock minerals. 6) They truncate all primary rock structures present.

These characteristics are evidence of an authigenic and epigenetic origin of the euhedral carbonates. Had they formed at the time of deposition the euhedral carbonates would not display truncation of primary structures, they would hardly be restricted to the neighborhood of chert nodules, and more subhedral to anhedral crystals would be expected.

Calcite occurs as fracture fillings in the cherts. Some of it may be derived from the limestone outside the nodule, but much of it seems to have been furnished by the carbonate minerals in the chert. The calcite is clear, free of opaque impurities, and occurs in large crystals. It truncates all other phases present. The chert around such veins is generally free of any but large crystals of carbonate minerals. Even the large masses are more corroded than similar material farther away from the vein.

Other Minerals

In addition to quartz and the carbonate minerals described above, fluorite and hydrous oxides of iron are present in some cherts as minor accessory minerals. A black material that is opaque but not metallic is observed in some thin sections. It is thought to be organic matter because it is most abundant in the more fossiliferous rocks.

Fluorite in cherts is limited in occurrence to the rocks near the fluorite mining area in southern Illinois. It is present in the cherts as isolated irregular bodies and as thin veins that replace the quartz preferentially leaving carbonate minerals apparently untouched. The veins and metacrystals are completely contained in the chert. They have not been observed to pass into the limestone at any place.

Reddish bodies of hydrated iron oxide occur around included masses of host rock in the chert as shown in plate 1, figure E, and in the host rock near the margins of chert nodules.

Mineral Relationships

The oldest mineral in the cherts is the carbonate of the host rock. Individual crystals, fossils, and masses of the host-rock carbonate have been observed that were cut by quartz, dolomite euhedra, and calcite fracture filling.

The dolomite rhombs truncate all structures in the host rock and are commonly found isolated in the quartz. They are in turn corroded by the quartz in some instances. It seems likely that the euhedral carbonates formed in and at the expense of the host-rock carbonates and were left in contact with the quartz by reason of their greater stability when the host rock was replaced by the quartz. Quartz truncates both host-rock carbonate minerals and the authigenic dolomite. Microcrystalline and cryptocrystalline quartz corrode all carbonate phases except the calcite that fills fractures. They are thought, therefore, to be the active minerals in the growth of the nodules.

Fibrous quartz is restricted to void fillings; it occurs only on free surfaces. The youngest mineral is the fracture-filling calcite. The fractures may have formed at any time after the final consolidation of the rocks. Such veins cannot be traced far into the carbonate rock but in the chert they are quite conspicuous.

FORMATION OF CHERT NODULES

Pettijohn (1957) recently summarized present theories of the origin of chert. Some workers in the past have favored the hypothesis of syngenetic origin, others have found evidence that the chert bodies were formed by replacement of the host rock.

The most commonly accepted criteria supporting the replacement concept are those given by Van Tuyl (1918): 1) the occurrence of chert along fissures in limestone; 2) the very irregular shape of some nodules; 3) the presence of irregular patches of limestone in some cherts; 4) the association of silicified fossils and chert in some limestones; 5) the presence of replaced fossils in some cherts; 6) the failure of some cherts to follow definite zones in limestone formations; 7) the occurrence of silicified oolites formed by the replacement of calcareous ones.

Additional evidence of replacement is furnished by two points suggested by the present study: 1) small quartz bodies in the interstices of the host rock near nodules, and 2) authigenic growth of carbonate minerals of a composition different from that of the host rock associated with the chert.

Small bodies of quartz in the interstices of the host rock are of irregular shape and random disposition about the nodule. They are not always present, but the fact that they are encountered at all is evidence that the associated nodule was never a subspherical mass of silica gel being rolled about on the sea floor. Had this been the case the small bodies surely would have become incorporated in the larger one. Moreover, some of the interstitial bodies embay and partially replace fossils, oolites, and other host rock structures.

The growth of authigenic carbonate minerals associated with chert nodules cannot be explained as a phenomenon of syngenetic chert formation because the authigenic crystals are associated with the lower boundaries of the nodules as well as with their sides and tops. The rhombs concentrated in the host rock near nodules, and distributed through the isolated carbonate masses and the chert proper, have been shown earlier to be authigenic and epigenetic.

Field Relationships

In the field, the following evidences of the replacement of chert nodules in carbonate rocks of Illinois are found: 1) masses of host rock are clearly visible in some nodules without the aid of magnification; 2) remnant bedding passes through some nodules; 3) some nodules have the oolitic or coquinoid texture of the rock preserved in silica; 4) euhedral carbonate crystals, some of which are much larger than crystals of the host rock, appear in some nodules; 5) fossils in all stages of replacement are found in many nodules; and 6) zones of silicified carbonate rock appear at the borders of some chert nodules.

Three negative evidences to support the replacement hypothesis of manner of origin were observed: 1) no nodules were found with bedding planes bent around them; 2) nodules are not consistently related to any primary structure of the host rock; and 3) dessication cracks, which would be expected in abundance as a result of dehydration of a silica gel, are absent.

Host-rock Inclusions

Perhaps the strongest evidence of epigenetic origin is the presence, in all cherts examined for this study, of host-rock material. The occurrence of host-rock carbonate has been described in an earlier section. It varies from large polycrystalline masses down to single crystals. Masses of host rock within a nodule are shown in figure 1, and the texture of one of the masses is shown in plate 1, figure E.

