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GEOLOGY OF THE NAZARETH QUADRANGLE, NORTHAMPTON COUNTY, PENNSYLVANIA

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By

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This report is preliminary and has not been edited or reviewed for conformity with Geological Survey standards

Chapter I

INTRODUCTION

Geologic Setting

The Appalachian Mountains of the eastern United States have stimulated the interests of geologists for nearly 200 years. The forests, waters, coal, oil, and mineral resources of this region provided the energy and raw materials that supported colonization and fostered the industrialization and economic growth of the United States. Some of the fundamental concepts of geology evolved from the observations and imagination of some of the early geologists who worked in the region, and the sparks of some of the great geologic controversies were struck there as well.

Despite this rich legacy, however, many geologic problems remain unsolved. Each new generation of geologists brings new techniques, new ideas, and new definitions to what are essentially old problems, and new solutions to some of these problems commonly raise as many questions as they answer.

The Great Valley and Reading Prong are physiographic sections of the middle Appalachians. The Reading Prong is a belt of prominent ridges that extends 140 miles northeastward from Reading, Pennsylvania, through northwestern New Jersey (Jersey Highlands) and southeastern New York into western Connecticut (fig. 1). The Prong is bounded on

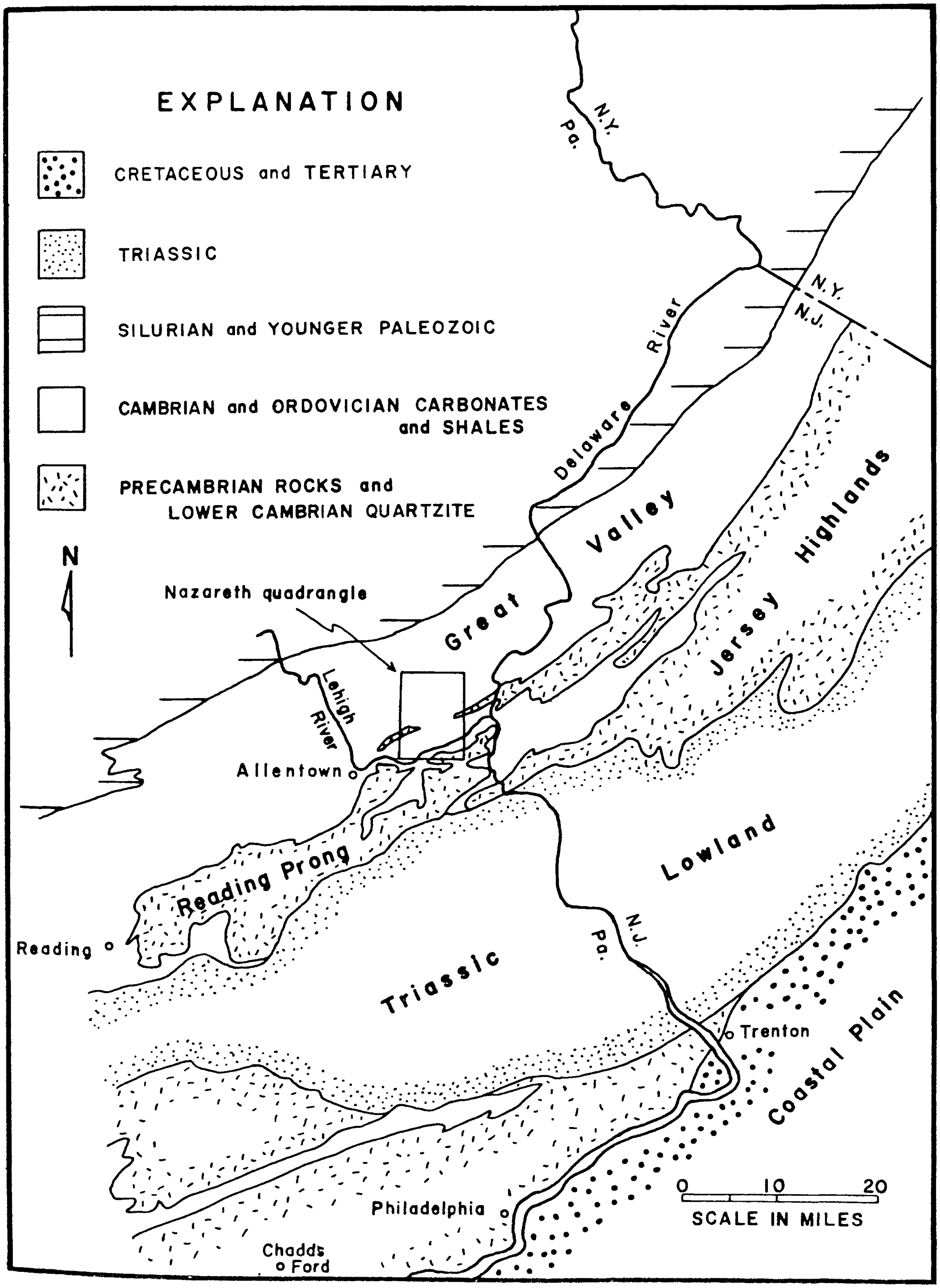


Figure 1.—Generalized geologic map of eastern Pennsylvania and western New Jersey.

the northwest throughout most of its length by the Great Valley, a broad, flat valley that is known by a host of local names. The ridges that constitute the Prong are composed largely of a variety of Precambrian plutonic and high-grade metamorphic rocks. The Great Valley and numerous major valleys within the belt of the Precambrian rocks of the Prong are underlain by Cambrian and Ordovician carbonate and pelitic rocks (fig. 1).

The geology of the Nazareth quadrangle, an area that comprises parts of both the Reading Prong and Great Valley in eastern Pennsylvania, is the subject of this dissertation. The general area has been studied by five generations of geologists, partly because of the variety and continuing interest of the geology, and partly because of the value of its natural resources, which include lead, zinc, and iron ores, commercial slate, limestones ideally suited to the manufacture of portland cement, agricultural limestones and dolomites, dimension stone, brick clay, sand, and gravel. However, in spite of the wealth of previous work in the area, perhaps because of it, there remain many unresolved conflicts and many important questions.

This study focuses mainly on two aspects of the geology of the Nazareth quadrangle: 1) the origin, depositional history, and stratigraphic significance of the Cambrian and Ordovician sedimentary rocks that underlie the Great Valley, and 2) the geologic structure and tectonic history of these rocks and their structural relationship to the Precambrian rocks of the Reading Prong. Its goal is to contribute to the understanding of the nature and history of geologic processes involved in the formation of the Appalachians in the eastern Pennsylvania-western New Jersey area. Although the study area is small compared to the geographic extend of even the middle Appalachians (Roanoke, Virginia, to Hudson River, New York), the rocks and structural features described herein extend far beyond the boundaries of the study area and conclusions based on the intensive study of such features therefore may have implications of broader regional significance.

Statement of the Problems

Most geologic studies in the general area of this study have concentrated on stratigraphic and structural aspects and have largely ignored the sedimentary history of the lower Paleozoic rocks themselves. This is unfortunate because these rocks contain at least a partial record of the beginnings and early geologic history of the Appalachians. Such history includes the heritage of the sediments, i.e., the position, lithologic nature, relief, climate, and tectonic character of the land area that supplied the sediments, the environment in which the sediments were deposited and the nature and kinds of processes that operated within it, and the post-depositional changes in the environment that led to alteration, lithification, and development of the rocks as we see them today. Systematic changes in rock properties that can be related to their source areas and depositional environments can reveal much about the history and direction of sea-level changes and the development of the Appalachian geosyncline, the deeply-subsiding elongate basin and sediment receptacle that is the basic element in the Appalachian tectonic framework.

The carbonate rocks pose some special problems because they largely reflect chemical, mechanical, and biological factors in the depositional basin. If they contain significant terrigenous debris, they may reflect source area conditions as well.

The dolomitic rocks are of particular interest in this study because they are typical examples of what for several decades has been called "the dolomite problem." The so-called "dolomite problem" is actually composed of several problems, some of which are well on the way to solution. Briefly summarized, part of the problem is that dolomites of great thickness, great lateral extent, and high purity occur throughout the geologic record, but none are forming on a comparable scale in modern carbonate depositional environments, and until the report of modern dolomite deposition in South Australia (Alderman and Skinner, 1957) none at all had been found in modern environments. Moreover, pure well-ordered, stoichiometric dolomite has not been synthesized under conditions that approach the temperature, pressure, and salinity of modern seas, in spite of the fact that dolomite should be a stable carbonate phase in the marine environment (Zen, 1960).

The scarceness of modern dolomite, its apparent stability in the marine environment, and the lack of success in synthesizing the pure mineral raise another question about the origin of certain dolomites that are common in the geologic record. Clearly, many dolomites originated by replacement of calcium carbonate sediments. The petrographic evidence is overwhelming. These rocks are coarse grained, commonly porous, contain fossils that unquestionably were composed originally of calcite or aragonite, and contain relict textures and structures that are typical of modern calcium carbonate sediments. On the other hand, many dolomites exhibit no vestige of a previous texture or structure, are uniformly very fine grained to aphanocrystalline, are strongly laminated, and are unfossiliferous. These dolomites have been considered by many to originate as primary chemical precipitates. Some isotopic studies, in addition to the inability to synthesize pure dolomite, suggest that such primary precipitation does not occur, however. In the Nazareth quadrangle both of the dolomite types described above are common.

The discovery of oil in carbonate rocks, including dolomite, and the discovery of modern dolomite forming in South Australia have generated a wealth of research activity on the nature and origin of these rocks. Dolomite now has been found forming in several localities in a variety of environments, and a much better interpretation of ancient limestones and dolomites now is possible.

Thus, the question to which the study of the lower Paleozoic sedimentary rocks in the Nazareth quadrangle is directed is this: what do these rocks reveal about their heritage, the environment in which the sediments were deposited, and the depositional processes that formed them, and of what significance are they in the interpretation of the Cambrian and Ordovician history of the Appalachians? The approach to this question involves both field observations and petrographic studies, and the results are incorporated into the systematic description of the stratigraphic units.

The second aspect of this study involves the structural relations in Nazareth quadrangle. Initially, the base of the Paleozoic rock sequence in the Great Valley was in depositional contact with the Precambrian rocks of the Reading Prong, and, to some extent, this relation still holds. However, the relations have been greatly complicated by subsequent tectonic events that deformed both groups of rocks and brought Precambrian rocks into contact with, or even on top of, rocks much younger than those at the base of the Paleozoic sequence.

The Paleozoic rocks have been deformed in at least one and perhaps as many as three (Taconic, Acadian, and Appalachian) orogenic episodes. The effects of each deformation are superimposed upon the structures that represent earlier events. The problem thus is both geometrical and historical. What is the structure of the rocks of the Great Valley? If more than one episode of deformation occurred, what are the individual and cumulative effects of each, and when did these episodes occur. By what processes did the structures originate and of what significance are they to the understanding of the origin and tectonic history of the Appalachians?

A related problem of continuing interest is the relation of the Precambrian rocks of the Prong to those of the Great Valley. Principally there are three poles of thought on the matter. The oldest is that the Precambrian rocks, as seen today, are deeply-rooted upthrown blocks of high angle normal or vertical faults.

Another is that the Precambrian rocks are part of a far-traveled overthrust plate that moved over the Paleozoic rocks on a very low angle (sub-horizontal) thrust plane. Both views are based mainly on reconnaissance and small scale mapping. Recent detailed studies by Avery A. Drake, Jr., of the United States Geological Survey, and his coworkers, including the writer, suggest that the Precambrian and Paleozoic rocks together are involved in one or more large recumbent folds of regional extent. If this is true, the Paleozoic structural and tectonic histories of these two groups of rocks are more intimately related than is implied by the two earlier ideas. This is a sweeping departure from earlier ideas and is of great significance in the interpretation of the nature and magnitude of the forces that shaped Appalachian structures, and to the economic geology of structurally and stratigraphically controlled rock and mineral deposits. Detailed mapping in the Nazareth quadrangle, in conjunction with geophysical studies and mapping by personnel of the United States Geological Survey and the Pennsylvania Geological Survey in this and nearby areas, should contribute to a better understanding of these regional structural relations and the tectonic forces and processes that were involved in their development.

Location

The Nazareth quadrangle occupies some 56 square miles between latitudes 40°37'30" and 40°45' north, and longitudes 75°15' and 75°22'30" west. The area is entirely within Northampton County in eastern Pennsylvania between the cities of Easton and Bethlehem and includes parts of both municipalities (fig. 2).

Physiography

The area is part of the Appalachian Highlands and includes the Reading Prong section of the New England province, and the Great Valley section, known locally as the Lehigh Valley, of the Ridge and Valley province.

The topography ranges from flat or gently rolling in the Great Valley to moderately hilly in the Reading Prong. The elevation of the divides in the Great Valley is about 400 feet where the bedrock is limestone or dolomite, but as much as 700 feet where the bedrock is shale. Summit elevations along the Reading Prong in this area range from 670 to 1016 feet. The maximum relief in the Nazareth quadrangle, from the highest point on top of Gaffney Hill in the southeasternmost part of the area to the lowest point along the Lehigh River, is 826 feet.

Pleistocene glaciers, thought to be pre-Wisconsin in age (Miller and others, 1939), covered most of the area north of the Lehigh River. However, aside from the very flat topography developed on patches of drift, there are no topographic forms in the area which can be recognized as being intrinsically glacial. The principal effect of glaciation was to subdue and to soften the profile of the preglacial topography.

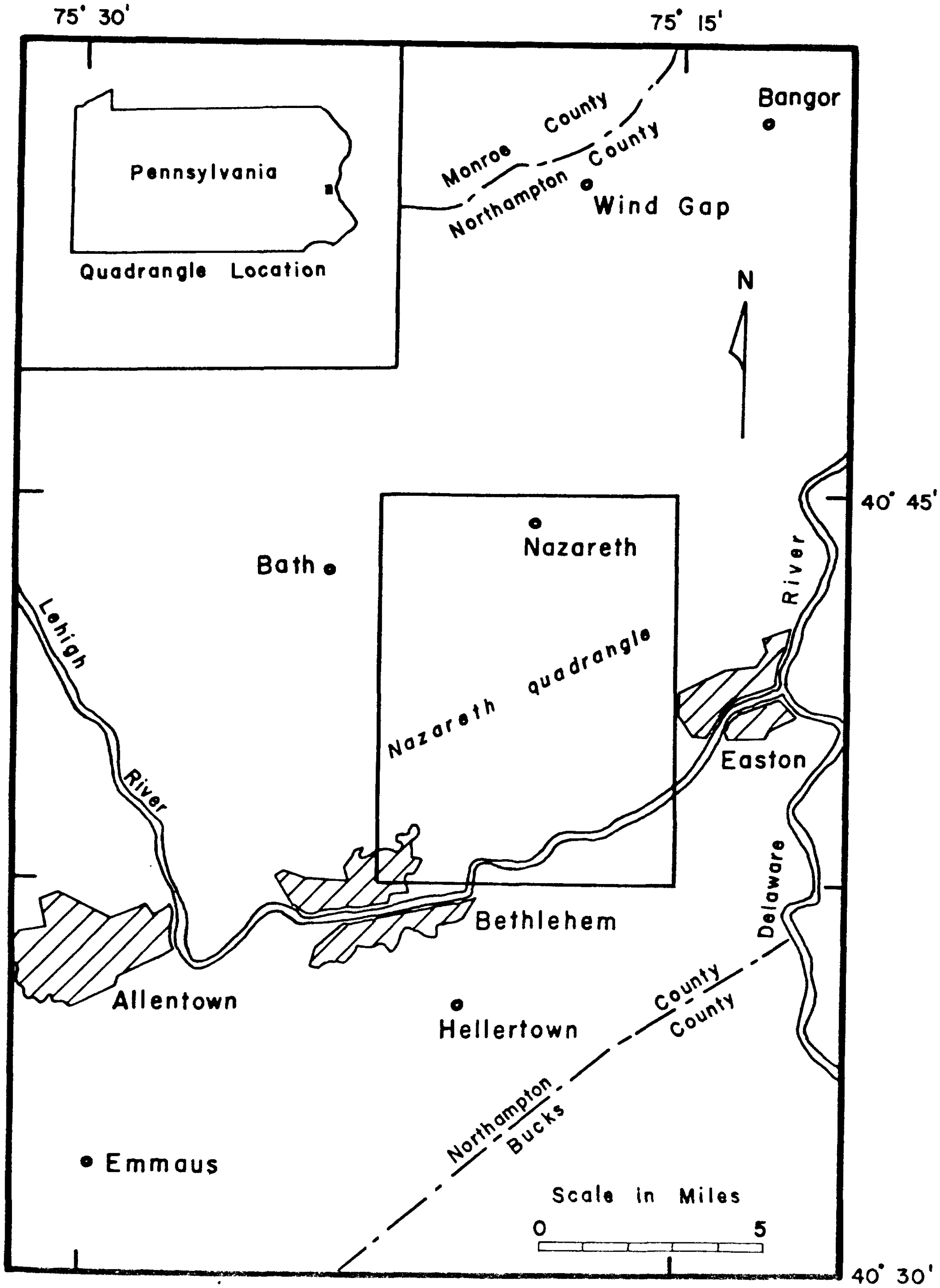


Figure 2.— Index map showing location of Nazareth quadrangle.

Previous Work

The area around Easton, Pennsylvania, in the Lehigh and Delaware valleys, has received a wealth of attention from geologists since the earliest days of geologic exploration in the United States. This is due in part to the important mineral resources of the area, principally iron ores, cement materials, and commercial slate, but also to the fascinating diversity of geologic features and problems in the area.

The first published maps and descriptions that include the area are those of Maclure (1809) and Finch (1824). Both described the "primitive" (Precambrian) crystalline rocks of the Reading Prong and the "transition" (Paleozoic) limestones of the Great Valley. Rogers (1858), directing the First Geological Survey of Pennsylvania, described the major rock units and structures, and included details on specific localities in the area.

The Second Geological Survey of Pennsylvania (Lesley and others, 1883, 1892) produced the first large scale systematic maps and descriptions of the area. The authors (Lesley and others, 1883) divided Northampton County, including the area of this report, into three geological regions which were termed the northern slate belt, the middle limestone belt, and the southern "syenite" (gneiss) belt. These regions plus the "Potsdam" sandstone (Cambrian) were also their geologic map units. They described in detail all aspects of the geology and geography of each region. The slate belt was thought to rest unconformably on the limestone belt, though it was admitted that no positive evidence for such a relation existed. The structural complexity of the limestone belt was fully appreciated, if not understood, as is evident from the statement of J. P. Lesley (in Prime, 1875) that

> . . . what seems so smooth and regular a surface conceals one of the most contorted, twisted, fractured, cleft, plicated, complicated, and even overturned set of subsoil rocks in the world.

The belts of Precambrian gneisses were considered to be anticlinal structures, asymmetrical and overturned to the northwest, which also included the overlying limestones.

B. L. Miller (1925) divided the "limestone belt" of earlier workers into three geologic map units. He separated the older, predominantly dolomitic rocks of the southern part of the belt from the younger argillaceous limestone (cement rock) and pure limestone (cement limestone) of the northern part of the belt. Precambrian rock units included graphitic limestone and undifferentiated gneisses. There was no further subdivision of the slate belt within the area of the present study, but isolated patches of limestone were mapped within the slates and shales a few miles to the west.

Miller (1925) included no structural cross sections in his report, and his discussion dismissed the structure as being too complex for satisfactory generalization. His map, however, suggests fold structures which include the Precambrian rocks. No faults or fold axes appear on the map, but the discussion indicates that folding and faulting are complex.