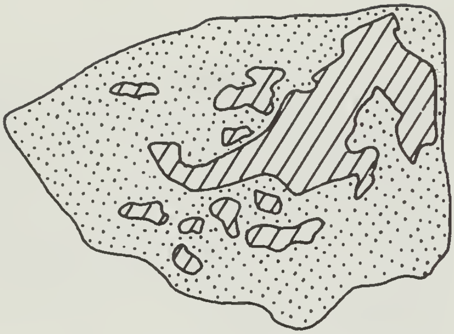


Fig. 1. - Host-rock masses in chert. Generalized sketch of a nodule from the Galena formation at Menominee Station, Illinois.
Stippled = chert; lined = carbonate
About actual size

Remnant Bedding

Remnant bedding passing through nodules is strong evidence in support of the replacement concept. Syngenetic chert nodules probably would not be bedded at all, and, if they were, it would be difficult to account for their bedding's being continuous with that of the surrounding rock. Relict bedding was observed both on the outcrop and in thin section. A line drawing of relict bedding in a nodule from the Kankakee formation at Savanna, Illinois, is shown in figure 2.

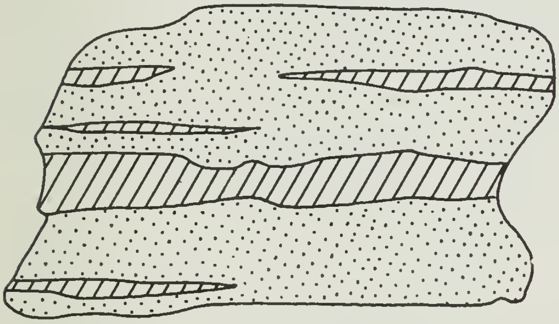


Fig. 2. - Relict bedding in a chert nodule from the Kankankee formation at Savanna, Illinois.
Stippled = chert; lined = carbonate

Preserved Texture

The preservation of oolitic or coquinoid texture of the carbonate rock in the chert is strong evidence of an epigenetic origin for the chert nodules. The texture and fabric of the host rock passes without interruption into the chert. This is irreconcilable with the syngenetic hypothesis, which holds that the regions occupied by the nodules had never been filled with the host rock.

Euhedral Carbonate Crystals

The presence of euhedral carbonate crystals in chert might be taken as evidence of a syngenetic origin. It might be argued that such crystals grew in the soft silica gel and so were able to complete their euhedral form. This argument would have some validity but for the fact that crystals of the same kind are found in the host rock at nodule boundaries. Close examination reveals that most of the euhedral carbonates are associated with host-rock masses in the chert and many of them actually are contained therein. The relationships of the euhedral carbonates to nodule boundaries and included host-rock masses can best be interpreted as due to epigenetic formation of the chert.

Replaced Fossils

Many calcareous fossils are found in the chert that have been partially or completely replaced by quartz (plate 1, fig. B). It has been argued that fossils are replaced by being dropped into dehydrating silica gel on the sea floor during sedimentation. The silicified fossils found in this study generally are associated with included host-rock masses that are somewhat replaced by quartz. The fossils, even when separated from recognizable masses of carbonate rock, are not found at or near the bottom of nodules, as would be expected of organisms dropped into a gelatinous silica, but are randomly distributed throughout the mass.

Silicified Carbonate Rock Zones

Zones of silicified carbonate rock at the edges of chert nodules have been thought to be due to the pressing of carbonate sediments into the still gelatinous silica. This argument fails to account for two conditions. First, the silicified zones are always as thick, and in some instances thicker, around the sides of nodules as at their tops and bottoms. Simple overburden pressure intruding partly consolidated sediments into silica gel would be expected to leave the sides of nodules more or less unaffected.

Second, a dehydrating silica gel is expected to harden in the peripheral regions first. If impression of carbonate sediments occurred, it would be expected that as deeper, softer zones of the gel mass were reached impression would be easier until finally the carbonate would completely fill the region formerly occupied by the silica gel. The silica would thus be reduced to an interstitial role in a normal carbonate rock.

Bent Bedding Planes

Nodules around which bedding planes of carbonate rock are bent are commonly thought to be syngenetic. Actually such a bending of bedding planes around nodules could simply represent differential compaction of the carbonate rock and the chert and merely mean that the host formation has undergone some local change in thickness since the chert nodule formed. This phenomenon was not observed in the rocks studied.

Nodule Distribution

Syngenetic chert nodules might be expected to be associated with bedding planes in the rock. It would be during temporary cessation of carbonate deposition that silica gel would be most likely to form independent nodular masses on the sea floor, and such gel bodies, if covered by later deposits after a reasonable degree of dehydration, might form chert nodules. The nodules found in this investigation are not restricted to such surfaces of nondeposition of carbonate. Rather they are randomly distributed throughout the rocks and are as common in thick beds as in the bedding planes between strata.

Dessication Cracks

Dessication cracks would be expected if silica gel bodies had hardened into chert nodules. According to Iler (1955) silica gels are highly variable in water content, some gels having much more water than silica. The exact nature of the silica gels that have been postulated to explain chert nodules has not been stated but it is presumed that a considerable fraction of the volume would be water. Chert is anhydrous. The supposed hardening of a gel body from the outside invites the supposition that the interior portions could harden only as rapidly as water could be lost. This would almost certainly demand that cracks develop in the outer regions. There are no evidences of dessication cracks in the nodules studied.

Thin-Section Evidence

In thin-section studies, replacement manner of origin was indicated by 1) the presence of relict masses of host rock; 2) the presence of pseudomorphs after single crystals or aggregates of crystals; 3) the presence of pseudomorphs after host-rock structures; and 4) the growth of authigenic carbonate minerals of a composition different from those of the host rock in and near nodules.