In a later report Miller and others (1939) further subdivided the various dolomitic rocks of the Great Valley and the gneisses of the Reading Prong. The general structure of the region was shown to consist of ridges of Precambrian crystalline rocks separated by narrow, steep-sided valleys. The valleys normally are floored with Paleozoic rocks that were down-faulted or down-folded, or both. The map patterns are complex, especially where Precambrian rocks are in contact with younger rocks. Faults, usually shown as very steep and normal, or vertical, were used to explain most aberrant contacts. Folds were recognized as generally asymmetrical to the northwest, but overturning was thought to be only a local complexity. Minor structures were ignored completely.

In 1957 the United States Geological Survey began a program of systematic mapping in the Delaware Valley, mostly in New Jersey, by Avery A. Drake, Jr., and others. Particular attention was given to differentiating the complexly-structured gneisses and granitic rocks of the Reading Prong, to refining the stratigraphy of the Great Valley, and to the orientation and significance of the minor structures. These studies have shown that overturning, far from being merely a local complexity as Miller (1939) thought, is in fact a regional phenomenon. Moreover, evidence gathered thus far strongly suggests a regional tectonic picture substantially different from that proffered by previous workers. The present study is an outgrowth and extension of this Geological Survey program.

Method of Study

The field work for this study was carried out during the summers of 1963 and 1964, the fall of 1965, and the spring of 1966. A total of about nine months was spent in mapping and related activity.

Field data were plotted on United States Geological Survey topographic map compilation sheets of the Nazareth quadrangle, and subsequently on the published edition. Recent (April, 1962) aerial photographs (approximate scale 1:20,000) were used for location of data stations in the field. Photo locations were then transferred to the topographic base map. Where photos did not permit accurate location, the Brunton compass was used for location by resection or pace-and-compass methods.

Approximately 350 hand specimens were collected in the field, and from these 125 thin sections were studied. In addition, studies were made of about 100 hand specimens and thin sections of related rocks outside of the Nazareth quadrangle. X-ray and chemical analyses were obtained where desirable.

Terminology

The classification and nomenclature of sedimentary rocks are by no means standard, as both are still somewhat in a state of development. Familiar terms commonly mean different things to different people, depending on what classification scheme is used.

In this study the nomenclature applied to detrital sedimentary rocks is essentially that of Krynine (1948), except that the term "claystone" is used for rocks in which at least half the grains are less than 0.004 mm in size, and which lack bedding fissility. The term shale commonly implies that a degree of lamination or bedding fissility is present, but such need not be the case in all pelitic rocks described in this study. The term shale will be applied only to pelitic rocks that exhibit bedding fissility.

Terminology for the thickness of stratification and parting units in sedimentary rocks is that of Ingram (1954). Cross-stratification is described according to the scheme of McKee and Wier (1953), which is cited where their terminology is used.

By comparison with detrital rocks, the classification of carbonate rocks is practically in the newborn stage. The recent quantum jump in the research effort on these rocks has generated not only more extensive knowledge and a deeper understanding of carbonate depositional processes, but also a plethora of new classification schemes and new, largely unfamiliar terminology. Most systems strongly reflect the interest, experience, and purposes of the authors, and most are largely untested outside the specific area in which and for which they were devised. Dolomotic rocks pose additional problems because they commonly have two, perhaps more, textural elements of different type and origin that must be considered. No single carbonate classification has won wide acceptance.

The classification adopted in this study is based on the pioneering system of R. C. Folk (1959), which was developed on rocks very similar to those in the area of this study, and which is descriptive of a broad range of marine limestones. Usage in this study differs slightly from that of Folk. Folk (1959) feels that 1 mm is a more meaningful boundary between the rudite and arenite classes of allochem textures than the traditional 2 mm boundary applied to terrigenous sediments. The 1 mm boundary is also considered a fundamental break in the size of calcite or dolomite crystals composing the rock. However, Folk (1959) presents no evidence to support these changes, and the traditional 2 mm boundary is retained herein. The grain size scales and nomenclature used in this study are presented in Table 1.

The treatment of dolomites is expanded somewhat in order to describe these rocks more thoroughly than would be done under the Folk system strictly applied. This simply means that the allochemical texture (oöids, intraclasts, etc.) will be specified as to both size and type, rather than type only. The complete dolomite rock name will consist of a) grain size of dolomite crystals (coarsely crystalline, etc.), b) size of allochems (medium, etc.) and c) a modifier specifying the allochem type (oölitic, etc.) or a main rock name that specifies the allochems, if they are intraclasts (dolarenite). Thus a dolomite composed of dolomite crystals between 1 and 2 mm in size and more than 25 percent oöids between 0.25 mm in size, would be termed a "very coarsely crystalline medium oölitic dolomite." A rock composed of dolomite crystals between 0.125 and 0.25 mm in size and intraclasts (relics) to 2 mm in size would be a "medium crystalline very coarse dolarenite."

Table 1. Size classification of textural and compositional elements of limestones and dolomites

mm	Orthochemical or Replacement	Allochemical Texture		mm
64	Extremely coarsely crystalline	Very coarse	Calcirudite (Dolorudite)	64
		Coarse		16
		Medium		4
2		Fine		2
1	Very coarsely crystalline	Very coarse	Calcarenite (Dolarenite)	1
	Coarsely crystalline	Coarse		0.5
0.25		Medium		0.25
	Medium crystalline	Fine		0.125
0.062		Very fine		0.062
	Finely crystalline	Coarse	Calcilutite (Dololutite)	0.031
0.016		Medium		0.016
	Very finely crystalline	Fine		0.008
0.004		Very fine		0.004
	Aphanocrystalline	Sublutite		

Algal stromatolites are termed as such, rather than "biolithmite," but crystal size is specified. The terms dolorudite, dolarenite, and dolutite are used where intraclasts constitute more than 25 percent of the allochems, and allochems compose more than 10 percent of the rock. Otherwise, where oöids or pellets constitute 25 percent or more of the allochems, the rock is termed an "oölitic" or "pellet" dolomite, as appropriate, with the size of the dominant allochem specified according to the size scale in Table 1. This usage is summarized in Table 2.

Table 2. Compositional classification of dolomites

<table>
<tr><td colspan="4"></td><td>Allochem relics >10 percent</td><td>Allochem relics <10 percent</td></tr>
<tr><td rowspan="4">Volumetric Allochem Composition</td><td colspan="3">Intraclasts >25 Percent</td><td>Dolorudite
Dolarenite
Dololutite</td><td rowspan="4">Dolomite (Specify crystal size)</td></tr>
<tr><td rowspan="3"><25 Percent Intraclasts</td><td colspan="2">Oöids >25 Percent</td><td>Oölitic Dolomite</td></tr>
<tr><td rowspan="2"><25 Percent Oöids</td><td>Pellets> Fossils</td><td>Pellet Dolomite</td></tr>
<tr><td>Fossils> Pellets</td><td>Biogenic Dolomite
(Specify fossils if possible)</td></tr>
</table>

Chapter II

PRECAMBRIAN ROCKS

Introduction

A great variety of older Precambrian high-grade metasedimentary, metavolcanic, and igneous rocks crop out in the Reading Prong and New Jersey Highlands. An apparently younger sequence of lower-grade metasedimentary and metavolcanic rocks occurs locally.

Early workers recognized the great diversity of lithologies but for mapping purposes they lumped them into a few broad and rather inclusive stratigraphic units. These comprise the Franklin Limestone (Wolff and Brooks, 1898), and the Pochuck, Losee, and Byram Gneisses (Spencer and others, 1908), all of which have been widely used throughout the Prong. In addition to these are the Pickering Gneiss (Miller, 1912) and the Moravian Heights Formation (Fraser, in Miller and others, 1939), which have been applied locally, mainly in eastern Pennsylvania. The Franklin Limestone includes mostly dolomitic marble with associated graphitic schists; the Pochuck Gneiss includes largely mafic gneisses in which hornblende, pyroxene, and biotite are abundant; the Losee Gneiss comprises intermediate gneissic rocks in which plagioclase feldspar is dominant; and the Byram Gneiss includes granitic gneisses in which potassium feldspar is abundant. The Pickering Gneiss was introduced in order to formally recognize certain gneisses associated with the Franklin Limestone in Piedmont rocks in Chester County, Pennsylvania. Bayley (1914, 1941) applied the name to gneissic rocks believed to be of sedimentary origin that are associated with the Franklin Limestone in the Reading Prong and Jersey Highlands. The Moravian Heights Formation comprises certain green, well jointed quartzo-feldspathic gneisses that characteristically contain streaks and layers of sericite or a serpentine-like mineral.

The Byram, Pochuck, and Losee Gneisses were believed by most to be of igneous origin (Spencer and others, 1908; Fraser, in Miller and others, 1939), and were thought to have intruded the Franklin, Pickering, and Moravian Heights, all of which were considered to be metasedimentary.

Recent workers (Hotz, 1952; Sims and Leonard, 1952; Sims, 1958; Drake, 1967a, 1967b; Buckwalter, 1959, 1962, 1963) have abandoned the system and terminology outlined above because rocks that are widely different in character and origin commonly fall within the same unit. Instead they have adopted a more objective mineralogic-lithologic approach that emphasizes the character and mappability of individual lithologies rather than presumed affinities with regional or subregional inclusive terminology. Such an approach is adopted in this study. However, the treatment remains at the essentially descriptive level because the area of Nazareth quadrangle that is underlain by Precambrian rocks is rather limited, as are outcrops within that area. Meaningful petrologic studies would require coverage of several quadrangles, at least. With but few exceptions detailed age relations between Precambrian rock units are generally unknown. In general, however, there are three distinct groups of rocks that make up the Precambrian terrane: an "older" high-grade metamorphic sequence and a "younger" low-grade sequence, both of which are metasedimentary or probably so, and intrusive rocks.

Older Precambrian Rocks

Amphibolite

Pods, lenses, and layers of amphibolite and pyroxene amphibolite occur sporadically in gneissic rocks of the older Precambrian sequence in the Nazareth quadrangle, and are widely distributed throughout the Reading Prong and Jersey Highlands. The lithology is more common than current mapping would suggest because many bodies are unmappable at 1:24,000.

The amphibolites in the Nazareth quadrangle are dark gray, greenish gray, and greenish black, medium to coarse grained equigranular to inequigranular rocks with crystalloblastic textures. Foliation typically is fair to good in the gneissic amphibolites, but is fair to poor in the coarser grained massive units. Foliation is manifest by the subparallel arrangement of tabular feldspars and hornblende or by alternations of feldspathic and mafic layers that result in a gneissic texture.

The essential minerals include xenoblastic plagioclase (andesine), hornblende, and clinopyroxene, with lesser orthopyroxene. Common accessories include sphene, apatite, magnetite, biotite, and quartz. Plagioclase typically constitutes between 35 and 65 percent of the rock. Hornblende ranges from 10 to about 50 percent, and forms polygonal grains that are intergrown with plagioclase and pyroxene. Clinopyroxene or orthopyroxene, or both, occur in quite variable proportions in almost all amphibolites, but clinopyroxene is the more common by far. The accessories collectively compose about 5 percent of the rock. Quartz in elongate composite grains with strongly undulose extinction was observed in one thin-section, and it constituted 5 percent of the section. Such an occurrence probably is related to amphibolite migmatite that has been mapped elsewhere in the Reading Prong by Sims (1958), Drake (1967a, 1967b), and others. A mode of a typical amphibolite appears in Table 3.

Quartzo-Feldspathic Gneisses

Gneisses described in this section characteristically are quite variable in composition. They typically occur together, however, and seem to represent a genetic group. Three variations seem to make up the bulk of this group. Minor occurrences of other types in the area were not separately mapped.

Potassium Feldspar Gneiss

Potassium feldspar gneiss is the most areally persistent of the quartzo-feldspathic gneisses in the Nazareth quadrangle. Rocks assigned to this unit occur at the western end of Chestnut Hill southeast of the village of Seipsville, at Pine Top and Camel Hump, two outliers of Precambrian rocks within the carbonate rocks of the

Table 3. Modes (volume percent) of typical Precambrian rocks from the Nazareth quadrangle

	1	2	3	4	5
Quartz	--	44.5	52.2	41.6	39.9
K-feldspar	--	47.2	22.8	0.5	--
Plagioclase	51.2	--	1.6	43.2	12.5
Microperthite	--	--	--	--	45.8
Sillimanite	--	--	14.3	--	--
Hornblende	26.7	--	--	--	1.8
Clinopyroxene	9.0	--	--	--	--
Orthopyroxene	2.7	--	--	--	--
Biotite	8.3	3.2	--	11.7	--
Sericite	--	1.4	7.2	--	--
Chlorite	0.5	--	--	0.5	tr
Epidote	0.4	--	--	--	--
Magnetite	0.5	1.6	0.9	1.3	tr
Apetite	0.7	tr	tr	--	tr
Sphene	--	--	--	--	tr
Zircon	--	tr	tr	--	tr
Garnet	--	1.1	--	1.2	--
Pyrite	--	1.0	1.0	tr	--
Total	100.0	100.0	100.0	100.0	100.0

Location of samples in Table 3 as follows:

1. Amphibolite. Collected from prominent knob on mountain side, 3,000 feet west of intersection of Apple Butter Road and Island Park Road, southern border of quadrangle.

2. Potassium feldspar gneiss. Outcrop from 500 feet elevation, southeast of Steel City.

3. Sillimanite gneiss. Sample from knob 750 feet north of Tumble Creek Road, southeast corner of quadrangle.

4. Biotite-quartz-plagioclase gneiss. Collected from north-facing slope, 1,750 feet north of Tumble Creek Road.

5. Microperthite alaskite. Collected from north facing slope, 750 feet south of Tumble Creek Road near its intersection with Island Park Road.

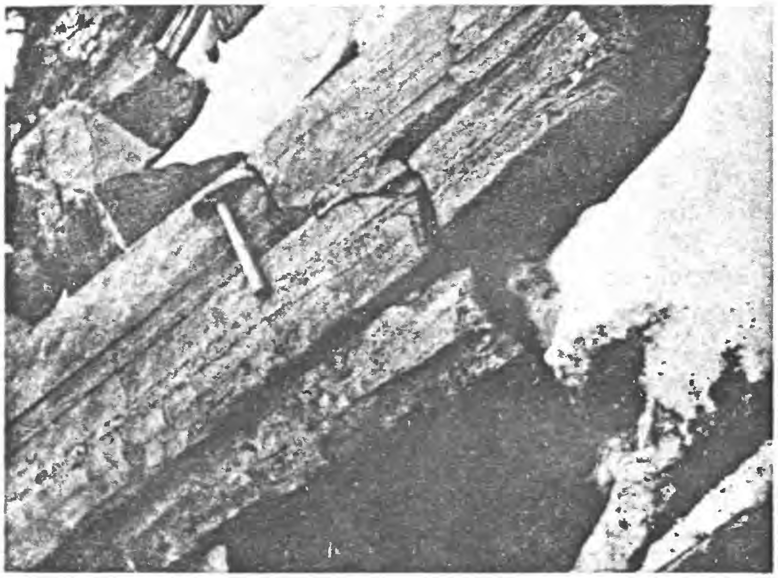
Lehigh Valley, and in the main Precambrian mass south of the Lehigh River. A typical outcrop is pictured in Figure 3.

Potassium feldspar gneiss is pinkish gray, light gray, and light greenish gray, fine to medium grained, and has a poor to very poor foliation. It is composed of xenoblastic quartz, microcline, and microperthite, with lesser oligoclase and biotite. Accessories include magnetite, which is abundant locally, apatite, garnet, zircon, hornblende and sphene. The texture is essentially granoblastic. Quartz, microcline, and microperthite occur in variable proportions, but together they typically compose 85 to 90 percent of the rock. Oligoclase and biotite rich phases are locally prominent variants, but are not mappable as such. Biotite typically is altered in varying degrees to chlorite, as is the sparse hornblende. Zircons typically are very round, which may suggest a metasedimentary origin for these rocks. In most areas rocks of this unit are permeated by microperthitic alaskite. A mode of typical potassium feldspar gneiss in the Nazareth quadrangle appears in Table 3.

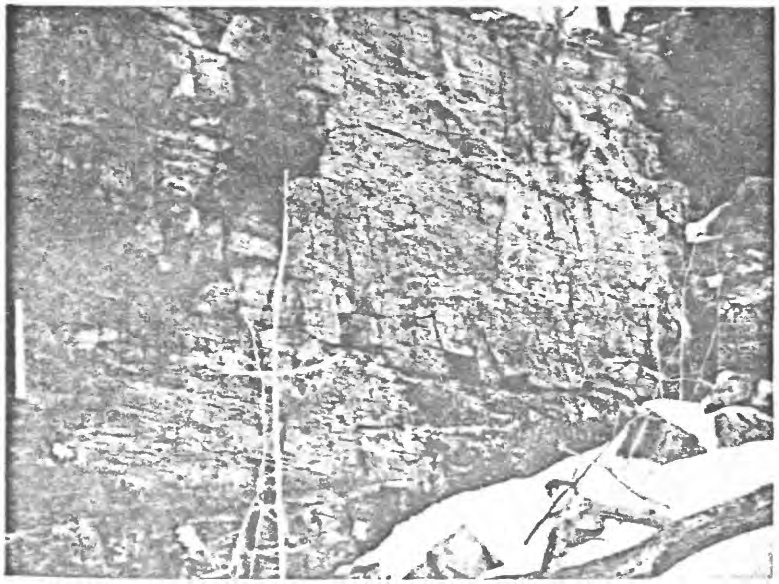
Sillimanite Gneiss

Quartzo-feldspathic gneiss characterized by the presence of sillimanite occurs in the southeasternmost part of the Nazareth quadrangle, where it is associated with potassium feldspar gneiss. Sillimanite gneiss is light greenish gray, light gray, and pinkish gray in color, fine to medium grained, and has a well-developed foliation. The texture is gneissic and inequigranular, but has a

Figure 3. Photographs showing outcrops of Precambrian rocks. _A_, Potassium feldspar gneiss, intersection of Apple Butter and Island Park roads. Note light color and well-defined foliation. _B_, Hornblende-clinopyroxene gneiss, Gaffney Hill.



A



B

strongly cataclastic fabric superimposed on it because the unit occurs within a prominent shear zone.

Xenoblastic quartz, microcline, microperthite, and oligoclase are the principal mineral constituents, with lesser sillimanite, biotite, magnetite, zircon, and apatite.

Sillimanite occurs as oriented needles and felty bundles in gneissic layers, and typically is altered in some degree to phyllosilicates, the specific nature of which is unknown. Oligoclase and perthitic plagioclase are strongly sericitized, and magnetite is altered in varying degrees to limonite. Quartz typically composes about 50 percent of the rock, potassium feldspar from 20 to 35 percent, and sillimanite from 10 to 25 percent. A mode of typical sillimanite gneiss is given in Table 3.

The composition and texture of the sillimanite gneiss, and its association with and similarity to potassium feldspar gneiss suggest that it may be simply a more aluminous phase of that unit. This gneiss is also typically permeated by veins, stringers, and layers of microperthitic alaskite.

Biotite-Quartz-Plagioclase Gneiss

Associated with the aforementioned potassium feldspar gneiss and sillimanite gneiss are small bodies of biotite-quartz-plagioclase gneiss. This gneiss is fine to medium grained and typically is light gray to medium gray or pale yellowish brown. A gneissic foliation is very well developed.