Relict Host Rock

Relict patches of host rock, noted in hand-specimen studies, were also observed in thin section. These patches range in size from fragments of a single crystal to polycrystalline masses 15 centimeters long. In thin section they have not been appreciably changed except by the invasion of quartz around grain boundaries and along cleavage planes. The carbonate crystals at the periphery of the isolated patches are more embayed than those at the center. In many cases, the peripheral grains are highly embayed and partially fringed by isolated fragments of carbonate material in optical continuity with the neighboring crystals of the main mass.

Relict patches of host rock in various stages of replacement are shown in plate 1, figures E and F, and plate 2, figures A and B.

The dolomite rhomb at the center of the photograph in plate 1, figure F, is in optical continuity with the dolomite of the surrounding crystal. The dark zone around the rhomb at the center is microcrystalline quartz. The replacement seems to have taken place along cleavage planes in the original crystal.

Pseudomorphs After Carbonate Crystals

In many thin sections, pseudomorphs after carbonate crystals are found in which single crystals of quartz replace several carbonate crystals but preserve the outline of the original carbonate. Plate 1, figure A, shows a single quartz crystal replacing a mosaic of dolomite crystals. The original outlines of the carbonate crystals appear as light lines in the photomicrograph.

Explanation of Plate 1

A. - Unit metacryst of quartz growing at the expense of dolomite and microcrystalline quartz. Shakopee formation near Utica, Illinois. Crossed nicols, 200X.

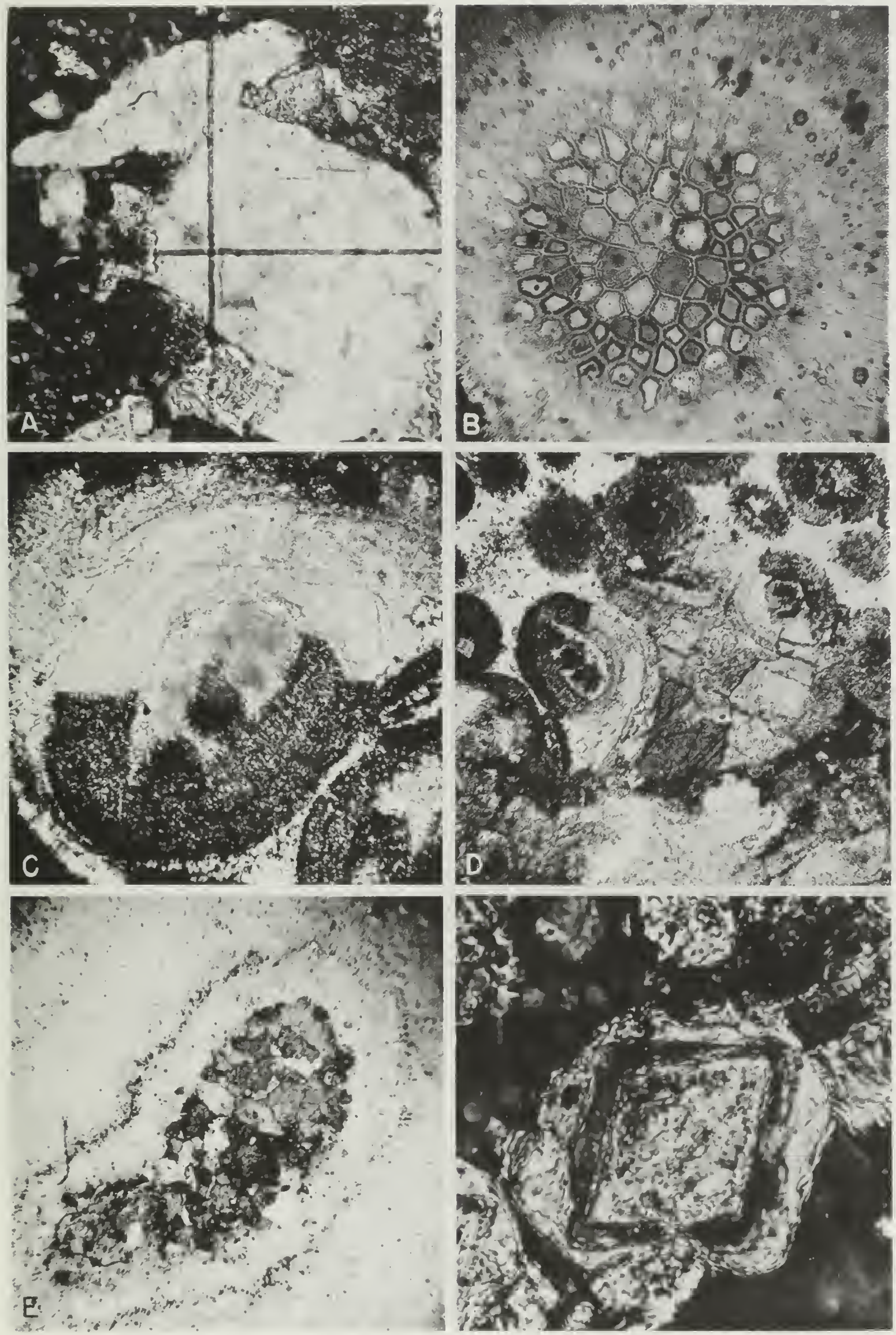
B. - Preservation of a coral upon complete replacement by quartz. Ste. Genevieve formation near Anna, Illinois. Plane polarized light, 25X.