Xenoblastic quartz and oligoclase are the principal mineral constituents. Both compose 40 to 50 percent of the rock. Biotite constitutes 5 to 10 percent of the rock and is conspicuous. Commonly it shows some degree of chloritic alteration. Accessory minerals include garnet, magnetite, and rare hornblende. The texture is crystalloblastic and ranges from schistose in the biotite-rich layers to granoblastic in the felsic layers. A mode of typical biotite-quartz-plagioclase gneiss is given in Table 3.

Metamorphic Grade of the Older Precambrian Rocks

The gneissic rocks of the older Precambrian sequence are metamorphosed to the alamandine-amphibolite or granulite facies of regional metamorphism. The key mineral assemblages are quartz-sillimanite-alamandine-microcline (-plagioclase-biotite) of the quartzo-feldspathic gneisses, and hornblende-plagioclase (-diopside-hypersthene-quartz) of the amphibolites. The mineral assemblage of the quartzo-feldspathic gneisses is typical of originally pelitic and quartzo-feldspathic sediments in both of the above metamorphic facies (Turner and Verhoogen, 1960), but it does not permit a clear distinction between these facies. The assemblage present in the amphibolites, however, clearly belongs to the hornblende-granulite subfacies of the granulite facies, thus the entire body of older Precambrian rocks in the Nazareth quadrangle probably was metamorphosed under conditions commensurate with the granulite facies. The microperthitic and microantiperthitic feldspars that occur in the alaskites that have so thoroughly "soaked" these rocks are consistent with the conditions of metamorphism represented by the above facies as well.

Retrogressive metamorphism of the older sequence is represented by alteration of plagioclase and sillimanite to sericite and other phyllosilicates, and the alteration of hornblende and clinopyroxene to biotite, chlorite, or epidote.

Origin of the Older Precambrian Rocks

There is little firm evidence concerning the origin of these older rocks. All of the textures are crystalloblastic and no original fabrics are preserved. However, some inferences may be made from the mineralogy and associations of the rocks themselves.

One clue as to the ultimate origin of these rocks is the presence of marble (some dolomitic) interbedded and folded with these and similar gneissic rocks in several areas of the Reading Prong (Drake, 1967a,b; Epstein and others, 1967) and Jersey Highlands (Spencer and others, 1908; Hotz, 1952; Hague and others, 1956). There is no evidence at this time that all of the reported occurrences represent a single marble horizon, or that they represent several different horizons. Two distinct and separate layers are known to occur in the Franklin Furnace area of New Jersey (Hague and others, 1956), however. In any case, the marbles and associated gneissic rocks occur together in linear outcrop belts that can be traced along strike for miles. The gneisses are very heterogeneous within and between map units, and Drake (oral communication, 1970) has found what he believes to be feldspathic quartzites interlayered with the gneisses. Such field relations suggest, to the writer at least, that the older sequence is metasedimentary in origin. The metamorphic mineral assemblage of the quartzo-feldspathic gneisses is consistent with this view. That assemblage, as previously noted, is typical of pelitic and quartzo-feldspathic sediments that have undergone metamorphism to the almandine-amphibolite and granulite facies. In terms of bulk composition the interbedded amphibolites could represent basic rocks, perhaps lavas, or perhaps metamorphosed siliceous dolomites. Relic carbonate has been described from some amphibolites in the Jersey Highlands by Hotz (1952), Sims (1958), and Drake (1970), among others. This is a complex and problematical lithology, but it seems likely that at least some was derived from dolomite.

Younger Precambrian Metasedimentary Rocks

Distribution and Character

A body of apparently younger, less metamorphosed Precambrian rocks occurs in the Nazareth quadrangle along Bushkill Creek south of Bushkill (pl. 1). Similar rocks were mapped by Drake (1967a) in Chestnut Hill in the Easton quadrangle, and have been recognized elsewhere in the Reading Prong and Jersey Highlands (Spencer and others, 1908; Bayley, 1914), particularly on the northern side of these Precambrian masses. Rocks of this type in the Nazareth quadrangle occupy too small an area to allow mapping of individual lithologies at a scale of 1:24,000.

Included in this unit in the Nazareth quadrangle are arkose, arkosic conglomerate, quartzite, marble and serpentinized marble, and alaskitic pegmatite. Elsewhere in the Prong the above lithologies are associated with schistose rocks composed of talc, chlorite, sericite, quartz, and feldspar, hard purple slates, and metarhyolite.

Arkose is pale red purple with detrital quartz and feldspars of medium sand size. The feldspars include microcline, microperthite, and strongly sericitized plagioclase. Chlorite and rare metamorphic rock fragments are also present. Most detrital grains are subangular to subround. In the arkose conglomerate, which is especially prominent at this locality, subangular to round clastic pebbles of vein quartz are abundant, but otherwise the mineral composition is similar to that of the arkose. Some of these conglomerates contain appreciable (and conspicuous) disseminated pyrite. Some quartz-rich rocks are associated with the arkosic rocks, and these typically are vitreous, dark gray, highly fractured, and severely iron-stained. Some of these rocks contain up to 35 percent hematite. Many grain boundaries are strongly sutured and locally the original sedimentary fabric is completely destroyed. Blocks of altered (tremolite-diopside-talc) dolomitic marble, serpentine, and small exposures of a coarse alaskitic pegmatite occur in the Bushkill Creek locality, but their relation to the clastic rocks is not clear because of alluvial and glacial cover. This association with the metasedimentary clastic rocks is not unusual, however, Perhaps the pegmatite contributed to the alteration of the marble by contact metamorphism.

Metamorphic Grade and Origin

The clastic rocks that compose this younger Precambrian sequence in the Nazareth quadrangle very clearly represent a lower metamorphic grade than do rocks of the older sequence. Metamorphic minerals in the arkose and arkosic conglomerate include sericite and chlorite, which are characteristic of the quartz-albite-muscovite-chlorite subfacies of the greenschist facies (Turner and Verhoogen, 1960).

The age and relation of these metasedimentary rocks to each other or to the higher grade gneisses, previously described, is largely unknown because of poor exposure. Their lower metamorphic grade and location along one border (the northern) of these Precambrian masses suggest that they represent a younger sedimentary sequence that bears a consistent structural or stratigraphic relationship, ostensibly superjacent, to the older rock sequence.

Intrusive Rocks

Hornblende-Clinopyroxene Gneiss

The most widespread unit in the Precambrian terrane of the Nazareth quadrangle is a body of hornblende-clinopyroxene gneiss. It underlies Gaffney Hill and most of the southeast corner of the quadrangle, and parts of this body also occur in the adjacent Riegelsville and Easton quadrangles (Drake, 1967a, 1967b), and the Hellertown quadrangle. There are only a few outcrops, however, and most mapping must be done on the basis of float. A typical outcrop is pictured in Figure 3.

The rock seems to be quite variable in composition. Most is greenish gray, dark greenish gray, and greenish black, fine to medium grained, with a fairly well-developed gneissic foliation. The foliation results from planar concentrations of quartz and feldspars and mafic minerals. The texture is crystalloblastic and mostly equigranular, but some quartz and feldspar occurs poikiloblastically in larger grains of hornblende.

Quartz occurs in sutured xenoblastic grains with an abundance of vacuoles, and composes from 10 to 30 percent of the rock. Plagioclase is xenoblastic and is mostly extensively sericitized andesine. It composes between 30 and 40 percent of the rock. Hornblende is a pale green variety that is only slightly pleochroic from pale yellowish green to green. It constitutes about 50 percent of the rock. Up to 20 percent clinopyroxene may accompany the hornblende locally. Accessory minerals include magnetite and sphene which commonly form at the expense of the mafic minerals and garnet. A gneissic rock composed essentially of hornblende, clinopyroxene and epidote is locally prominent within this unit.

This gneiss appears to be folded and metamorphosed with the older Precambrian rock sequence. It is intruded by veins and thin layers of microperthite alaskite that affect the older complex also. The alaskite thus post-dates the intrusive activity represented by this hornblende-clinopyroxene gneiss.

Clinopyroxene-Quartz-Garnet Granofels

A distinctive body of clinopyroxene-quartz-garnet granofels surrounds much of the above-mentioned body of hornblende-clinopyroxene gneiss in the southeastern corner of the Nazareth quadrangle (pl. 1). The surrounding envelope of garnet granofels also occupies parts of the Riegelsville, Hellertown, and Easton quadrangles where they join the Nazareth quadrangle.

This rock is light brownish gray to dark greenish gray, fine to medium grained, and equigranular to inequigranular. Foliation is poor to nonexistent and, where present, is expressed as planar concentrations of quartz or pyroxene.

All mineral constituents in the granofels are quite variable in abundance. Some hand specimens are almost entirely garnet, whereas others may contain almost none. Fresh and weathered surfaces have the appearance of a garnet clotted gneiss.

Garnet is the most conspicuous constituent of this unit. It is reddish-brown and occurs as single small crystals or as irregularly distributed clots that range up to 6 inches in size. Its distribution is so variable that it is difficult to say precisely how much of the unit it composes. It is roughly estimated to constitute from 20 to 40 percent of the rock. Clinopyroxene occurs as stubby xenoblastic crystals that are in various degrees of alteration to magnetite and chlorite. Some is virtually unaltered, however. It makes up a highly variable proportion of the rock, estimated to range from 35 to 60 percent. Quartz composes from 25 to 50 percent of the granofels and occurs as irregularly shaped grains of fine to medium size. The quartz grains contain numerous inclusions of xenoblastic to idioblastic garnet that is similar in petrographic expression to other garnet present in the rock. Small amounts (5 to 10 percent) of strongly sericitized plagioclase are present, as are trace amounts of sphene and epidote.

The texture and composition of this granofels, and its spatial relation to the hornblende-clinopyroxene gneiss suggest that this body is an alteration zone of some sort, associated with the hornblende-clinopyroxene gneiss. It seems reasonable to suppose that the hornblende-clinopyroxene gneiss is a small metamorphosed intermediate to basic intrusion and that the pyroxene-garnet-quartz granofels is the metamorphosed hornfelsic alteration zone associated with that intrusion.

Microperthite Alaskite

A small lenticular body of microperthite alaskite occurs in the middle of the northern slope of Gaffney Hill, in the southeastern corner of the Nazareth quadrangle (pl. 1), and material of this composition permeates most of the older gneissic rocks in the area. It also has cut the hornblende-clinopyroxene gneiss that is itself believed to be intrusive. Small bodies of similar shape and composition apparently are common throughout the Reading Prong and Jersey Highlands (Hotz, 1953; Sims, 1958; Drake, 1967a, 1967b; Drake and others, 1967), and Drake (oral communication, 1970) reports that this type of granite becomes increasingly abundant in the Prong southwestward from the Delaware River.

The microperthite alaskite is very light gray, pinkish gray, and very pale orange to light brownish gray in color. It is characterized by an almost complete lack of mafic minerals, and is composed essentially of quartz and feldspar.

Microperthite is the principal mineral constituent and constitutes 35 to 55 percent of the alaskite. Quartz composes about 30 to 35 percent of the unit, and oligoclase makes up 15 to 25 percent of it. Microantiperthite is abundant locally, instead of microperthite, and alaskites of such composition have been mapped separately elsewhere in the Reading Prong by P. K. Sims (1958) and by A. A. Drake, Jr. (1967a, 1967b). The plagioclase feldspars are moderately to strongly altered to sericite. Hornblende, rare clinopyroxene, and biotite occur locally in variable quantities that do not total more than 5 percent, and trace amounts of zircon, apatite, and magnetite also occur in the alaskite. A mode of the typical alaskite in the Nazareth quadrangle is presented in Table 3.

The texture of the alaskite is variable and has both igneous and metamorphic affinities. In some specimens the texture is hypidiomorphic granular to pegmatitic and seems clearly to be of igneous origin. Where mafic minerals, especially hornblende, are more abundant, the alaskite has a poor to fair foliation.

The shape, composition, texture, and lithologic setting of this alaskite suggest that it is a syntectonic phacolithic intrusion into the metasedimentary or metavolcanic quartzo-feldspathic gneisses and the hornblende-clinopyroxene gneiss, hence is younger than these rocks. It seems likely that, on a regional scale, this alaskitic material is related in some way to larger microperthite- and microantiperthite-bearing granite plutons that intrude the same or similar quartzo-feldspathic gneisses like at numerous localities throughout the Reading Prong and New Jersey Highlands.

Metamorphic History of the Precambrian Rocks

Isotopic data on metamorphic events in the Reading Prong are rather sparse. The significance of the few studies that have been attempted is diminished by a lack of detailed mapping and poor correlation of lithologic units between study areas, by application of analytical results to the inclusive mega-units of early workers (Byram and Losee gneisses, etc.) and by the very complexity of the geology itself in relation to the scant analytical data. Almost no data are available for the Pennsylvania segment of the Reading Prong-Jersey Highlands Precambrian terrane, and the data in New Jersey are mostly auxiliary to studies directed mainly at the New York City and Manhattan Prong areas. There are, nevertheless, some reasonably consistent metamorphic and igneous events that are of broad regional significance.

Two distinct Precambrian metamorphic or heating events are suggested by available isotopic studies. A number of potassium-argon (K-Ar) and rubidium-strontium (Rb-Sr) determinations on biotites from gneissic rocks in the Jersey Highlands indicate a metamorphic or reheating event at about 835 ± 35 million years (m.y.) ago (Long and Kulp, 1962). This apparent age is consistent with similar determinations on biotites from the Storm King Granite and Canada Hill Gneiss (Tilton and others, 1960; Long and Kulp, 1962), both of which occur at Bear Mountain, New York, and both of which are part of the Highlands extension into New York State. Three uranium-lead (U-Pb) determinations on uraninite in marble (Franklin Limestone?) along the Delaware River, near Phillipsburg, New Jersey, give apparent ages that range from 825 to 915 m.y., and monazite in Losee Gneiss at Chester, New Jersey, is reported to yield a U-Pb age of 720 m.y. (Long, Cobb, and Kulp, 1959). Thus, the event at about 835 m.y. seems well-established throughout the New Jersey-New York Highlands. This event also is confirmed in the Reading Prong of Pennsylvania by a biotite K-Ar apparent age of 820 m.y. (Long and Kulp, 1962), from gneiss near Wernersville, at the southwestern extremity of the Prong.

A distinctly older event at about 1150 m.y. ago is indicated by concordant U-Pb apparent ages of zircon from the Canada Hill Gneiss at Bear Mountain, New York. The concordant ages include 1140 m.y., 1150 m.y., 1170 m.y., and 1030 m.y., determined respectively from Pb^{206}/U^{238}, Pb^{207}/U^{235}, Pb^{207}/Pb^{206}, and Pb^{208}/Th^{232} (Tilton and others, 1960). The Canada Hill Gneiss is the host rock to the Storm King Granite (Lowe, 1950), and its relationship and correlation with gneissic rocks southwestward in the New Jersey-Pennsylvania parts of the chain are not clear. However, brief petrographic descriptions (Lowe, 1950; Long and Kulp, 1962) suggest that the Canada Hill Gneiss may be equivalent to what has been mapped as Losee gneiss by Hague and others (1956), as oligoclase-quartz-biotite gneiss by Sims (1958), and as oligoclase quartz gneiss by Drake (1967a,b), all of which seem to be essentially equivalent units in the older Precambrian sequence. The description of Sims' (1958) oligoclase-quartz-biotite gneiss especially is suggestive of the Canada Hill Gneiss. It should be noted that the K-Ar apparent age of biotite in the Canada Hill Gneiss, as previously mentioned, belongs to the 835 m.y. event.

Thus metamorphic or reheating events at about 1150 and 835 m.y. ago seem well established. The event 1150 m.y. ago represents the original metamorphism or crystallization of rocks in the New Jersey-New York Highlands, and presumably in the Reading Prong. A discrete period of reheating and metamorphism occurred about 835 m.y. ago. It is not clear whether the biotites on which the 835 m.y. event is based actually crystallized in that event or whether they first crystallized 1150 m.y. ago and were subsequently altered by heating to reflect the 835 m.y. event.

In terms of Precambrian rocks in the Nazareth quadrangle, the older sequence, comprising the amphibolites and quartzo-feldspathic gneisses, apparently first recrystallized about 1150 m.y. ago. Subsequently these rocks were intruded by material now represented by hornblende-clinopyroxene gneiss which formed an alteration zone now represented by the clinopyroxene-quartz-garnet granofels. Whether this intrusion dates from the 835 m.y. event is uncertain. All of these rocks later were intruded, permeated, or cut by microperthitic granitic rocks, presumably during the 835 m.y. event. At that time the isotopic composition of biotites in older rocks was altered to reflect this event, new biotites may have formed, and uraninite and monazite were formed or recrystallized in some rocks. The sedimentary and volcanic rocks of the younger sequence evidently post-date the 835 m.y. event because they have undergone only incipient low-grade metamorphism, despite their close association with granulite facies rocks. Also, the texture and many of the detrital minerals in the younger sequence would readily have been altered under the metamorphic conditions represented by the 835 m.y. event.

Chapter III

STRATIGRAPHY AND PETROLOGY OF PALEOZIC AND CENEZOIC ROCKS

Cambrian System

Hardyston Quartzite

Introduction

The Hardyston Quartzite is about 100 feet thick and forms the base of the Paleozoic section in eastern Pennsylvania and western New Jersey, where it rests unconformably on Precambrian crystalline rocks of the Reading Prong. It has, of course, been recognized and mapped since the beginning of geologic work in the area, but all the early work and much of that done more recently was concerned with problems of stratigraphic definition and mapping and the resultant structural and tectonic implications. Little attention has been given to the Hardyston as a unit of sedimentation that records important aspects of its history not normally revealed simply by mapping. Most of the following data were obtained from exposures in the Nazareth quadrangle, Pennsylvania (fig. 4), and the quadrangles immediately east and south (Easton, Bloomsbury, Hellertown, and Riegelsville), but important localities in other quadrangles also were studied.