C. - Same as figure D but magnified 57X.

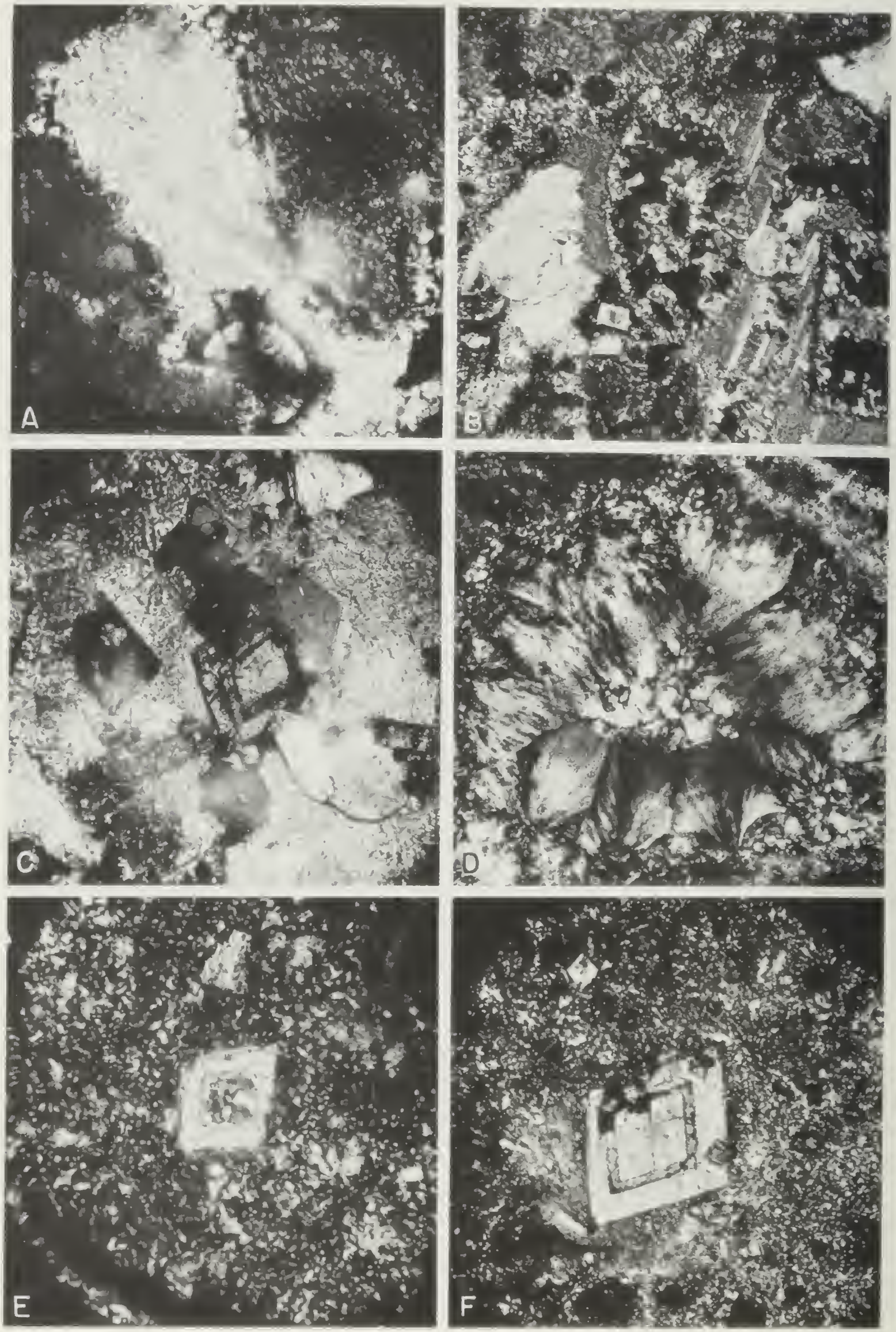
D. - Calcareous oolite (light) partially replaced by quartz (extinct). Ste. Genevieve formation. Crossed nicols, 25X.

E. - Mass of dolomite isolated within chert. Note the dark ring of iron oxide surrounding the dolomite. Galena formation near Menominee Station, Illinois. Plane polarized light, 15X.

F. - Selective replacement of dolomite along cleavage planes by microcrystalline quartz. The core and rim of the dolomite are optically continuous. The dark rhomb-shaped zone is microcrystalline quartz. Shakopee formation near Utica, Illinois. Crossed nicols, 380X.



Biggs — Petrography and Origin of Nodular Cherts


A
B
C
D
E
F

Explanation of Plate 2

A. - Carbonate crystal (at maximum illumination) between other carbonate crystals at extinction displaying a ring of interstitial quartz crystals and a fibrous quartz rosette. Burlington formation near Pearl, Illinois. Crossed nicols, 130X.

B. - Partially replaced mass of host rock. The quartz in the crinoid stem plate is at maximum illumination; the calcite is at extinction. Ste. Genevieve formation near Anna, Illinois. Crossed nicols, 25X.

C. - Compositionally zoned dolomite crystal in limestone that has replaced parts of several calcite crystals. Burlington formation near Pearl, Illinois. Crossed nicols, 130X.

D. - Chalcedonic quartz showing a dark band at the tips of fibers which may be due to concentration of light scattering pores. Burlington formation near Pearl, Illinois. Crossed nicols, 150X.

E. - Dolomite metacryst probably like that of figure F with the carbonate mineral at the center replaced by microcrystalline quartz. Burlington formation near Florence, Illinois. Crossed nicols, 130X.

F. - Compositionally zoned carbonate crystal in chert. The core and rim of the crystal are ankeritic and the intermediate zone is dolomite. Salem formation, cen. NW 1/4 sec. 18, T. 13 S, R. 1 E. Crossed nicols, 57X.

The origin of these lines is not understood, but they may be due to slightly different composition of the original material at crystal boundaries. The quartz is a single crystal that extinguishes as a unit. A dolomite crystal partially replaced by quartz is shown in the lower central portion of the photomicrograph.

Pseudomorphs After Host-rock Structures

Pseudomorphs in which aggregates of quartz crystals replace carbonate rock structures are shown in the completely quartzitized coral in plate 1, figure B, and in the partially replaced oolites in plate 1, figures C and D. Both the coral and the oolites are from the Ste. Genevieve formation near Anna, Illinois. In figure C, the two large oolites have been partially replaced by quartz. The quartz here is at maximum illumination and the calcite is near the extinction position. The white material between oolites is fibrous quartz. In the upper left portion of the figure, other oolites that have been completely replaced are seen. Figure C is the upper large oolite of figure D photographed at greater magnification and rotated so that the quartz is at extinction and the calcite is at maximum illumination.

It is improbable that the oolites originally formed with this composite arrangement and also unlikely that the oolites were partially replaced in a dehydrating silica gel. They are part of an uninterrupted oolitic fabric that extends into the host rock. Corals such as that shown in plate 1, figure B, are known to be calcareous forms. This coral has been completely replaced by quartz. It does not lie in the bottom of a nodule, as would be expected had the replacement occurred after it had been covered by a deposit of silica gel.

Authigenic Carbonates

Euhedral carbonate crystals are shown in plate 2, figures C, D, and E. The skeletal rhomb in figure C truncates several crystals of the surrounding limestone. This photograph of the limestone was taken about two millimeters from the chert-limestone interface. The number, and degree of completion, of euhedral rhombs increases toward the interface.

A large euhedral rhomb is seen in plate 2, figure F. It is a zoned ankerite completely surrounded by quartz. Several smaller rhombs are seen nearby. Dark, vaguely defined masses in the lower part of the photograph are ghosts of completely replaced carbonate crystals that have left small amounts of impurities marking their former position.

A skeletal dolomite rhomb is shown in plate 2, figure E. It is surrounded, and the opening at the center is filled, by microcrystalline quartz. Probably the replacement was controlled by the presence of carbonate of slightly less stable composition at the center of the rhomb. If this was the case, quartz has now completely replaced the less stable phase.

The relationships of the authigenic carbonates to the boundaries of nodules indicate genetic control by the chert. The rhombs are found in and at the borders of nodules. Those in the central portions of larger nodules seem to have been attacked by the quartz to a greater extent than those at the margins.

Evaluation of Evidence

There is thus no lack of evidence for the epigenetic origin of cherts studied. Remnant bedding and stringers of carbonate rock in nodules can hardly be explained if it is supposed that the chert originated as gel bodies on the sea floor.

The preservation of textures and structures implies that the material that is now chert was once homogeneous with the surrounding carbonate rock. It cannot be assumed that oolites and fossils were dropped into a dehydrating gel and there replaced by quartz for this process would have scattered the gel if it were weak enough to permit entry of such small bodies. Moreover, if the gel were very soft and fluid, the likelihood of complete, rather than the commonly observed partial replacement, probably would be increased.

Patches of host rock with the same fabric and structure as the surrounding carbonate would not be likely to result from crystallization of carbonate muds and fossil fragments placed in a silica gel. Some such relict patches of host rock are fifteen to twenty centimeters in the maximum observed dimension. For a silica gel body on the sea floor to maintain its identity and shape against overburden pressure as it becomes more deeply buried, it must dehydrate and solidify in a comparatively short time. The shrinkage involved in this process would vary with the silica-to-water ratio of the gel but in any case it would not be negligible (Iler, 1955). The forces exerted on the included limestone patches by the shrinkage would not be the same as those on the surrounding rock of the same composition. The more or less centripetally distributed forces set up by the shrinkage might tend to mold the unconsolidated carbonate material to a more nearly spherical shape and move it toward the nodule's center. Such forces would certainly not promote, and probably would tend to destroy, the very irregular nature of some relict patches of host rock.

Masses of silica gel might logically be expected to solidify from the outside inward. If this happens it is difficult to explain the silicified limestone around some nodules. If the carbonate material was impressed into the gel while it was soft, a question arises as to what prevented its intruding all the way into the nodule.

Pseudomorphs after crystals and rock structures cannot occur in syngenetic chert nodules because presumably no rock structures or crystals of the carbonate host rock were there.

The association of authigenic dolomite with the chert also would be most improbable if the chert were syngenetic. Syngenetic chert probably would demand syngenetic dolomite, if the relationships that have been observed were found. The dolomite would, of necessity, be deposited at the same time as, and only in association with, the nodules and not between the nodules. Even by this hypothesis it would be difficult to account for the euhedral dolomite crystals that are associated with the lower surfaces of the nodules.

Such evidence supports the belief that chert nodules in Illinois limestones and dolomites are epigenetic concretions formed by metasomatic processes. The silica that composes the nodules is considered to have been deposited synchronously with the carbonate rock as a dispersed phase that aggregated into nodules during diagenesis.

Source of Silica

Chert nodules may be formed from silica introduced into carbonate rocks in several ways. Volcanic springs active on the sea floor may contribute silica directly to the accumulating sediments; hydrothermal solutions may introduce silica into already lithified rocks; groundwater may dissolve silica from siliceous rocks and deposit it in carbonate environments; and silica dissolved from rocks undergoing subaereal weathering may be transported to the sea and there deposited with the carbonate sediments.

Volcanic Springs

Volcanic springs would be expected to leave some evidence of their operations. Silica deposited from them probably would be localized about the vents. Silicified pipes should be found leading downward into older rocks. Moreover, such springs would be localized in one or several portions of the deposition basin, not uniformly distributed over wide areas. Evidence of none of these things was observed in this study.