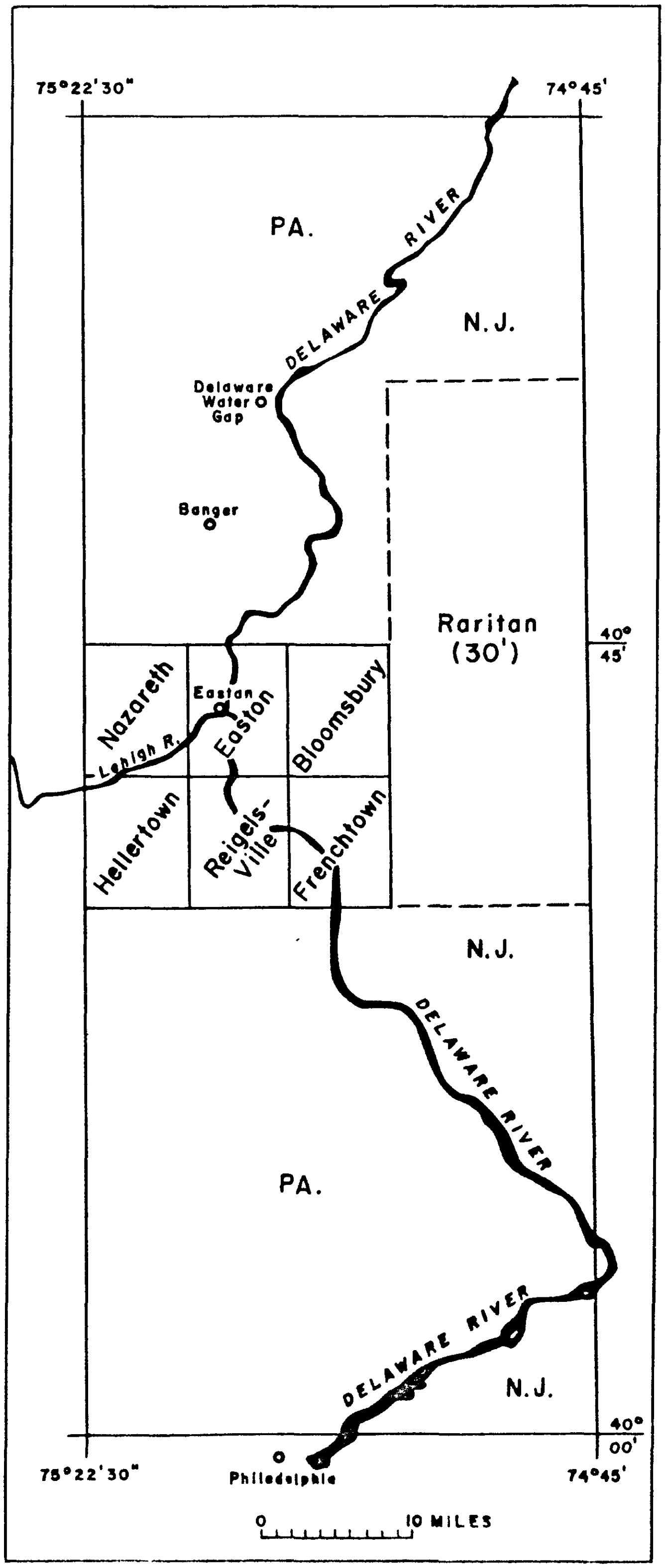


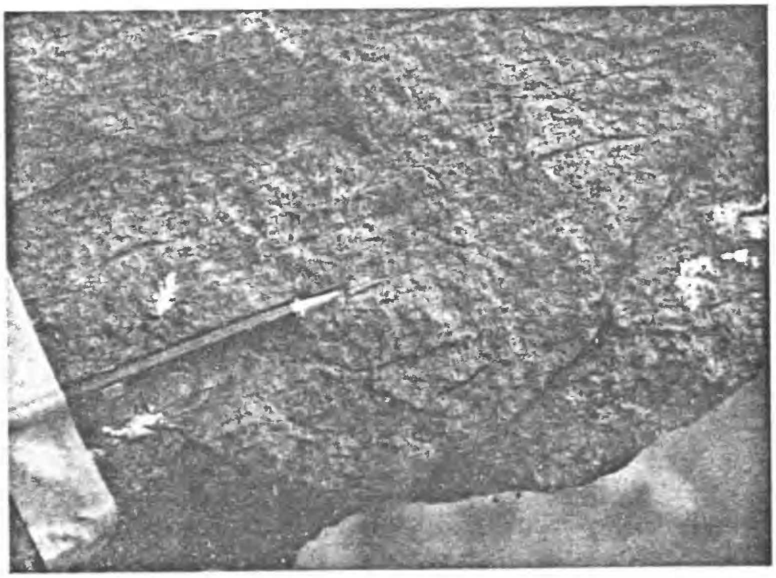
Figure 4. Index map of quadrangles in the vicinity of Nazareth.

Lithology

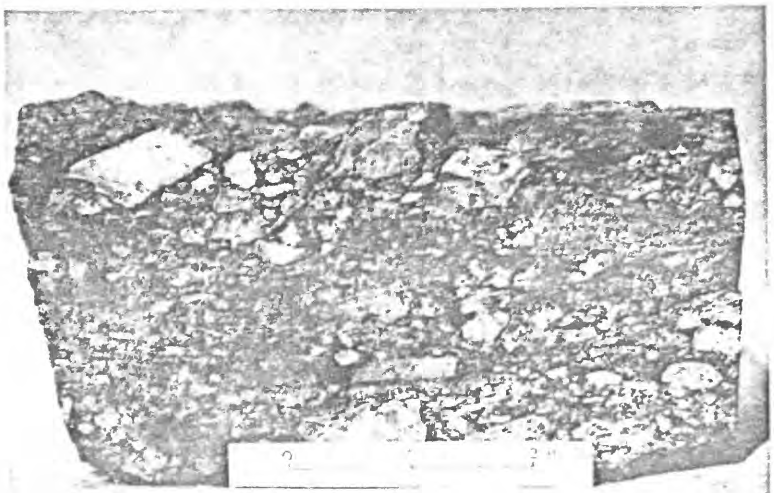
The Hardyston Quartzite contains a rather varied assemblage of lithic types, principally feldspathic sandstone, arkose, and orthoquartzite, with lesser quartz-pebble conglomerate, silty shale, and, locally, jasper. Carbonate rocks occur in the Hardyston in New Jersey. There is a broad textural gradation within the Hardyston; conglomerate and coarse sandstone are more common at or near the base and shale is more common at or near the top. These units are not generally mappable as stratigraphic subdivisions, however.

Arkose and feldspathic sandstone are light gray to very light gray and orthoquartzite is grayish orange-pink; all weather to moderate brown or moderate yellowish brown. Conglomerates are composed largely of iron-stained light to dark gray granules and pebbles of quartz and feldspar in a dark, poorly sorted, argillaceous matrix of finer quartz and feldspar (fig. 5). The granules and pebbles range in maximum diameter from 2 mm to 100 mm but most are less than 50 mm. They are subangular to subround, and the larger ones are platy to elongate. Shales are pale red or pale yellowish-brown, weather moderate yellowish brown, and are silty and sandy. In all rock types, quartz grains are vitreous light to dark gray, but vitreous blue quartz is abundant locally. Pyrite is commonly a conspicuous constituent; it occurs as a fine "dust" (0.1 mm diam.) widely disseminated throughout individual hand specimens, or as larger (0.2 mm-0.5 mm) blebs and cubes.

Figure 5. Photographs of Hardyston Quartzite. A, *Scolithus* tubes in orthoquartzite. Pencil points to well-developed tube emerging from plane of outcrop. B, Typical conglomerate from lower part of Hardyston. Large clasts are potassium feldspar.



A



B

Stratification in sandstone and conglomerate typically ranges in thickness from 5 cm to 1.5 m or more, and in small exposures may be very difficult to locate. Stratification in all Hardyston rock types is irregular in thickness and rather discontinuous laterally, especially in the lower part. Cross-strata, small scale and planar (McKee and Wier, 1953), are present locally, but the relative paucity of outcrops and the complexity of folding preclude meaningful statistical study of these features.

Name and Distribution

The Hardyston Quartzite was named for Hardistonville (Wolff and Brooks, 1898), a village near Franklin Furnace, Sussex County, New Jersey, where it is particularly well exposed. The name was shortened to Hardiston (Kümmel and Weller, 1901), the township name, which later was changed to Hardyston.

Formerly, the unit has been designated as Formation I of the lower Secondary rocks (Rogers, 1838), the Primal White Sandstone (Rogers, 1858), and the Potsdam sandstone (Prime, 1883). The discovery that the Potsdam sandstone in New York State (Van Ingen, 1902; Weller, 1903) was much younger than rocks called Potsdam in New Jersey and Pennsylvania led to the abandonment of the term in the latter areas.

The Hardyston Quartzite commonly is closely associated with Precambrian rocks, for both stratigraphic and structural reasons. Mapping in the area has shown that even where Precambrian rocks are allochthonous and believed to be far-traveled, the Hardyston usually is carried along. The same is rarely true for the younger carbonate rocks that overlie the Hardyston. Thus, Hardyston is exposed or present as float on one or both flanks of most of the Precambrian ridges in the area. Figure 6 shows the regional distribution of the Hardyston in easternmost Pennsylvania and northwestern New Jersey.

Thickness

There is much uncertainty concerning the thickness of the Hardyston in eastern Pennsylvania and western New Jersey; outcrops are scarce, long sections are nonexistent, and contacts are commonly faulted.

In the Delaware Valley the Hardyston is reported to be as much as 200 feet thick (Johnson and Willard, 1957). Miller (1939) reported a range of 0-200 feet in the Lehigh Valley. Mapping and construction of geologic cross-sections by Drake (1967a; Drake and others, 1967) and by the writer support a thickness of no more than 100 feet in these quadrangles. A section 45 feet thick was measured at low water in the Delaware River in the Riegelsville quadrangle (Drake, 1969).

In many areas the Hardyston locally is missing from its stratigraphic position. In some of these cases faulting clearly is responsible, but in others there is insufficient evidence or exposure to completely rule out non-deposition as a factor.

Regionally, the Hardyston thickens considerably southwestward along the Reading Prong. Buckwalter (1963) reports a thickness of

Figure 6. Generalized bedrock geologic map showing distribution of Hardyston Quartzite in part of eastern Pennsylvania and western New Jersey (in the quadrangles outlined in fig. 1). Data compiled from: Drake, 1967a,b; Drake and others, 1961, 1967; Miller and others, 1939, 1941; Willard and others, 1959 (Hellertown quadrangle); and J. M. Aaron, unpub. data (Nazareth quadrangle).

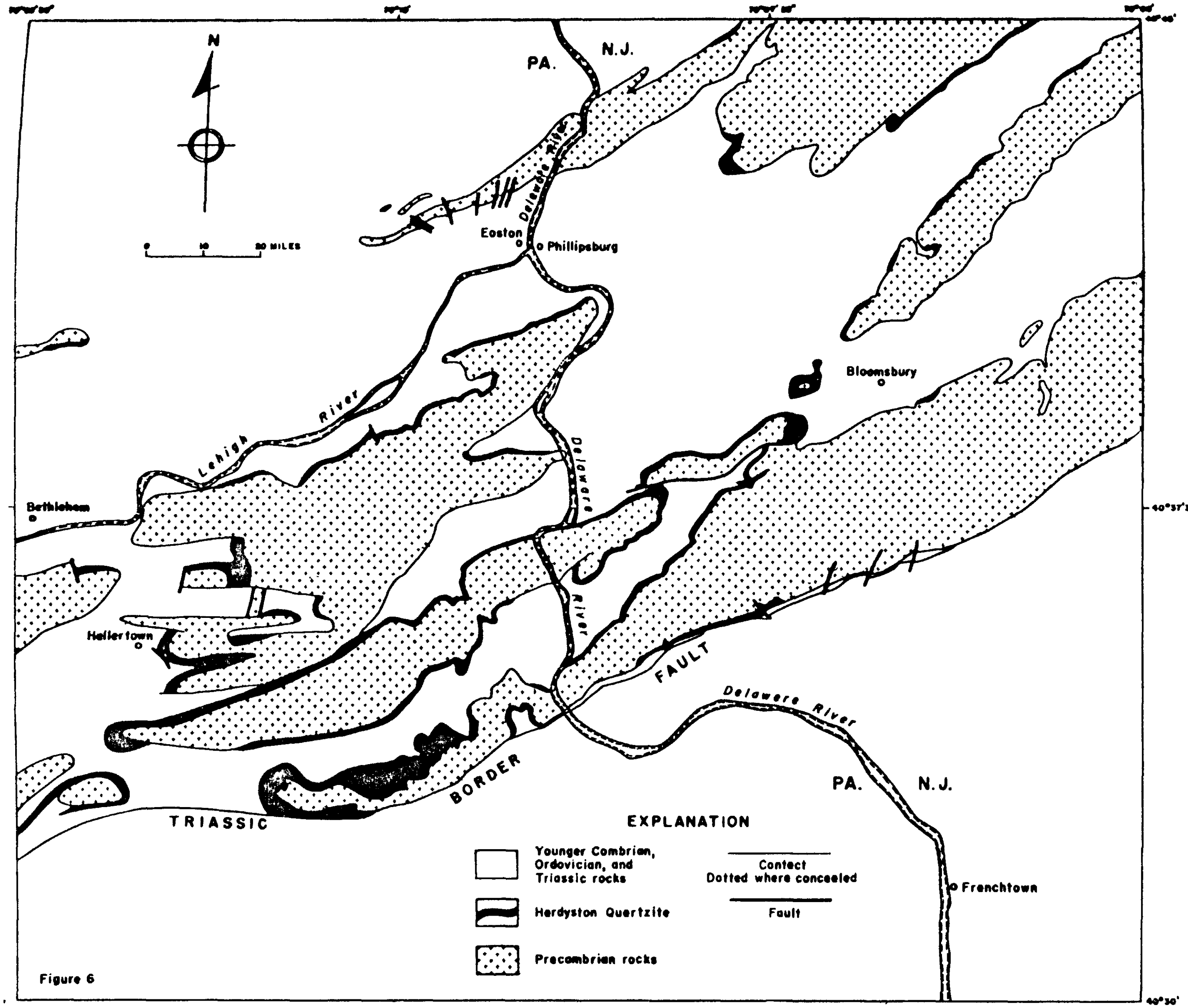

N
0 10 20 MILES
PA.
N.J.
Delaware River
Easton
Phillipsburg
Bloomsbury
Lehigh River
Bethlehem
Hellertown
Delaware River
FAULT
BORDER
TRIASSIC
Delaware River
PA.
N.J.
Frenchtown
40°37'30"
40°30'
EXPLANATION
Younger Cambrian, Ordovician, and Triassic rocks
Hardyston Quartzite
Precambrian rocks
Contact
Dotted where concealed
Fault

Figure 6

600-800 feet in the Womelsdorf quadrangle, at the extreme southwest end of the Prong.

Stratigraphic Relations

The Hardyston Quartzite uncomformably overlies Precambrian crystalline rocks. The contact is rarely exposed, and its location must usually be inferred from float.

Locally, in this area, the Hardyston-Precambrian contact is marked by a deposit of light-green cryptocrystalline material that has been called pinite (Miller and others, 1939; Virgin, 1956), a fine mixture of muscovite with some serpentine or chlorite mineral and iron oxides. This material has been interpreted as a metamorphosed remnant of a residual soil developed on underlying gneiss (Miller and others, 1939; Virgin, 1956). An alternative interpretation, favored by the writer, is that the so-called pinite is the result of shearing and mylonitization of argillaceous sediments at the Hardyton-Precambrian interface. Such an interpretation would imply that differential movement has taken place at the contact, but does not necessarily mean that the contact is faulted. Such shearing could result from folding and should be most intense where variations in relative competence are greatest.

The contact with the superjacent Leithsville Formation is transitional through a series of silty and shaly beds. This contact is poorly exposed and commonly faulted. As with the Precambrian contact, the geologist must rely on his regional experience to determine its location as nearly as possible.

Age

No fossils of conclusive chronologic significance have been found in what is currently mapped as Hardyston in eastern Pennsylvania. The only fossil in the area is Scolithus linearis (fig. 5), found at a few localities in the Lehigh Valley. However the taxonomic history of Scolithus is so diverse and its paleontologic significance so obscure that any interpretations based on it are highly questionable.

Olenellus thompsoni has been found in considerable numbers in calcareous beds of upper Hardyston at three localities in the Raritan quadrangle (fig. 4), New Jersey (Weller, 1903; Bayley and others, 1914). On this basis the Hardyston in eastern Pennsylvania is considered to be Early Cambrian in age.

Correlation

In eastern Pennsylvania and adjacent New Jersey the Hardyston directly overlies the Precambrian igneous and metamorphic complex (tab. 4). It would appear, therefore, to correlate physically with the Hellam Conglomerate Member of the Chickies Quartzite of southeastern Pennsylvania, with the Weverton Sandstone and Loudoun Formation of south-central Pennsylvania and adjacent Maryland and Virginia, and with the Poughquag Quartzite of southeastern New York. All these units are basal clastics in roughly equivalent stratigraphic sections.

In contrast to this apparent physical correlation is the location of the lowest Olenellus horizon within the stratigraphic section at the above localities. This horizon is in the upper part

Table 4. *Stratigraphy and correlation of Cambrian rocks in the Appalachian Valley, Pennsylvania, New Jersey, and New York.*

Age	Cumberland Valley, Pa. (Field Conf. Pa. Geol., 1966) Autochthonous	Lebanon Valley, Pa. (Field Conf. Pa. Geol., 1966) Allochthonous	Nazareth quadrangle, Lehigh Valley, Pa., Delaware Valley, Pa.-N.J. (Drake, 1967 a, b) Allochthonous	Kittatinny Valley, N. J. (Lewis and Kümmel, 1940) Allochthonous	Eastern New York (Broughton and others, 1962) Allochthonous (?)
Late Cambrian	Conococheague Group 2,150 feet Elbrook Formation 3,000 feet	Conococheague Group 2,000 ft. Buffalo Springs Formation 1,000 feet	Allentown Dolomite 1,700 feet	Kittatinny Limestone 2,500-3,000 feet (thru Early Ordovician)	Stockbridge Group 4,000 feet (thru Early Ordovician)
Middle Cambrian	Waynesboro Formation 1,000 feet	Leithsville Formation 1,000 feet	Leithsville Formation 1,000 feet		
Early Cambrian	Tomstown Formation 1,000-2,000 feet Antietam Sandstone 500-800 feet Harpers Formation 2,7 0 feet Weverton Sandstone 1,250 feet	Hardyston Formation 250-600 feet	Hardyston Quartzite 100 feet	Hardyston Quartzite 5-200 feet	Poughquag Orthoquartzite 250 feet
Precambrian	Volcanic rock	Gneiss	Gneiss	Gneiss	Gneiss

of the Hardyston and Poughquag in New Jersey and New York, respectively, but southwestward to central Pennsylvania, Maryland, and northern Virginia, it is progressively 1,200 to 2,600 feet stratigraphically higher, in the Antietam Quartzite. Paleontologically, therefore, the upper part of the Hardyston would appear to be correlative with the Antietam.

Petrography

Mineral Composition

Modes, based on detailed examination of thin sections of representative Hardyston lithic types, are presented in Table 5. Each mode represents approximately 400 counts. The total percentage of accessories is based on the point count, but the variety and relative abundance of these species is based on estimates by visual scan of the entire section and by examination of heavy-mineral separates from crushed rock.

Arkose and orthoquartzite are the end-member rock-types and all gradations exist between these rock-types, which have much the same character over the entire area.

Quartz is the most abundant mineral in all Hardyston rock types, constituting from 33 to 80 percent of the rock (tab. 5). Most quartz is sand size and angular to subangular. Well-rounded quartz grains are rare.