Hydrothermal Solutions

Hydrothermal solutions supplying silica to lithified rocks would be subject to the same restrictions as volcanic springs. Instances where chert is shown to have been deposited from hydrothermal solutions (Fowler et al., 1935) are restricted to narrow zones of ore deposition. The lead ores in northwestern Illinois and the fluorite ores of southern Illinois are not associated with hydrothermal chert, nor is there any evidence that hydrothermal solutions have produced the chert nodules found elsewhere.

Groundwater Deposits

Chert deposited from groundwaters should be restricted to zones of groundwater movement. Thus, chert nodules might be expected to accompany other groundwater deposits. They are not so found. Solution channels are not more common in cherty rocks than elsewhere. Instances in which carbonate rocks have been silicified by groundwaters moving downward through siliceous rocks are known (Rubey, 1952), but the replacement is complete, not nodular. Moreover, the silicified limestone noted by Rubey contains preexisting chert nodules.

Weathered Silica Deposits

Silica dissolved in sea water, derived from any source, may be deposited with the sediments. Clarke (1924) estimated that the annual increment of silica contributed to the sea by weathering was 319,000,000 tons. The amount of silicon contained therein would be 150,000,000 tons and is exceeded in Clarke's computations only by calcium (557,000,000 tons) and sodium (258,000,000 tons). Silicon freed by weathering and brought to the sea in streams is thus a more than adequate source for the chert found in carbonate rocks.

Whether the silicon is deposited by organisms or inorganic means is not known in all instances. The fact that silicon is the fourth most abundant cation in limestones (Rankama and Sahama, 1950) is evidence enough that large quantities of it are deposited with the rocks. Pettijohn (1957) cited the presence of interstitial quartz bodies in limestone and they have been observed near chert nodules in this study.

Evaluation of Sources

The very widespread occurrence of chert, both areally and stratigraphically, and its random distribution make a volcanic, hydrothermal, or groundwater source of silica most improbable. Evidence for the precipitation of silica from sea water is found in 1) the large amount of silica freed by weathering, 2) the large amount of silica carried annually to the seas by rivers, 3) the low concentration of silica in sea water, 4) abundance of silica-secreting organisms and ample opportunity for inorganic precipitation of silica known to exist in the sea, 5) the presence of silica in limestones in a form other than megascopic chert nodules, and 6) the presence of silica in recent carbonate sediments.

Concentration of Silica to Form Chert Nodules

Recrystallization tends to occur when rocks are subjected to changes in pressure and temperature. Ramberg (1952), who discussed the relationships of rock components during recrystallization, was concerned chiefly with conditions in the field of metamorphism where both pressure and temperature of rock masses vary over wide ranges.

Diagenesis can be separated from some forms of metamorphism as being different only in the degree of severity of pressure and temperature variations. Correns (1950) defined diagenesis as changes (other than weathering) in a sediment between its sedimentation and metamorphism. He found little difficulty in separating changes due to weathering from those of diagenesis but admitted that the boundary between diagenesis and metamorphism is an arbitrary one.

Perhaps changes in pressure are more important than variations in temperature in causing diagenetic changes. Correns (1950), Pettijohn (1957), and many other workers regard diagenesis as a low-temperature process; the simple overload of superjacent sediments, however, imposes rather large changes in pressure. It may be expected then that it is safe to ignore the effects of temperature in considering the formation of chert nodules.

Diagenesis of carbonate rocks includes recrystallization of phases formed at the time of deposition. Such recrystallization possibly begins almost immediately after deposition and continues at varying rates for an indefinite period. The process is not simple and so no rate can be assigned to it that would be valid for even identical components in close proximity.

Diagenetic recrystallization may be considered from the standpoint of two relationships stated by Ramberg (1952) as

$$\left(\frac{\delta F}{\delta P}\right)_{TX} = V \tag{1}$$

$$f_c = \left(\frac{\delta P}{\delta F}\right)_{TX} = \frac{1}{V} = \frac{d}{M} \tag{2}$$

where F is the free energy of a phase, P is the pressure on the phase, V is the molal volume of the phase, f_c is the force of crystallization, d is the density of the phase, and M is its molecular weight. The subscripts T, X indicate that temperature and composition of the phase are held constant.

If temperature, T, and pressure, P, in the rock are constant, the constituent mineral phases and interstitial fluids are in equilibrium with themselves and each other; the molal free energy of the crystal, F_c, and the corresponding components in the surroundings, F_s, are equal. Because we have made the assumption that T does not vary widely in diagenesis we shall ignore its changes; but if P increases then F must increase, and the phase in question becomes unstable if the pressure is applied to it and only to it. In this case, the crystal should dissolve in whatever phase or phases surround it.

In the event that $F_s > F_c$ the crystal may continue to grow even though the pressure on it is greater than that on its surroundings. The force of crystallization was defined by Ramberg (1952) as "the excess pressure which must be applied on a crystal in order that the crystal may be in chemical equilibrium with its surroundings, which are supersaturated to a given degree."

In a rock containing only a few crystals of a mineral and none of its constituent parts in the pore fluid, the crystals in question should perhaps disappear during recrystallization. If, however, a phase makes up an appreciable part of the rock, and the pore fluid may be expected to contain some of its constituent ions, the probability rises that any given crystal of the phase may be able to increase in size from dissolved constituents. If some crystals by reason of a favorable position continue to grow, the removal of constituent ions from the pore fluid permits further decomposition of nearby crystals that are unstable. Thus a mechanism whereby larger crystals grow at the expense of the smaller is established.