Microscopic examination of thin sections reveals that almost without exception quartz in the Hardyston occurs as single grains with strongly undulose extinction (fig. 7). The rare exceptions are

Table 5. *Modes* (volume percent) of typical Hardyston rock types

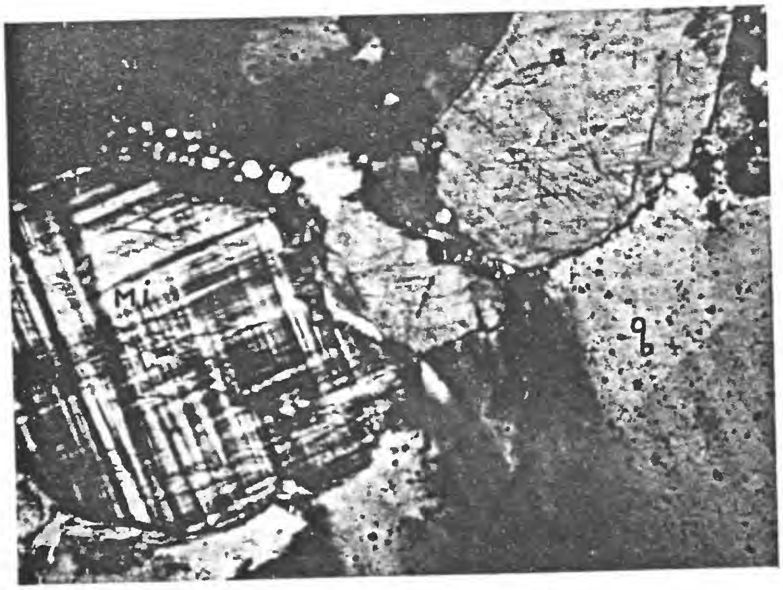
	Arkose		Arkosic Sandstone		Orthoquartzite	
Field Number	166-21	B-0-7-2	152-9	166-66	R-K3-32-C	166-48
Quartz	56.4	33.0	75.3	60.4	73.3	78.5
Potassium feldspar	20.4	15.0	10.3	12.0	3.3	1.0
Perthitic feldspars (all varieties)	4.8	16.7	4.7	4.8	--	1.3
Plagioclase	--	0.5	--	--	--	--
Silica cement	11.7	20.6	6.2	6.2	19.0	13.4
Clay and micromica	4.3	13.7	1.6	8.6	3.4	2.5
Accessories	2.4	0.5	1.9	8.0	1.0	3.3
Zircon	C	tr	tr	tr	tr	--
Monazite	C	--	C	C	tr	tr
Sphene	tr	--	tr	tr	--	--
Tourmaline	tr	--	C	C	--	--
Garnet	--	tr	--	--	--	--
Rutile	--	--	tr	--	--	--
Topaz	--	--	tr	--	--	--
Hornblende	--	tr	--	--	--	--
Pyrite	tr	--	tr	C	tr	--
Magnetite	A	A	A	A	A	C
Limonite	C	C	C	C	--	C

Location of samples in Table 5 and explanation of symbols are as follows:

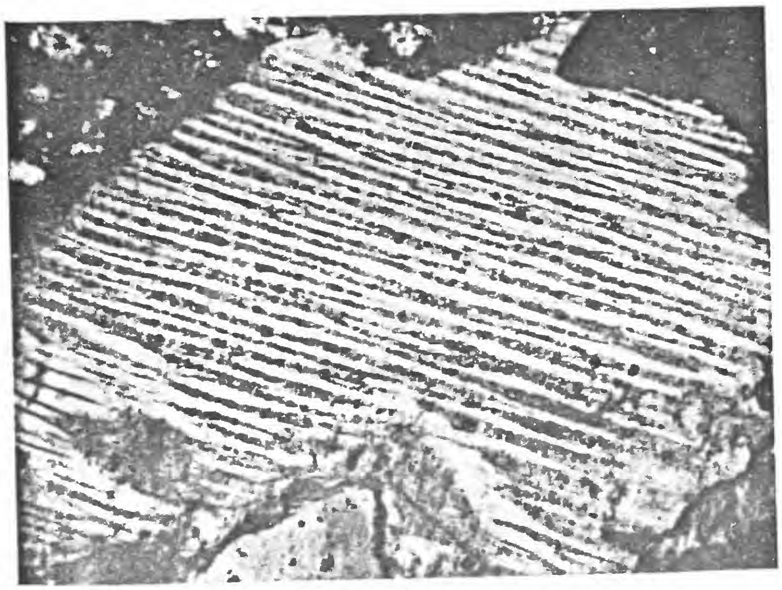
Nazareth quadrangle: 166-21, 152-9, 166-66, 166-48
Bloomsbury quadrangle: B-0-7-2 (collected by A. A. Drake, Jr.)
Riegelsville quadrangle: R-K3-32-C (collected by A. A. Drake, Jr.)

Visual estimates of relative abundance of species in the detrital accessory fraction are limited as follows: A, 50 percent or greater; C, 10 to 50 percent; tr, <10 percent.

Figure 7. Photomicrographs showing some petrographic features of the Hardyston Quartzite. A, Arkose with quartz showing undulose extinction (q) and microcline (mi). Crossed polarizers, X100. B, Differentially weathered detrital mesoperthite grain. Lighter lamellae are potassium feldspar, darker are altered sodic feldspar. Crossed polarizers, X100.



A



B

composite grains with the same type of extinction. In terms of microscopic morphology, the quartz in the Hardyston in no way differs from that of the several quartz-bearing gneisses and metamorphosed granitic rocks that constitute a major part of the Reading Prong in this area.

Although the morphology of quartz types by itself is of limited usefulness in provenance studies of sedimentary rocks, the strong predominance of a single type in the Hardyston at least suggests that the bulk may have had a common origin in high grade metamorphic rocks similar to those of the Prong. This suggestion is reinforced by the generally angular nature of most quartz, and by other evidence to be introduced below.

Feldspar is common to abundant in most of the Hardyston. The most abundant variety is potassium feldspar, mostly microcline with prominent grid twinning (fig. 7). Feldspars having microperthitic, mesoperthitic, or microantiperthitic structure are also common, and in some areas these collectively constitute the principal feldspar in the rock (for example B-0-7-2, Table 5). Plagioclase, chiefly oligoclase, is rare.

All of the above minerals are common in the Precambrian rocks presently exposed in the Reading Prong. Surprisingly, however, plagioclase is much less abundant in the Hardyston than one might expect if the Reading Prong as it stands today were the ultimate source of the detritus, as plagioclase, chiefly albite-oligoclase-andesine, is quite abundant there. The dominant feldspars in the Hardyston are typical only of a small fraction of rocks in the

Reading Prong. The predominance of the potassium feldspars must be due either to the relative unavailability of the plagioclase-bearing gneisses to erosion and transport, an unlikely possibility, or to the superior resistance and stability of the potassium feldspars. Differentiation of the latter sort is considered quite likely as there are clear examples of it in the coastal plain of New Jersey (J. P. Owens, oral communication, 1967). The possibility that the Reading Prong is not the source of the Hardyston detritus is considered highly unlikely in the light of other data to be presented.

The feldspar, like the quartz, is chiefly angular to subangular and most is of sand size. The feldspar in any given specimen tends to be in the same size range as the quartz in the same specimen, with the possible exception of some of the conglomerates. In these rocks feldspars may be very much larger than the quartz (fig. 5). Such rocks probably are of very local derivation from the coarser gneisses or granitic pegmatites in which feldspars may exceed 50 mm in size.

The feldspars are fresh to moderately altered, but on the whole remarkably fresh. Alteration of microcline consists of patchy areas of dusty brown cloudiness or turbidity, which conventionally is ascribed to kaolinitization, but which also may be due to vacuolization or bubble formation (Folk, 1961). In no case, however, are microline grains reduced to a mass of micromica or clay. The alteration is always of minor volumetric importance. Perthitic feldspars are altered differentially according to the composition of the exsolution lamellae (fig. 7). Sodic lamellae are weakly to moderately altered to sericite; the potassic lamellae are quite fresh or are altered weakly in the fashion described for microcline. Plagioclase is weakly to moderately altered to sericite.

Clay and micromica are minor constituents of the Hardyston, constituting less than 5 percent of the sandstones and about 50 percent of the shales. These minerals are predominantly sericite-illite, presumably derived by weathering of minerals, chiefly feldspars, in the source area, and by diagenetic alteration of feldspars after deposition. It is not clear which process has been more important in producing the clay minerals, but the relatively unaltered nature of most feldspars in thin sections suggest that diagenesis has been the less important.

Magnetite is the most abundant heavy accessory mineral; it composes from 35 to 95 percent (by volume) of the heavy fraction, but averages about 50 percent of that fraction. Limonite, used here for poorly crystalline hydrated iron oxides of uncertain identity, is commonly associated with magnetite. Both occur as irregularly shaped grains and rare euhedral and subhedral crystals in the finer fraction of the sandstones and conglomerates. Limonite also occurs, though rarely, as very round opaque grains from 0.5 to 1 mm in diameter, and both the enclosing sand grains and the round limonite grains are very well sorted. These could possibly be pseudomorphs after glauconite as glauconite readily oxidizes under subaerial conditions. The magnetite is detrital, but much of the limonite is authigenic.

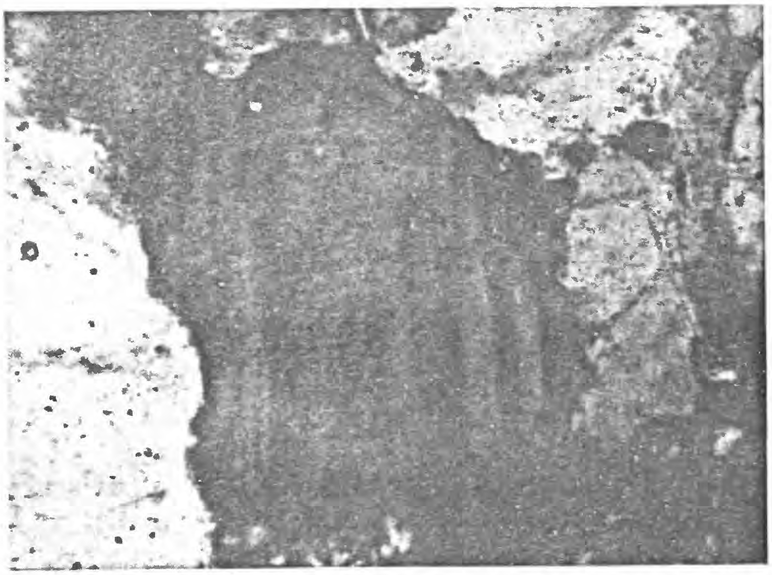
Pyrite is present in conspicuous amounts in some Hardyston sandstones but volumetrically it is a minor constituent. It occurs as cubes and irregularly shaped grains widely scattered throughout the sediment. Undoubtedly it is authigenic in origin.

Monazite, an unusually common accessory in the Hardyston, is second in abundance only to magnetite and limonite. It occurs as colorless to pale yellow, angular to subround, subequant to slightly elongate sand- and silt-size grains. Subhedral and euhedral detrital grains are common (fig. 8).

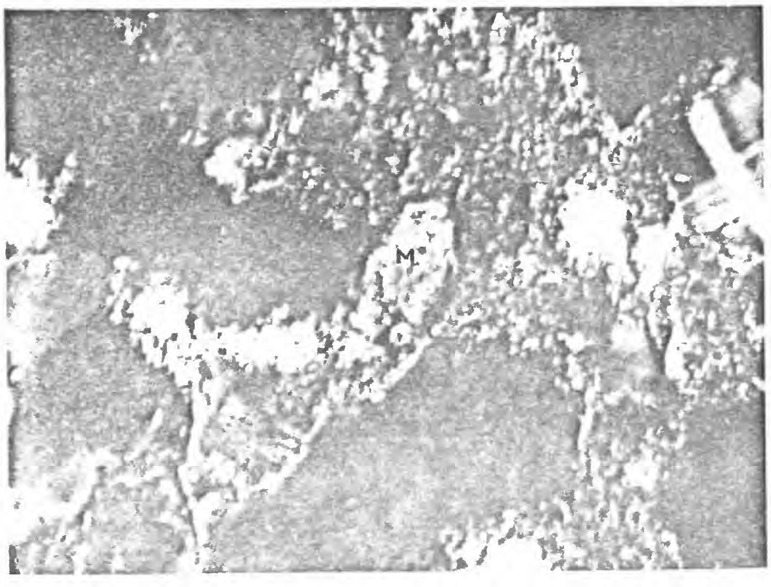
Although monazite is quite resistant to chemical destruction (Pettijohn, 1957), it is not notably resistant to abrasion (Friese, 1931; Cozzens, 1931). The angular, subhedral, and euhedral detrital grains suggest an extremely local source. Monazite is a common accessory in granites and pegmatites, both of which occur in the Reading Prong. Moreover, the prong is known to have several monazite-rich areas within the gneissic terrane in New Jersey (Williams, 1967); therefore, a local source for the monazite in the Hardyston is all the more probable. It is unlikely that the monazite is authigenic as most grains exhibit ample evidence of a detrital origin. The very euhedral grains show no evidence, relative to their surroundings, of having grown in place, and no authigenic overgrowth or detrital grains have been observed.

Zircon occurs in trace amounts throughout the Hardyston. The grains are small (0.125-0.05 mm), colorless, extremely well rounded (in marked contrast to the monazite), and contain very few inclusions. Very probably the zircon is multicyclic and was derived from

Figure 8. Photomicrographs showing some petrographic features of the Hardyston Quartzite. A, Authigenic feldspar overgrowth on detrital feldspar grain. Crossed polarizers, X100. B, Euhedral (above) and subhedral detrital monazite grains (m) with angular quartz grains. Crossed polarizers, X100.



A



B

metasedimentary rocks similar to those of the Reading Prong and Canadian Shield.

Tourmaline is common in some specimens, virtually absent from others. It occurs in subequant to elongate, angular to rounded grains of medium to fine sand. The grains are strongly pleochroic but the colors are variable; typically, they range from very pale brownish-yellow to pale green (E) and from dark brownish-green to dark green (O). Neither inclusions nor zoning is present; the grains are clear. These color varieties would appear to be part of the dravite (Mg-rich)-schorlite (Fe-rich) series that occurs in granitic and metamorphic or metasomatic rocks (Deere and others, 1966). In the Reading Prong tourmaline is present in some gneisses and granites, and in Precambrian pegmatites.

Garnet, hornblende, rutile, sphene, and topaz are found in trace amounts in some heavy separates of the crushed rock. They are very rarely seen in thin section and are not common enough to generalize or discuss their optical properties or occurrence.

Texture

The microscopic textural fabric of most Hardyston specimens is punctuated by streaks and stringers of magnetite, clay or micromica, and larger grains of quartz and feldspar. Otherwise the fabric is essentially structureless.

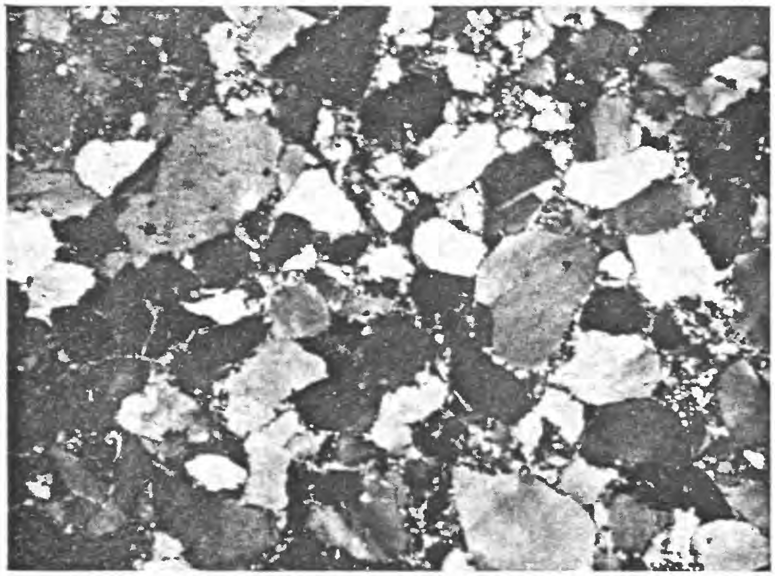
The degree of compaction or tectonic compression is indicated by the nature of grain-to-grain contacts. Most contacts range from straight to concavo-convex, suggestive of minor to moderate pressure solution (Adams, 1964). However, all possible contact relations, from sand grains "floating" in silica or clay cement to highly sutured grains, may be found at one place or another in the Hardyston (fig. 9).

Grain enlargement by secondary overgrowth on quartz and feldspar is common (fig. 8). The material forming the overgrowth presumably is the result of diagenetic precipitation but the diagenetically precipitated material may have been generated by pressure solution. Some of the most striking overgrowths, particularly on feldspar, occur in samples with abundant highly sutured grain boundaries. Abraded overgrowths have not been observed in the Hardyston.

Results of grain-size analyses using the microscope and sieve-size conversion methods developed by Friedman (1958; 1962) are summarized for four representative samples (tab. 6). Grain-size frequency distributions were determined by measuring the apparent long axes of grains in a thin-section and converting this apparent size to the equivalent sieve size according to Friedman's regression equation. The highly indurated character of all Hardyston rock-types precludes grain-size analysis by more conventional means. Thin sections with abundant sutured grain boundaries were excluded from the size analyses. The moment measures, determined graphically according to the methods of Folk (1961), are the statistical parameters used to summarize the frequency distributions.

Grain shape, a complex function of original shape, internal anisotropy, composition, directional properties, and depositional

Figure 9. Photomicrographs of Hardyston Quartzite.
A, Moderately sorted, subangular to subround quartz and feldspar grains in clay and silica cement. Note nature and variety of grain-to-grain contacts. Crossed polarizers, X30. B, Orthoquartzite showing sutured grain boundaries. Crossed polarizers, X30.



A

B

Table 6. Summary of textural parameters of typical Hardyston rock types

	Arkose	Arkosic Sandstone	Orthoquartzite	
Field No.	166-21	152-9	$R\text{-}K^3\text{-}32\text{-}C$	166-48
Mean grain size (mm)	0.34 Medium sand	0.7 Coarse sand	0.37 Medium sand	0.5 Medium-coarse sand
Sorting	Poor	Very poor	Moderate	Moderate-well
Standard deviation (Ø)	1.21	2.2	0.79	0.53
Skewness	Coarse (−0.23)	Strongly coarse (0.32)	Nearly symmetrical (0.07)	Fine (0.23)
Kurtosis	Very platykurtic (0.61)	Platykurtic (0.72)	Leptokurtic (1.20)	Leptokurtic (1.25)
Textural maturity	Submature	Submature	Mature	Mature
Grain shape	Very angular	Angular	Subangular to subround	Subangular

history, was determined on quartz and feldspar in the Hardyston by estimation and comparison with the roundness scale of Powers (1953). Thin sections with abundant sutured grain boundaries and grain overgrowths were not considered. Results are summarized in Table 6. Roundness ranges from very angular to subrounded, but most sand-size grains are subangular to angular. Roundness sorting is fair to good, and there is no apparent variation in roundness with composition. Shapes (sphericity) do vary with composition, however, because feldspar grains commonly are bounded by cleavage surfaces.

Textural maturity (Folk, 1961) is a function of degree of winnowing (clay content), rounding, and sorting. Hardyston rock types range from immature (greater than 5 percent clay and fine mica) to mature (less than 5 percent clay, well sorted but not well rounded). The Hardyston, however, is most typically submature (less than 5 percent clay, but grains poorly sorted and poorly rounded).

The textural properties described above tend to vary with stratigraphic position. In general, detrital grains are coarser and more poorly sorted in the lower part of the unit. This is not a precise trend however, and cannot be used to determine the stratigraphic position ("upper" or "lower") of an isolated outcrop.

Origin of the Hardyston Quartzite

The Cambrian stratigraphic sequences and lithostratigraphic correlations in the middle Appalachians are for the most part, rather well known (tab. 4). However, strong and ever-mounting evidence of very complicated structural relations and tectonic history of these rocks greatly hampers attempts to produce an integrated system- and process-oriented sedimentation picture.

Recent work in the Great Valley by the United States Geological Survey (Drake, 1967a,b; Drake and others, 1967; J. M. Aaron, this report) and the Pennsylvania Geological Survey (Field Conference of Pennsylvania Geologists, 1966) has shown convincingly that much of the Cambrian and Ordovician sequence between the Susquehanna and Delaware Rivers is regionally inverted and allochthonous. These rocks are involved in large-scale nappes and are displaced westward, probably by as much as 35 miles. In contrast, rocks of similar age and gross character in the Cumberland Valley section of the Great Valley, west of South Mountain and southwest from the Susquehanna River, are autochthonous (Field Conference of Pennsylvania Geologists, 1966). Thus tectonic events have juxtaposed initially widely separated parts of the depositional basin, telescoping the original facies relations and obscuring the depositional strike. Whatever relation, if any, exists between the ancient predeformation facies boundaries and the present regional strike of the rocks is totally unknown. Thus it is extremely difficult at this time to frame the origin of the Hardyston Quartzite in very meaningful paleogeographic terms.