Recrystallization of limestones after deposition leads to the growth and purification of crystals. Any silica occluded in the carbonate minerals or crystallized in the rock is affected by the process just as the carbonate minerals are. Some quartz crystals whose free energy is lower than that of SiO_2 in the surroundings will remain stable during recrystallization and grow at the expense of other quartz. In such a densely packed medium the space for growth afforded to the quartz is slight indeed. A growing crystal can replace other crystals provided it has a higher force of crystallization and a supply of constituent ions. The force of crystallization of quartz, by equation 2 above, is greater than that of calcite or dolomite. Therefore, we should expect to find that replacement of carbonate by quartz is common, but that replacement of quartz by carbonate is not so frequently obvious. Carbonate crystals must replace quartz crystals, however, where ever the phases carbonate and quartz are in contact and the surroundings are supersaturated with carbonate constituents but deficient in silicon ions.

Perhaps not all quartz found in the rock formed around a nucleus of quartz that was stable from the beginning of diagenesis. Activated particles, Si^{++++} and O^{--} ions, may nucleate quartz crystals on mineral interfaces when, by reason of locally lower pressure or other anomaly, crystals may form at weak supersaturation. Once formed, these newly nucleated crystals can grow at the expense of surrounding crystals.

Favorable sites for such nucleation are probably found in 1) open voids formed by the shells of organisms; 2) small fractures opened in the rock; and 3) pressure shadows developed around coherent bodies in the rock. These, or similar sites, provide low-pressure areas where crystal nuclei may form from relatively dilute pore fluids. Once nucleated, quartz crystals continue to grow if a supply of constituent ions is available or if the crystal is stressed beyond its equilibrium value they will resorb.

Theoretically, quartz growth should fill open spaces before new growth begins replacing other crystals. The actual growth of quartz bodies in carbonate rocks proceeds in this fashion, as is evidenced by the incomplete replacement of carbonate minerals at the nodule boundary and by the occurrence of small quartz bodies in the interstices of the limestone outside the nodule. The apparent method of growth is complex, involving the formation of a small quartz core surrounded by a diffuse halo of partially replaced host rock. The diffuse area continues to grow outward into the limestone accompanied by growth of the more or less carbonate-free core of firm chert.

The excellent preservation of structures and textures and the lack of any evidence for displacement of the structures suggest that the replacement is isovolumetric. Barth (1948) supposed that such replacement occurred without movement of the oxygen ions. Concerning the role of oxygen in replacement he says :

"In view of what has been said in the preceding paper about the role of the oxygen ion in rocks and in mineral lattices, it is reasonable to suppose that the mechanism of such isovolumetric alterations is that of a migration and exchange of cations in a medium composed of relatively stationary oxygen ions. Only in this way can one explain the preservation of the delicate structural features; for removal of the large oxygen ions would break up the minerals and destroy the fragile and fine structure patterns. In harmony with this idea is, e.g., the mechanism of the weathering of biotite (bleaching, baueritization), which is effected by selective removal of the metal ions while oxygen and silicon remain in the residual lattice."

The shape of chert nodules is dictated by the ease of replacement of the host rock. Because the host rock is essentially isotropic in this respect, the nodules often approach sphericity. Because they are somewhat flattened parallel to the bedding of the host rock, the nodules commonly have the form of oblate spheroids. This flattening is due to the greater ease of replacing the host rock in the horizontal plane. Replacement in the vertical plane must take place against overburden pressure. Pressure on the upper and lower surfaces of the nodule would be expected to be greater than that on the sides, and would thus lower the growth rate of the quartz.

Zones in the host rock that are, for any reason, more resistant to replacement than others, complicate the shape of the nodules. A stylolite or shale zone in the limestone commonly results in a flat side on an otherwise rounded nodule. Smaller inhomogeneities in the limestone cause indentations in the surface of the nodule.

The chert-limestone contact generally is irregular as seen in thin section. Stringers of quartz invade the interstices between calcite crystals and partially replace them. Individual crystals of carbonate minerals and islands of host rock are more abundant near the contact than deeper within the nodule. The central portions of large nodules generally are composed of quartz with little or no carbonate present.

The contact area is in many instances marked by large, concentrically zoned carbonate rhombs. The rhombs are most common on the host-rock side of the interface. Their composition varies from dolomite to ankerite. The position of the zoned rhombs, crowded at the chert-limestone interface, suggests that they are products of the interaction of the quartz and the carbonate of the host rock. This suggestion is strengthened by the fact that the zoned carbonates are more common in impure limestone than in other rocks.

If calcite crystals containing small amounts of magnesium and iron were replaced by quartz, it is probable that the two impurity elements would find more stable positions by concentrating at the replacement front. At the contact of the limestone and quartz, crystal structures are already weakened. The magnesium and iron would find entry and be able to form mixed crystals with the calcium already present. Such crystals would be more stable with respect to the quartz than clacite with slight impurities because enough magnesium and iron would be added to complete the structure. The mixed crystals formed would have a lower free energy and better ordering, and so be more stable than the solid solution of the original calcite.

The increased stability of the dolomite rhombs is shown by the fact that they are less corroded in the chert than are similar fragments of limestone.

Special Characteristics of Chert in Dolomite

Chert nodules formed in dolomite contain fewer fossils than do those found in limestone. This is possibly due to the destruction of fossils during the dolomitization. If the paucity of fossils is due to such destruction, the chert nodules must have been formed at or later than the time of dolomitization. Generalizations can not be made about the time of dolomitization of limestones as each dolomitized formation is an individual problem. Careful study of the relations of the chert nodules to the host rock might help to solve the time-of-dolomitization problem.