Consideration of the physical and mineralogic properties of the Hardyston (tab. 7) does not necessarily lead to a unique solution regarding the origin of the detrital material and its environment of deposition. Ancient, complex, poorly exposed units of this sort, including its relatives throughout the Appalachians, do not have

Table 7. Summary of properties of Hardyston Quartzite

Composition: Mineralogy is very similar to subjacent Precambrian rocks, which are probable ultimate source. Quartz and microcline are principal minerals. Cement is both chemical and detrital. Faunal remains are absent to scarce and restricted. Little argillaceous material except in upper part. Authigenic limonite, pyrite, feldspar, and quartz.

Texture: Sorting is poor to moderate in lower part, moderate to good in upper. Rounding is fair to poor. Sand sizes predominate in all mineral species. Highly variable from outcrop to outcrop and from bed to bed.

Sedimentary structures: Crossbedded; exposures too poor to judge abundance reliably; no data on orientation. *Scolithus*(?) burrows in well washed sands. Beds tend to be lenticular and discontinuous, with erosional scour.

Internal organization: Irregularly asymmetric, with grain size decreasing stratigraphically upward. Conglomerates common in lower part.

Contacts and associated lithologies: Nonconformable basal contact. Upper contact transitional into silty shales and carbonates.

entirely convenient analogies among modern sediments. Modern sediments abound with cases in which the application of the interpretative principles that are conventionally applied to ancient rocks would produce conclusions that are diametrically opposite to what we know to be true by direct observation, i.e., interpretations of wind, water current, climate, source, etc. It is much easier, obviously, to understand sediment properties, types, and dispersal patterns when most of the controlling variables are reasonably well known. The reverse path, from sediment properties to the controlling variables, is infinitely more difficult, correspondingly more speculative, and less reliable. Some reasonable generalizations, however, can be made about the Hardyston, and there are some possible implications which must await further work.

The detrital mineralogy of the Hardyston rather strongly indicates that the ultimate source of the material was a high grade metamorphic and igneous terrane essentially similar to the Precambrian rocks currently exposed in the Reading Prong. The textures of the Hardyston suggest that the source of the detritus was quite local, because even the detrital grains that should have undergone fairly rapid physical modification are little altered, i.e., little rounded and nonspherical. Indeed many crystal faces and forms are still preserved. Although some quartz and feldspar, not significant amounts, and other minerals (notably zircon) occur locally as very rounded grains, their presence is not necessarily inconsistent with the above conclusion. Such materials could be extensively reworked (in an earlier sedimentary cycle) or multicyclic and still be very locally derived from some of the Precambrian metasedimentary rocks of the Reading Prong.

All the tangible evidence and the absence of any evidence to the contrary indicate that Hardyston detritus was transported and deposited in water. That at least the upper part of the Hardyston is marine is indicated by the very well washed, well-sorted sands, the possible glauconite, and the presence of marine fossils in calcareous beds. The poorly-sorted, poorly-washed, laterally and vertically variable, unfossiliferous arkosic sandstones and conglomerates of the lower Hardyston may well be alluvial, however. The paleogeographic picture thus envisioned is one of an Early Cambrian sea gradually inundating the low-lying flank of a shield of Precambrian rocks, which were thinly veneered by a broad apron of alluvial deposits. The encroaching sea reworked the upper part of the alluvial material. Subsequent deposits (upper part of Hardyston) were better sorted, better washed, finer (as the shoreline became more distant), and less variable.

The presence of relatively fresh feldspars in the Hardyston raises some interesting questions about the climate and relief in the source area. It may indicate that the relief was great enough to allow streams to breach the mantle of weathered rock (if there was any), thereby exposing fresh rock to vigorous stream action. Although the lack of strongly altered feldspars of the same initial composition as the fresh feldspars would argue against this, there is the possibility that the altered material was virtually destroyed in transport. One possible clue as to the relief on the basement complex is found in the Blue Ridge of northern Virginia, near Luray, where the Catoctin Formation of Reed (1955) (upper Precambrian or Lower Cambrian volcanic rocks) rests on a granitic basement having a relief of as much as 1,000 feet. Presumably the geomorphic history of the basement complex in northern Virginia was not too significantly different from that of the basement in eastern Pennsylvania, but this is questionable. Another possibility is that the local climate simply was too dry or too cold to support chemical weathering; or perhaps seasonal variations were extreme, and whatever clays and micromicas formed in a given wet season were spalled away, along with fresh rock, by physical extremes in a thermally variable dry season, and washed away by the next round of torrential rains. If the latter were true, however, claystones ought to be more common throughout the Hardyston rather than virtually confined to the transitional zone with the overlying Leithsville. Or, perhaps the clays were transported out of the immediate depositional area of the upper Hardyston by marine currents and were deposited farther from the shore.

Although the location of the source and the sediment dispersal patterns from it are more obscure than its petrologic nature, in all probability the Hardyston detritus came from a westerly or northwesterly direction, specifically a more extensively exposed Canadian Shield. Indeed, some paleogeographic reconstructions of the Early Cambrian (for example Lochman, 1956) place the shield well into eastern Pennsylvania even after partial marine submergence. Moreover, the Precambrian rocks of the Reading Prong itself probably were once part of the shield as the rocks of the prong are quite similar to those of the Grenville province of Canada and New York, and the Adirondacks, as noted by Engel (1956).

Studies of crossbedding in the Weverton Sandstone (Whitaker, 1955) and the Chickies Quartzite (A. Hohl, unpublished data, 1964), both at least partial equivalents of the Hardyston, indicate a westward source for these sediments. Both studies show paleocurrent azimuths trending east and southeast. Unfortunately neither has much to conclude about the probable origin of these units. Both are unfossiliferous and in both the variability of crossbedding orientations is rather low (75 percent of Whitaker's Weverton orientations are within 30° either side of due east), suggestive of an alluvial origin (Potter and Pettijohn, 1963). The plausible possible marine environments (beaches, bars, tidal basins, shallow-water marine) ought to produce either bimodal or more variable orientations (Pettijohn and others, 1965).

Other than the deposits of possible alluvial origin, the character, distribution, and age of the lower Cambrian clastic rocks in eastern and central North America clearly record a westward marine transgression. The basal Cambrian sequence from southeastern Pennsylvania to southeastern New York is Early Cambrian (Howell, 1956; Fisher, 1956). In the Nittany Valley of central Pennsylvania, the oldest known Cambrian is the Waynesboro Formation of Swartz (1948) of Middle Cambrian age, which is believed to have a westward source (Rodgers, 1956). Along the southeastern flank of the Adirondack Precambrian mass and throughout the eastern and northern interior of the United States the earliest Paleozoic clastic rocks are Late Cambrian (Fisher, 1956; Lochman-Balk, 1956; Bell and others, 1956). Thus, the ubiquitous Cambrian clastic rocks basal to the Paleozoic in the eastern United States are time transgressive and decrease in age and thickness northward and westward. The Upper Cambrian clastic rocks of the central and eastern interior are viewed broadly as facies equivalents of the coeval shales and carbonate rocks of the eastern states, by then far removed from shorelines and sources of coarse clastic sediment.

On a local scale, the Hardyston and Leithsville should not necessarily be considered in strictly "layer-cake" terms. They may indeed be partly contemporaneous facies equivalents, being no more than different parts of one and the same transgressive carbonate-orthoquartzite-arkose system. At any given Hardyston locality the superjacent Leithsville would, of course, be younger, with the contemporaneous Hardyston-equivalent Leithsville lying somewhat to the east. The transitional contact between these units and the presence of calcareous beds and scattered interbedded carbonate rocks in the upper part of the Hardyston are consistent with this interpretation.

Leithsville Formation

Introduction

The Leithsville Formation marks the transition from deposition of mostly terrigenous clastic sediment to the deposition of marine carbonates. The unit is at the bottom of a long sequence of carbonate rocks that underlies the Lehigh Valley. The lower and middle parts of the sequence chiefly are dolomites; the upper part mainly is limestone.

Lithology

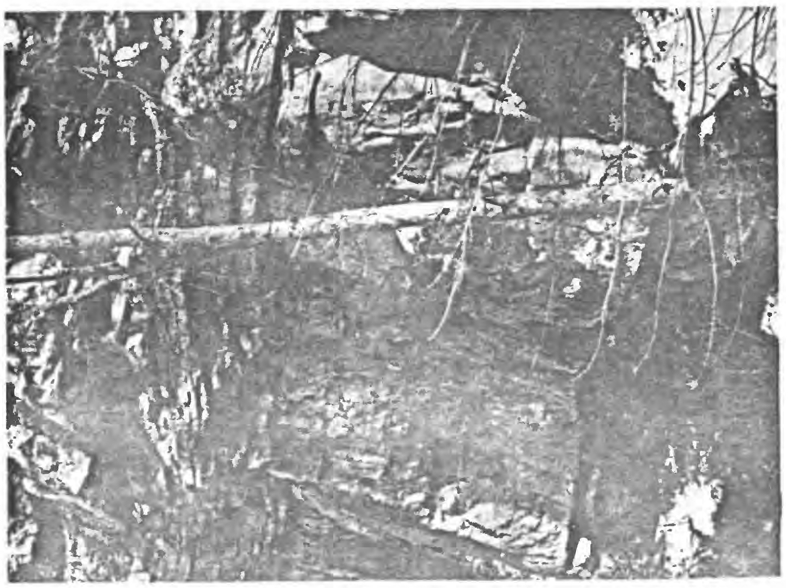
Principally the Leithsville Formation is medium to dark gray, light olive gray, and brownish black very finely crystalline to coarsely crystalline dolomite and dolarenite. It weathers very light gray, grayish orange, dark yellowish orange, or pale to dark yellowish brown. The dolomites are interbedded with thin pale brown sericitic shale, silty or dolomitic shale, quartz siltstones and sandstones, and dark gray calcitic dolomite. Sporadically present are dark gray bedded and nodular chert, oölitic and pisolitic dolomites and, in the upper part of the unit, poorly organized, non-laminated, indistinct hemispherical masses that the late V. E. Gwinn (oral communication, 1964) recognized as algal stromatolites.

Medium to coarse sand size subangular to subround vitreous grains of clear and bluish quartz, and feldspar, are scattered throughout the dolomites. Pyrite is visible locally and iron oxides conspicuously stain the rocks where it is concentrated.

Stratification in dolomitic rocks ranges from 1 cm to about 1 m. Partings ranges from 1 cm to more than 3 m. Very thickly parted (>100 cm) dolomites commonly have thinly laminated stratification and rocks of this type are rather characteristic of the Leithsville (fig. 10).

Figure 10. Photographs showing outcrop features of the Leithsville Formation. _A_, Very thick bedded dolomite and shaly dolomite in small abandoned quarry near Springtown, Hellertown quadrangle. _B_, Thick bedded laminated dolomite outcrop on River Road, south bank of Lehigh River, Nazareth quadrangle.

B



A

Shales and siltstones are present throughout the unit but appear to be more common in the lower parts. Stratification in these rocks is thin (1-5 cm), and aggregate thickness typically is less than about 30 cm, but may be greater locally.

Cross strata (cross laminated to very thinly cross bedded) are fairly common in the unit, as are graded beds, ripple marks, mud cracks, and "fucoids." Oölitic dolomite, dolorudite (edgewise conglomerate), and algal stromatolites are very sparingly present in the upper part of the Leithsville, and signal the initiation of depositional conditions and processes that produced the Allentown Dolomite, in which these features are very common.

The major lithotypes in the Leithsville are repetitive in a generally cyclic fashion, but poor exposure limits detailed knowledge of the nature, extent, and significance of the cycles. In general, however, a typical cycle begins with an influx of sand size terrigenous material that becomes progressively finer and more dolomitic upward, culminating in thinly laminated, very thickly parted beds of almost pure dolomite. The base of the cycle typically is a thin bed of quartzose material (orthoquartzite, impure quartz sandstone, or quartz-rich dolomite) that passes progressively upward into sandy siltstone, silty shale, calcareous (dolomitic) shale, and finally to thick units of laminated to thinly laminated, finely to aphanocrystalline dolomite. The Leithsville cycles are asymmetrical, and tend to be on the order of 10 to 20 feet thick. The variation in thickness is great however. These cycles are distinctly different from those of the superjacent

Allentown Dolomite, and can be used in mapping where exposures are adequate.

Name and Distribution

Wherry (1909) named the formation for exposures of gray dolomite with interbedded buff shales in the area around Leithsville, Pennsylvania, a village about four miles south of Bethlehem. Formerly these rocks were included in the Kittatinny Limestone, a formational name first used in New Jersey for the entire Lower Cambrian through Lower Ordovician carbonate section. Miller (in Miller and others, 1939, 1941) discarded Wherry's local usage and mapped these rocks as the Tomstown Formation, a name widely used in south-central Pennsylvania and adjacent Maryland, and northern Virginia.

Recent work (Field Conference of Pennsylvania Geologists, 1961, 1966; Drake, 1967a, 1967b; Aaron, this report) in eastern Pennsylvania has demonstrated widespread and previously unknown structural complexities involving allochthonous and far-traveled Lower Paleozoic rocks. The inter-regional correlation implied by Miller's Tomstown Formation are therefore suspect, and local names are preferable until further work unravels the tectonic framework. Moreover, the Leithsville contains dolomitic lithotypes that are suggestive not only of the Tomstown of south-central Pennsylvania, but also the Elbrook and Waynesboro Formations (tab. 4). Wherry's (1909) original definition of the Leithsville has been reinstated by most recent workers in the area.

Although rocks similar in age, stratigraphic position, and character to those of the Leithsville Formation occur throughout the Appalachians, rocks so named and mapped are restricted to the upper Delaware Valley of Pennsylvania and New Jersey, and to the Lehigh and Lebanon Valleys of Pennsylvania. A. A. Drake, Jr. (written communication, 1970) feels, however, that this rock-stratigraphic unit may be viable well north of the Lehigh-Delaware Valley, into central Massachusetts.

In the Nazareth quadrangle most of the rocks mapped as Leithsville occur south of the Lehigh River (pl. 1), along the flanks and base of the highlands of Precambrian rocks. It is well exposed in several quarries and bluffs along the river, but elsewhere exposures are scarce. Like the Hardyston, the Leithsville commonly is masked by rubble of Precambrian rocks from higher elevations.

Miller (1939) reported Leithsville along the south flank of Pine Top and Camel Hump (pl. 1), the Precambrian outliers. There are, however, neither outcrops nor other surface indications to support his contention.

Thickness

The thickness of the Leithsville Formation in the Nazareth quadrangle is about 1000 feet, as determined by geologic mapping and construction of cross-sections. This is in general agreement with reports of other workers in the area. Johnson and Willard (1957) report 900 feet in the Delaware Valley; Howell and others (1950) report 800-900 feet in Lehigh Valley and 400-500 feet in Buckingham Valley, an isolated patch of Paleozoic rocks within the Triassic (Newark) Basin. Drake (1967a) reports 1000 feet of Leithsville in the Easton quadrangle.

Stratigraphic Relations

The Leithsville conformably overlies the Hardyston Quartzite and the contact is transitional and has been described previously.

The Allentown Dolomite overlies the Leithsville with apparent conformity and the contact is transitional. This transition is marked by changes in the nature of the cyclicity, which is distinctive in each unit, as well as by changes in parting, and dolomite lithotypes present. In measured sections the contact is drawn at the bottom of the first typical Allentown cycle. In mapping, the contact is drawn at the stratigraphically lowest oölitic dolomite or well organized, laminated algal stromatolite.

Despite the apparently distinctive criteria for mapping the Allentown-Leithsville-contact, in practice it is not at all easy. These units have many attributes in common, and single hand-specimens and small exposures rarely are characteristic. In general, however, it might be said that the criteria for mapping the Leithsville are mostly negative (i.e., lack of the many distinctive features that characterize the Allentown) whereas the criteria for mapping the Allentown are quite positive.

Age

Wherry's (1909) definition of the Leithsville was based purely on lithologic criteria, among which was the absence of "cryptozoa" (algal stromatolites). As neither these nor any other fossils were known in the Leithsville at that time, Wherry assigned the Leithsville a Middle Cambrian age on the basis of its stratigraphic position between the relatively better dated Hardyston Quartzite (Early Cambrian) and Allentown Dolomite (Late Cambrian).

Although poorly organized, non-laminated stromatolitic structures are now known to occur in the Leithsville, their chronologic value is nil. A. A. Drake (oral communication, 1966; 1969, 1970) has reported a find by the New Jersey Geological Survey of specimens of the exterior opercula and impressions of internal surfaces of what is thought to be the Early Cambrian mollusk _Hyolithellus micans_ in the Leithsville, 8 to 10 feet above the Hardyston Quartzite, in New Jersey. This find is as yet undocumented, unfortunately. Transitional contact relations with the underlying Hardyston Quartzite (Lower Cambrian) and with the overlying Allentown Dolomite (Upper Cambrian) suggest that the Leithsville may, in fact, range in age from Early Cambrian to Late Cambrian. Although the bulk of it _may_ be Middle Cambrian in age, there is no specific evidence on which to base such an assignment. In this study the Leithsville is carried as Middle Cambrian(?).

Correlation

The Leithsville correlates with the lower parts of the Kittatinny Limestone of New Jersey and the Stockbridge Group of New York. Both are undivided carbonate sequences that overlie Lower Cambrian clastic rocks (Hardyston or equivalent) and comprise Lower Cambrian through Lower Ordovician rocks. Southwestward the Leithsville correlates with the Leithsville and part of the Buffalo Springs Formations of the allochthonous sequence in the Lebanon Valley, Pennsylvania (tab. 4), and the Tomstown and Waynesboro Formations, and part of the Elbrook Formation of the south central Pennsylvania authochthonous sequence (tab. 4).

Petrography

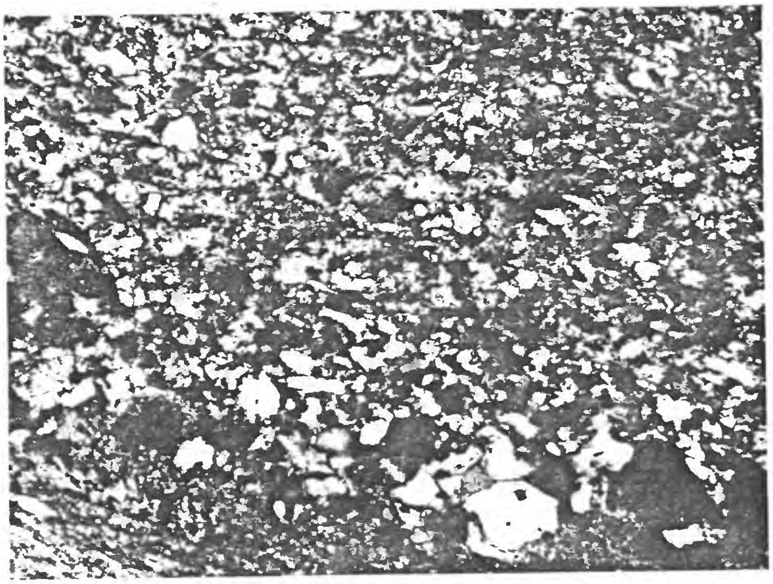
Mineral Composition

Dolomite is the principal mineral constituent of the Leithsville Formation. It occurs typically as subhedral to anhedral sparry crystals that range in size from 0.015 mm to about 0.2 mm (maximum diameter). Palimpsest relics, or "ghosts," of replaced small (average size about 0.05 mm) carbonate intraclasts, oöids, and pellets occur sparingly in mosaics of this type of dolomite. Also present, particularly in laminated dolomitic rocks in the upper part of the typical Leithsville cycles, is an essentially aphanocrystalline dolomite that is generally finer than 0.002 mm. In thin sections this variety is brownish gray and subtranslucent. Under crossed polarizers crystals appear as barely more than pin-points of light. This variety has been observed only in some laminae, and typically is associated with peculiar mottled textures (fig. 13).

Detrital quartz (fig. 11) is ubiquitous in the Leithsville except in the thinly laminated dolomites. It occurs throughout the sand size range in generally angular, poorly sorted, single grains. Invariably quartz grains display strongly undulose extinction, and are similar in every respect to quartz grains in the Hardyston Quartzite and to quartz in the Precambrian rocks, their probable source. Quartz in sheared rocks has a pronounced platy or bladed habit. Microcrystalline quartz mosaics petrographically similar to chert, are common in some sheared rocks, and constitute the "silicification" of such rocks. The abundance of quartz in the Leithsville is completely variable and the entire range from pure dolomitic rock to pure orthoquartzitic rock occurs.

Detrital feldspar is evenly divided between microcline and mesoperthite, and both are common, though volumetrically minor, wherever detrital quartz occurs. Microcline typically is prominently grid-twinned and shows little or no alteration. Mesoperthite invariably is differentially altered according to composition, with the sodic lamellae partly to completely converted to micromica or a clay mineral and potassic lamellae altered hardly at all (fig. 11). This type of alteration is much advanced over similar phenomena in the Hardyston. By way of comparison, very little mesoperthite occurs in the overlying Allentown Dolomite. Plagioclase, except for that in mesoperthite, is rare. All the feldspars are variable in both size and abundance. Sizes range from 0.006 mm to about 1 mm, and

Figure 11. Photomicrographs showing textural and mineralogic features of the Leithsville Formation.
A, Quartzose silt lamina in shaly dolomite. Note grading (base of a "microcycle") and angularity of detrital quartz grains; crossed polarizers, 100X. B, Differentially altered sand size detrital grain of mesoperthite. Sodic lamellae (dark) strongly altered; crossed polarizers, 250X.



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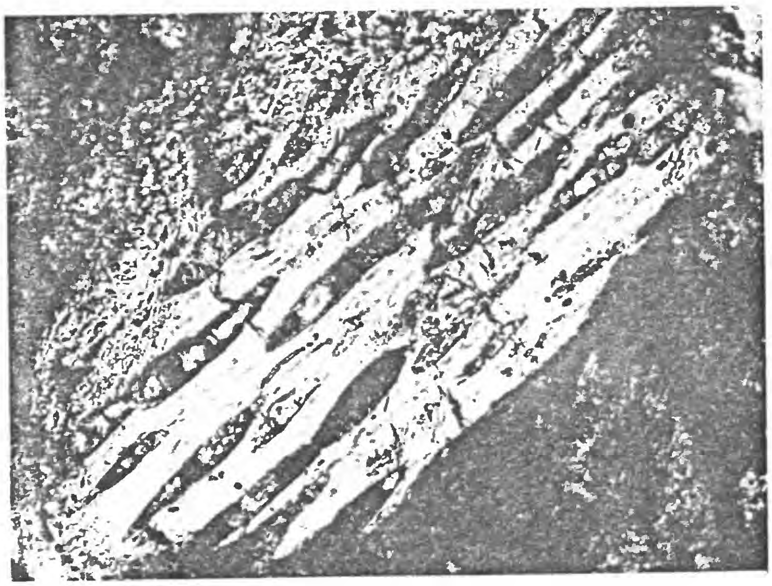

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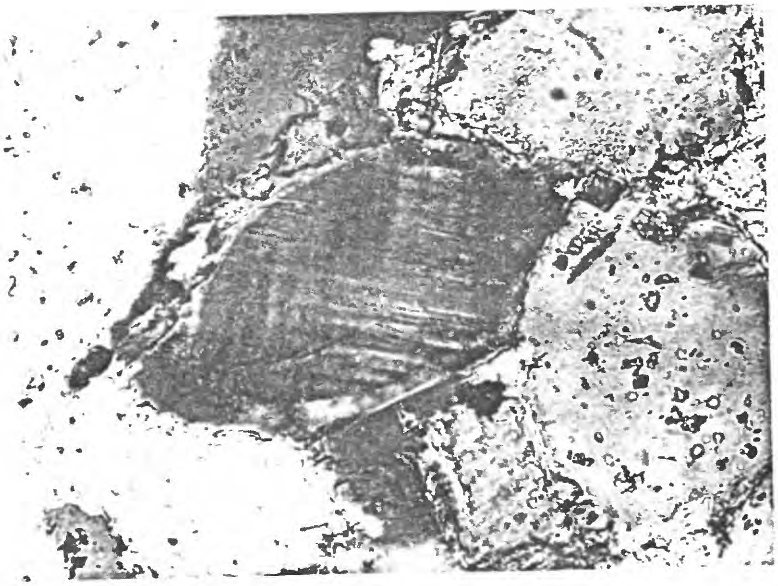
tend to be somewhat smaller than the average size of quartz grains in the same rock. Feldspar abundance varies directly with that of quartz, though on a much smaller scale. Even in very quartzose rocks (>75 percent) feldspar constitutes less than 10 percent, by volume, of the rock. Authigenic overgrowths occur on both microcline (fig. 12) and quartz, but are not particularly common on either.

Pyrite, magnetite, and limonite occur in practically every thin section of Leithsville. Pyrite occurs typically as small irregular blebs or well defined cubes that range up to 0.03 mm in size. Magnetite occurs as clastic silt-size grains. Limonite is the alteration product of both, and is common in weathered or sheared rocks. The sum of these three constituents rarely exceeds 1 percent of any given sample. The pyrite clearly is authigenic, and tends to be somewhat more abundant in the dolomitic and shaly rocks than in the more quartzose rocks, where detrital magnetite is common.

Monazite, tourmaline, zircon, and rutile are the only detrital heavy minerals, other than magnetite, that consistently are observed in thin sections. The occurrence of monazite is similar to that of the Hardyston, but the average grain size is somewhat less, typically medium to fine silt size. The grains are very angular and crystal outlines are preserved in many. Tourmaline is angular to subround, pleochroic green, and silt size. Authigenic overgrowths occur on tourmaline but these are very rare. Zircon is very round, clear, and silt size.

Figure 12. Photomicrographs showing textural features of the Leithsville Formation. A, Authigenic overgrowth of potassium feldspar on detrital microcline grain in quartzose unit; plane light, 100X. B, Contact of mottled allochemical(?) dolomite lamina and fine-grained sparry dolomite lamina showing no textural relics; plane light 100X.



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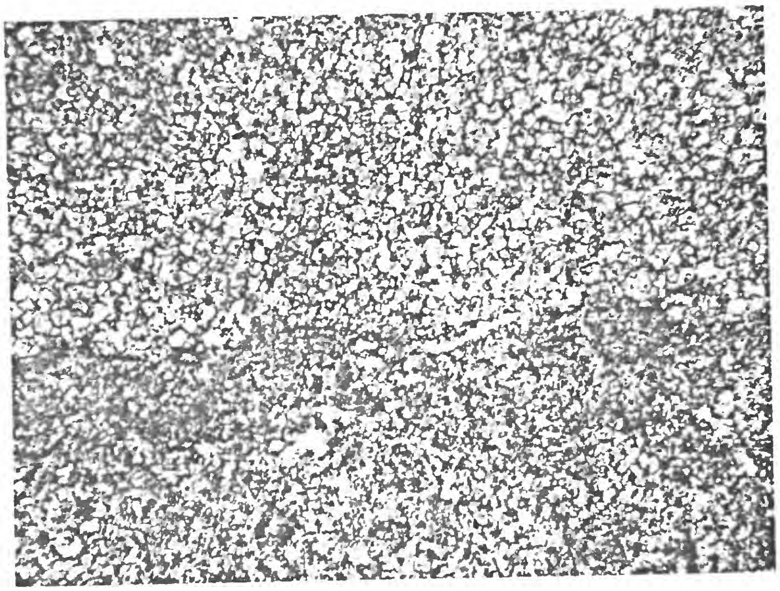

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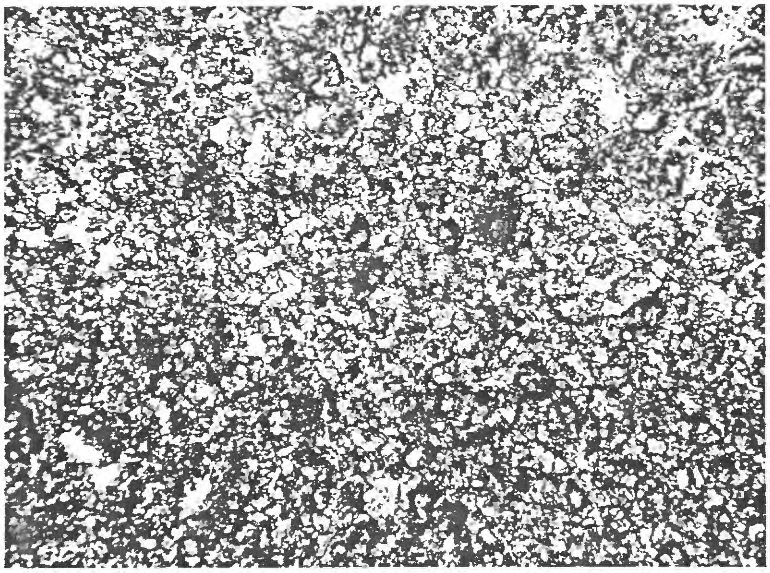
Shaly and phyllitic rocks are composed mostly of clay-sized phyllosilicates, mostly sericite with lesser chlorite, and contain abundant silt and clay size quartz, feldspar, and authigenic pyrite. Dolomite usually is present interstitially in variable amounts.

Texture

The principal texture of the Leithsville dolomites typically is hypidiotopic to xenotopic equigranular. The only exception occurs in some strata of the laminated rocks where a peculiar mottled texture (figs. 12 and 13) is inequigranular, presumably xeontopic. Allochem ghosts are common and are relics of the calcite pre-dolomitization texture. They indicate a clastic origin for rocks containing them. Dolomite crystals commonly transect allochem boundaries and such relations attest to the replacement origin of rocks containing them. Many fine to medium grained sparry dolomites in the Leithsville contain no recognizable textural relics at all, however, and in these cases, assuming that any evidence of a clastic texture was not simply destroyed, the original sediment must have been essentially chemical rather than chemiclastic. This lithology commonly is associated with mottled laminae in the laminated rocks (fig. 12).

The nature of the mottled texture of some laminae is difficult to fathom. The finer grained dolomite in this inequigranular texture (fig. 13) appears to be only a coating of variable thickness on ghost-like structures that resemble carbonate allochems. The shapes of most such allochems suggest intraclasts, oölites, or pellets.

Figure 13. Photomicrographs showing Leithsville dolomite textural features. A, Mottled texture in laminated dolomite. Mottlings defined by dark very fine grained dolomite, and resemble carbonate allochem ghosts; plane light, 35X. B, Same; plane light, 250X.



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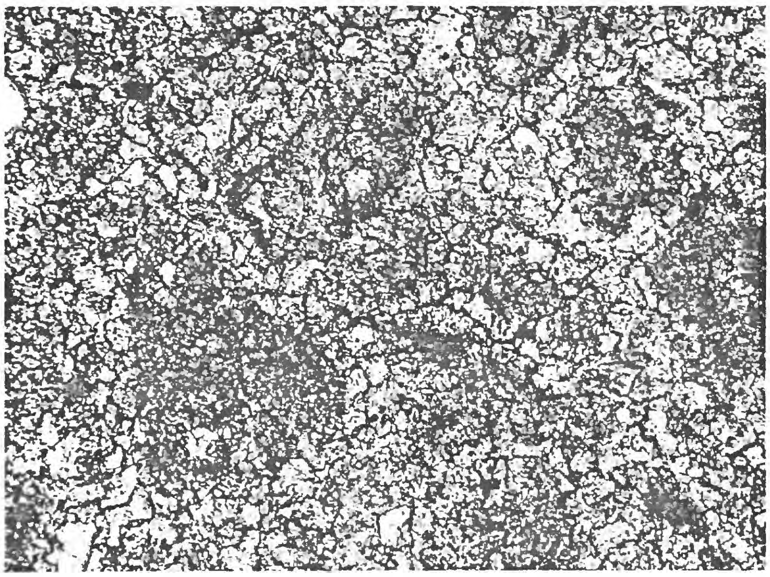


B

However, some of the very fine-grained material also occurs as irregular clots within the sparry dolomite or as what appears in thin section to be no more than a dust over the sparry dolomite mosaic that is visible in the background.

Quartz occurs as widely scattered grains and in thin lenses and stringers. Stratified concentrations typically are graded (fig. 11). Dolomite crystals engulf quartz grains in most cases, and even highly quartzose rocks (e.g. quartz sandstones) contain much interstitial dolomite cement, a petrographic feature that readily distinguishes these rocks from Hardyston sandstones.

Sericite in shaly and phyllitic rocks is randomly to strongly oriented. Strong alignment generally is associated with sheared or otherwise strongly deformed rocks. Other detrital minerals in such rocks are similarly affected. Most sericite, except in sheared rocks, is detrital.

Chemical Composition

Three chemical analyses of dolomites from the Leithsville Formation in or near Nazareth quadrangle are presented in Table 8. All reflect the rather high degree of purity of these rocks; and some approach the composition of some of the purest known dolomites. The most variable components are SiO_2 and $Fe_2O_3+Al_2O_3$, which mostly reflect variable proportions of detrital material.

Table 8. Chemical analyses of dolomites from the Leithsville Formation, Northampton County, Pennsylvania

	1	2	3
$CaCO_3$	52.16	53.43	57.05
$MgCO_3$	43.28	41.43	37.90
Al_2O_3	0.58	2.20	2.62
Fe_2O_3			
SiO_2	1.90	4.77	2.70
S	0.020	--	--
P	0.006	--	--
	97.95	101.83	100.27

1. Quarry at Island Park, Easton quadrangle. Average of 108 carloads (Miller, 1934).

2. Bethlehem Steel Co. quarry at Redington, Nazareth quadrangle, Pa. (Miller, 1934).

3. Bethlehem Steel Co. quarry at Redington, Nazareth quadrangle, Pa. (Miller, 1934).

Origin of the Leithsville Formation

The Leithsville is partly clastic and partly chemical in origin, and, along with the Hardyston Quartzite, records the gradual submergence of the seaward flank of the North American Precambrian shield that lay mainly to the west and northwest of the eastern Pennsylvania area in Cambrian time. As previously described in the discussion of the Hardyston, the Leithsville Formation probably represents, at least in part, the further offshore phase of a transgressive orthoquartzite (upper Hardyston)-carbonate (Leithsville) system. Such a relation is suggested by the transitional contact between the two units, the abundance of terrigenous clastic debris in the Leithsville, and by the similar ages of the two units.

Much of the dolomite in the Leithsville is of replacement origin. The original sediment was composed largely of calcium carbonate, either as calcite or as aragonite. The evidence for replacement of such sediment is the presence as dolomite of such features as oöids, pseudoöids, pellets, and algal stromatolites, all of which have been observed to form only as calcite or aragonite, never as dolomite. Petrographically such features occur only as ghosts in the equigranular dolomite fabric. Dolomite crystals transecting the boundaries of these features are further evidence of replacement.

The apparent ease with which vestiges of the original sediment were preserved in these rocks suggests that the dense, very fine grained, laminated dolomites that lack textural relics may have had a somewhat different origin. One possibility is that deposition of calcium carbonate occurred in waters so little agitated that features such as oöids, pseudoöids, carbonate intraclasts, ripple marks, and the like could not form. However, one might still expect to find algal stromatolites under such conditions, but these features are not present in the fine-grained laminated rocks either. Neither are any other fossils nor other indications of animal activity, such as fecal pellets or burrowings. Another possibility is that these dolomites are primary in the sense that the position of a given dolomite crystal in the rock was not previously occupied by calcite or aragonite. The dolomite itself may have been the product of syngenetic replacement of calcite or aragonite, but replacement occurred before the crystals attained a coherent calcium carbonate texture. A third possibility is that the dolomite composing these fine-grained laminates is the product of direct chemical precipitation from sea water. Such deposition might have been possible in very quiet marine waters with little or no circulation, under evaporitic conditions. Such conditions are associated with relatively high water and air temperatures and high pH, and would be hostile to many forms of marine life. Moreover, such conditions would be consistent with the laminated character of the bedding that indicates protection from most currents, and the lack of fossils, borings, and current features, and the presence of mud cracks, which indicates shallow protected water that is, on occasion, subjected to desiccating conditions.

The lithologic cycles in the Leithsville record repeated changes in depositional pattern that may be in response to several different causes that acted singly or perhaps simultaneously. The cycles are essentially large scale graded-beds with sand size terrigenous material at the base grading upward to progressively finer noncarbonate clastic material and progressively more carbonate (now dolomite). Such changes may or may not be related to tectonism. Non-tectonic causes include 1) a stable sea level with progressive changes in the size and abundance of terrigenous material brought to the carbonate depositional basin owing to periodic flooding, and 2) an oscillating sea level (eustatic) that would allow into the depositional basin more or less, coarser or finer, terrigenous material according to the proximity of the depositional area to the shifting shoreline. Tectonic causes could include sea level changes and concomitant changes in the vigor of erosion and sediment transport brought about by uplift or depression of the source lands or depositional basin.

The data on hand do not permit a conclusive choice among these possibilities. It is conceivable that more than one factor could have played a part in the development of cyclicity. However, some suggestion of tectonic activity may be found in the fact that many sedimentary features of the Leithsville, such as mud cracks, algal stomatolites, and oöids, typically form only in shallow to very shallow water (perhaps no more than a few inches deep). For a unit of 1000 feet or more to develop with such features, some subsidence of the depositional basin must have occurred more or less continuously with deposition.

It is therefore reasonable to postulate that Leithsville cyclicity is somehow related to tectonic activity and basin subsidence. A pulse of tectonism and associated basin subsidence could trigger local changes of sea level or changes in the balance of processes in the geomorphic-sedimentation system in the shield area and give rise to an influx of coarse terrigenous sediment in the marine depositional area. As the effects of such a pulse died out, terrigenous clastic sediment would become finer and less abundant and gradually would give way to carbonates, first mostly clastic limestones and later, in the quiet, waning stages of the cycle, mostly material of dominantly chemical origin, including dolomites.

Dolomitization probably accompanied the late stages of the cycles, when conditions favored either direct chemical precipitation of dolomite or syngenetic replacement by dolomite of calcium carbonate, either of which case would produce dolomites that would be essentially primary in the petrographic sense. Such an environment also would be conducive to diagenetic replacement of calcium carbonate sediments that formed earlier in the cycle.

The foregoing remarks about the origin of dolomite in the Leithsville Formation raise questions that apply equally to other stratigraphic units of concern to this study. Although the subject will be broached repeatedly as these other units are described and discussed, more specific discussion of dolomitization and the carbonate depositional environment will be deferred until all the pertinent units have been described.