Nodules from dolomites contain more carbonates than do those from limestones, and the chert fills interstices to a greater extent. This may be due to the aggregation of chert after the limestone has been replaced by dolomite. Perhaps the dolomite was more stable with respect to quartz than calcite, and the silica was forced to fill interstices to a much greater extent.

AGE OF CHERT NODULES STUDIED

The problem of assigning an age for the chert nodules has not been solved. Early workers held that chert was formed only at the weathering interface. This idea became untenable, however, as well records showing the presence of chert far from the outcrop became common.

Other ideas of a late genesis have not been so easily disproved. The general lack of association of chert nodules with ore deposits, groundwater channelways, or other secondary features of the rocks weighs against any hypothesis claiming the chert to be much younger than the enclosing formation.

The presence of residual chert conglomerates in Illinois was noted by Poor (1925). He found that the detrital chert in the basal conglomerate of the Pottsville formation in southern Illinois contained fossils of Chester age (late Mississippian). Chert pebbles bearing Chester age fossils could not occur in the Pottsville formation if the genesis of the chert was much later than the deposition of the Chester rocks. The same relationship of Mississippian age chert pebbles in the basal conglomerate of the Pottsville formation was observed by Rubey (1952) in the Hardin and Brussels quadrangles.

Delicate fossils commonly are found in the chert whereas their counterparts in the limestone are shattered. Bryozoa found in the chert of both the St. Louis and Ste. Genevieve formations are better preserved than those in the surrounding limestone. The most probable reason for better preservation is that the fossils were incorporated in chert early in the history of the deposit before overload crushed them. The chert would resist crushing much better than would the relatively incompetent limestone. Some crushed fossils have been found in the cherts but they are thought to have been broken before replacement. If the better fossil preservation is due to protection by a more rigid matrix, chertification must have preceded or at least accompanied the compaction and lithification of the rock.

In thin section the same replacement phenomena are seen. Delicate macrofossils are better preserved in the chert than in the limestone. Included host-rock patches, however, have the same texture as the surrounding rock, so the nodules probably are diagenetic in age rather than syngenetic. That host-rock patches in a rigid chert nodule would undergo recrystallization resulting in the same texture as that of the host rock outside the nodule is unlikely because the patches would not have had the same thermal and pressure history as the carbonate minerals outside the nodules.

The high porosity and permeability of poorly consolidated sediments (Pettijohn, 1957) and the plentiful supply of chemically active aqueous solutions would favor the aggregation of chert during diagenesis. Better preservation of delicate fossils in the cherts and the presence of residual cherts of an age but little younger than the parent formation are evidences favoring an early time of origin for chert. The similar texture of host rock outside the nodules and included host-rock patches is evidence that the chert was aggregated no earlier than the recrystallization of the host rock.

It seems that many, if not most, chert nodules in the rocks studied, originated during the recrystallization of the host rocks. The time that the nodules ceased to grow is harder to state. Probably they increase in size until the chemical and physical factors controlling the process cease to operate because of interference by outside factors. The rate of growth probably has varied from one nodule to another, and probably has not been uniform within the same nodule. No evidence of recent growth of nodules was found. The growth probably ceases altogether when rocks come into the zone of weathering.

SUMMARY AND CONCLUSIONS

The silica mineral present in the nodular cherts of the carbonate rocks of Illinois is quartz. This determination, based on study of material in thin section, is supported both by infrared absorption and x-ray diffraction studies. All cherts examined contain some carbonate minerals. Some of the carbonate materials are fragments of host rock or individual crystals having the distinctive markings of the crystals in the host rock. Others are authigenic dolomite crystals which generally differ in composition from the carbonates of the formation. Quartz and the carbonates, calcite and dolomite, make up the bulk of the chert. Fluorite and iron oxide are present in minor portions in some chert.

The only evidence that might favor a syngenetic origin of the nodules is the presence of better preserved fossils in the chert than in the carbonate rock. As has been shown, this is really a criterion of age, not manner of formation. Other evidence of syngenetic origin is lacking. Remnant bedding, textural preservation, patches of host rock, authigenic dolomite crystals, partially-to-completely replaced fossils, rim zones of silicified limestone, pseudomorphs of crystals or aggregates of crystals, and pseudomorphs of rock structures in the nodules are all considered evidence that the nodules originated by replacement of the pre-existing host rock.

The chert nodules and the rocks in which they occur offer no evidence favorable to the hypothesis that the silica was introduced after the consolidation of the original rock. It is certain that siliceous organisms, which are rare in Illinois rocks, cannot be considered a source of silica for chert formation.

The most probable source of the large quantities of silica found in cherts is the weathering of land masses. Large quantities of dissolved silicon are brought to the sea each year. Moreover, it is known that Si^{++++} is not concentrated in the sea but must be deposited almost as fast as it is brought into the sea. Silica is thought to be deposited along with other sediments because they commonly show siliceous cementation, and because Si^{++++} is the fourth most abundant cation in limestones.

Silica may be deposited with the carbonate rocks by organic or inorganic processes and probably precipitation actually occurs due to both processes. Seemingly, the inorganic mechanism would be more nearly ubiquitous than organic precipitation, but organisms may be locally important. The presence of silica in the limestones is a matter of record, though further studies should be made to show quantitatively how much quartz a given limestone holds.

The preservation of fossils and rock textures in chert, seems to indicate that the replacement occurred between the beginning of diagenesis and uplift. The replacement of the limestone was not complete. Fragments of structures and textures of the host rock present in the chert refute the syngenetic hypothesis and make it possible to understand the origin of these cherts.

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