Allentown Dolomite

Introduction

The Allentown Dolomite is a very heterogeneous assemblage of dolomites that are characterized by the nature and variety of sedimentary textures and structures that they contain. Also characteristic is the cyclicity of dolomite lithotypes present. These positive attributes make the Allentown the most distinctive rock unit in the succession of dolomitic rocks that underlie the Lehigh and Delaware Valleys. The Allentown is of unusual petrographic and petrologic interest because the above characteristics, and others, combine with relatively good exposures to reveal much about carbonate depositional processes, environments of deposition, and sedimentary history of the rocks.

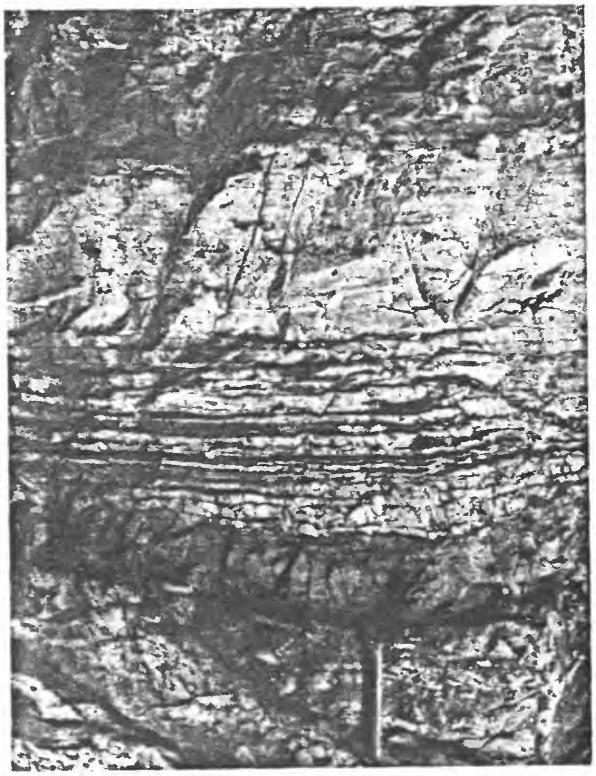
Lithology

The Allentown is composed of dolomites that range from aphanocrystalline to coarsely crystalline. These include fine to very coarse oölitic and pellet dolomites, very fine to very coarse dolarenites, fine to coarse dolorudites, several varieties of algal stromatolites as well as dolomites that lack such distinctive subtextures and structures. The allochems in the dolarenites and dolorudites are broken and abraded carbonate fragments (intraclasts), and are entirely dolomitic. The dolomites are medium light gray, medium gray, medium dark gray, dark gray, and light brownish- to light olive gray. They weather to lighter or darker shades of gray, generally very light gray, light gray, light olive gray, medium dark gray, and dark gray. Alternation of these lighter and darker weathering colors is quite conspicuous in most outcrops and sections, and is characteristic of the unit. Also characteristic is the cyclicity of textures and features mentioned above and sedimentary structures, which include large and small scale oscillation ripple marks, mud cracks, cross stratification, and minor disconformities (fig. 14).

Thin beds of quartz sandstone, orthoquartzite, and bedded or nodular chert are interstratified with the dolomites, but these compose a minor part of the unit. Quartz sand is widely scattered throughout the unit and is particularly conspicuous in outcrops and hand specimens of the dolarenites and dolorudites.

Oölitic dolomites are conspicuous and they constitute between 10 and 15 percent of the unit, mostly in the lower and middle parts. The upper part of the unit contains much less than 5 percent oölitic dolomite. Oöids occur in a variety of types and sizes but are well sorted within a given depositional unit. Sizes range from 0.25 mm to 2.0 mm in diameter (medium to coarse sand). Some are dark gray in a light gray matrix; others are light gray in a dark gray matrix. Concentric structure is apparent in many, in hand samples, but in others concentricity can be determined only under the microscope. Most are spherical but some are strongly flattened. Flattening is due to both tectonic and geostatic deformation, but it is not always easy to ascribe the flattening to one process or another in any given sample. Oöids whose long axes are strongly oriented in the plane of rock cleavage and perpendicular to fold axes presumably

Figure 14. Photographs showing some lithologic features of the Allentown Dolomite. A, Thin quartz sandstones (middle, dark bands), and desiccation dolorudites (lower right, upper left). Photo taken along tracks of Central Railroad of New Jersey, north bank of Lehigh River, Nazareth quadrangle. B, Cross-stratified quartzose dolarenite. Photo taken along Bushkill Creek, just east of road intersection one-quarter mile north of Walters, Nazareth quadrangle.



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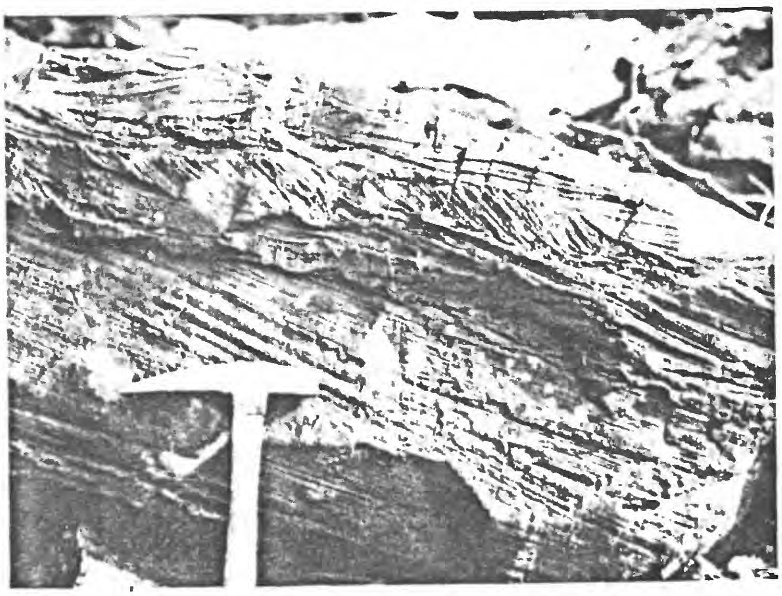

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were flattened by tectonic processes. Squashed or ruptured oöids with no strong or consistent orientation relative to each other or to observable rock structures would seem to result from geostatic deformation. However, the problem is complicated by the fact that the rocks commonly are part of recumbent isoclinal folds, hence the plane of bedding would contain both tectonically deformed oöids lying in the plane of cleavage (which would be parallel to bedding in fold flanks) and oöids deformed by the weight of accumulating superjacent sediment.

Peculiar oöids whose internal structures suggest an unusual and complex history occur in the Allentown. Some oöids are divided internally into light and dark halves separated (in cross section) by a line that is convex upward (stratigraphically). These have been called "half-moon oölites" (Wherry, 1916). Others have internal shells that have collapsed onto the core, but maintain a spherical outer shell. Still others have outer shells that are quite asymmetrical in terms of thickness and have collapsed inner shells as well. Although such features are rare and structural details are observable only in thin sections, they can be used as top and bottom indicators. More detailed descriptions and interpretations of these features will be presented in the section on petrography.

Algal stromatolites are another distinctive and useful feature of the Allentown and, like oölites, are more abundant in the lower and middle part of the unit. Although stromatolites also occur in the Leithsville Formation, they are much less common there and tend to be rather massive and lack the fine lamination of those in the Allentown. Aside from the rare occurrences in the Leithsville, algal stromatolites are not known to occur in any other lower Paleozoic stratigraphic unit in the Lehigh or Delaware Valleys.

Algal stromatolites in the Allentown Dolomite are mound-like hemispherical structures composed of laminae of sand, silt, and clay-sized detrital sediment and finely crystalline dolomite (presumably after aragonite). They are the remains of algal colonies which, by analogy with modern or Recent forms, built the laminated structures by trapping and binding sediment and fine calcite or dolomite crystals by means of an organic film that probably was a complex of filimentous and unicellular green (chlorophyta) and blue-green (cyanophyta) algae (Logan and others, 1964).

Genus and species were used in the earliest descriptions of algal stromatolites and the practice has persisted to the present. Accordingly, at least five "species," distinguished by the various forms taken by the laminae constituting the structure, occur in the Allentown Formation. Some of these are illustrated in Figures 15 and 16.

Some modern work (Logan and others, 1964; Aitken, 1967) has shown convincingly that the diversity of forms on which the binomial nomenclature is based is the result of the interaction of the algal film (or organic mat), detrital sediment, and physical environmental factors. Moreover, the sediment-binding algal mat itself may contain a large number of different species, the remains of which are seldom preserved in ancient sediments. Thus algal stromatolites

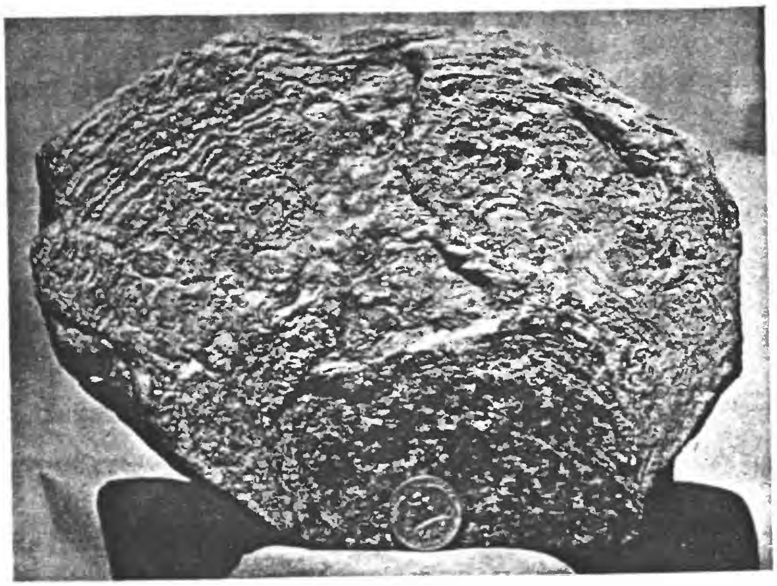
Figure 15. Photographs showing organo-sedimentary structures (algal stromatolites) in the Allentown Dolomite. Structures in both photos are compounded of two or more form-species. Both photos taken at 13th street intersection, Rt. 22, city of Easton, Easton quadrangle.

A

B

Figure 16. Organo-sedimentary structures in the Allentown Dolomite. A, Cryptozoon-type algal stromatolites in strongly overturned (nearly horizontal) dolomite beds. Photo taken along tracks of Central Railroad of New Jersey, north bank of Lehigh River, Nazareth quadrangle. B, Large algal head with structure compounded of Cryptozoon and Collenia form-species growing on large piece of carbonate rubble. Specimen collected along Nancy Run, Nazareth quadrangle.

A



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should more properly be treated as organo-sedimentary structures rather than as fossil organisms, and the conventional biologic nomenclature probably should be abandoned.

Algal stromatolites in the Allentown have neither particular chronologic significance nor value as diagnostic stratigraphic markers within the unit. However, regardless of the names, possible environmental modifications, and whether they are organic or inorganic, the upwardly convex shapes are extremely useful as top and bottom indicators in this complex terrane. The writer knows of no case in this area, or any other, where tops of beds as indicated by the arched laminae were contradicted by other primary sedimentary indicators.

Dark gray bedded and nodular chert is present, mostly in the upper part of the unit. It becomes progressively more common toward the top, as oolitic dolomite and algal stromatolites disappear.

Perhaps the most remarkable and diagnostic characteristic of the Allentown Formation is the cyclic repetition of dolomite rock types. The light and dark alternations are very conspicuous for some distance in the larger weathered cuts and sections, and the unit seldom is misidentified where such cuts are available (fig. 15). The color differences rarely are significant in fresh rock, however.

Although the cyclicity of the Allentown has been recognized for some time, little was known in detail until Zadnik (1960) studied two long sections on the Delaware River in New Jersey. He considered the ideal cycle to be as follows, from bottom to top: 1) structureless dololutite, 2) dolarenite, 3) oölitic dolarenite,

4) dolorudite, 5) cryptozoan dolomite, and 6) desiccation dolorudite. Individual cycles typically are not complete, as not all rock types are represented in every cycle. The cycles are asymmetrical and single cycles range in thickness from 5 to more than 25 feet. The Allentown cycles will be discussed in more detail in a later section after chemical and petrographic data are presented.

Name and Distribution

Wherry (1909) proposed the name Allentown Limestone for the alternating light and dark oölitic and stromatolitic carbonate rocks of the Lehigh Valley, but his definition was not based on proven mappability. Miller and others (1939) adopted Allentown Formation for the same stratigraphic unit in their map and description of Northampton County. Howell and others (1950) revised the Cambrian stratigraphic nomenclature of the area and subdivided Miller's (1939) Allentown Formation into the Limeport Formation, below, and the Allentown Formation (restricted), above, separated by a middle Late Cambrian disconformity. Their subdivision was based on their notions of the stratigraphic and chronologic significance of algal stromatolites, and on the fact that only Dresbachian (early Late Cambrian) and Trempealeauian (late Late Cambrian) fossils had been found in the Allentown Formation of Miller and others (1939). Howell and others (1950) presented no evidence to suggest that they had tested the subdivision by mapping. Drake (1965) found the subdivision of Howell and others (1950) unmappable in the Delaware Valley and rejected it, reinstating Wherry's (1909) original definition but changing the lithologic name to the more appropriate dolomite. Although many of the critical localities cited by Howell and others (1950) in their definition of the Limeport and Allentown (restricted) Formations are within the Nazareth quadrangle, field mapping does not support the subdivision, and Drake's (1965) terminology is herein adopted.

The Allentown is widespread throughout the Lehigh and Delaware Valleys. Its outcrop belts thin and disappear northeastward, in New Jersey, owing to structural complications. Southwestward along strike the Allentown has been mapped into eastern Berks County, beyond which detailed relations are either unknown or other names (see Table 4) apply.

In the Nazareth quadrangle the Allentown is well exposed in a broad belt trending east-northeast across the south central part of the area. Exposures are superb adjacent to the bed of the Central Railroad of New Jersey, on the north bank of the Lehigh River, from Freemansburg northwestward to Easton (pl. 1), a distance of nearly six miles. Unfortunately these cuts roughly parallel the regional strike so that the stratigraphic thickness actually exposed is rather limited. Good exposures are also available in and along streams that drain into the Lehigh River from the north, particularly Nancy Run (a local name which does not appear on topographic maps), and along Bushkill Creek, northwest of Easton, adjacent to the eastern border of the area.

Thickness

The thickness of the Allentown Dolomite is estimated to be about 1700 feet, but estimates vary. No complete sections of the Allentown are known, and structural relations invariably are complex. Zadnik (1960) measured 1610 feet of Allentown at the reference section on the Delaware River at Carpentersville, New Jersey, where the base of the formation is not exposed.

Stratigraphic Relations

The Allentown Dolomite conformably overlies the Middle Cambrian Leithsville Formation. The contact is transitional, and represents a change from characteristic Leithsville cyclicity to the typical Allentown cycle. In mapping, the lower contact is drawn at the lowest observed oölite or algal stromatolite. In measured sections the lower contact is drawn beneath the lowest Allentown-type cycle.

The Allentown is overlain conformably by the Lower Ordovician Rickenbach Dolomite, the lower unit of the Beekmantown Group in this area. The contact is transitional and in mapping it is drawn above the highest typical Allentown algal stromatolite, oölitic dolomite, edgewise dolorudite, or orthoquartzite, or beneath the lowest very dark gray, very coarsely crystalline, thin bedded calcareous dolomite. In measured sections the contact is drawn above the highest Allentown cycle.

Age

The Allentown Dolomite is considered Late Cambrian in age on the basis of relatively scarce paleontologic evidence. No fossils of any kind have been found in the Allentown in Nazareth quadrangle. There are only seven very widely scattered fossil localities in the Allentown throughout eastern Pennsylvania and western New Jersey. Some of these localities have been collected only once and have never been reestablished, despite repeated attempts to do so. Others are isolated in the sense that no detailed mapping has established the stratigraphic and tectonic control required to make meaningful use of the paleontologic age determinations.

Weller (1900) dated the lower part of the Allentown Dolomite as early Late Cambrian (Dresbachian) on the basis of the trilobite <u>Welleraspis jerseyensis</u> that he collected in a now abandoned quarry in the Carpentersville, New Jersey, reference section. Howell (1945, 1957) reported Dresbachian fossils in the Allentown at Peapack, New Jersey, and Limeport, Pennsylvania. The Limeport locality is part of a fault block of lower Paleozoic rocks within the Triassic basin. Weller (1903) and Howell (1945, 1957) reported late Late Cambrian (Trempealeauian) fossils from the upper Allentown at Portland, Pennsylvania, Blairstown, New Jersey, Newton, New Jersey, and Andover, New Jersey.

The middle Late Cambrian disconformity postulated by Howell and others (1950) between their Limeport and Allentown (restricted) Formations, and to some extent their proposed stratigraphic subdivision itself are based on the fact that no middle Late Cambrian

(Franconian) fossils have been found in the rocks that constitute the Allentown Dolomite of this report. However, the lack of such fossils hardly seems significant in the light of such rare fossil occurrences and the few genera that are represented. All fossil localities are acknowledged to have been from "uppermost" or "lower" Allentown. The apparent hiatus may simply reflect the fact that no fossils have been found in rocks of the "middle" Allentown.

Correlation

The Allentown Dolomite is correlated with the Kittatinny Limestone of New Jersey (Lewis and Kümmel, 1940) and the Stockbridge Group of New York (Broughton and others, 1962), both of which range from Lower Cambrian through Lower Ordovician, undivided. Southwestward, along strike, the Allentown is correlative with the upper Cambrian Conococheague Group and part of the Buffalo Springs Formation of Lebanon Valley, Pennsylvania, and the upper Cambrian Conococheague Group and part of the Elbrook Formation of Cumberland Valley, Pennsylvania (Field Conference of Pennsylvania Geologists, 1966). These relations are summarized in Table 4.

Petrography

Mineral Composition

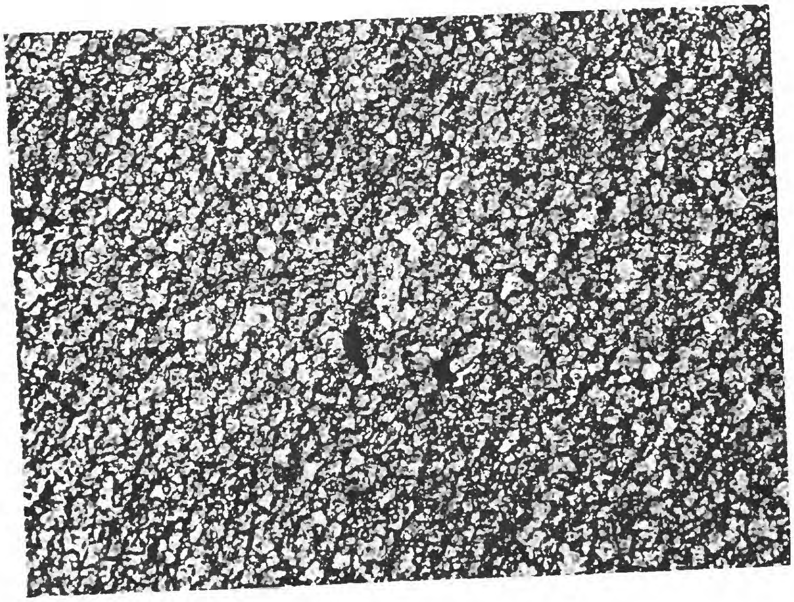
Twenty thin sections of representative Allentown lithotypes, six insoluble residues, and 37 sawed rock slabs were examined in detail under petrographic and binocular microscopes, using both transmitted and reflected light. X-ray mineral identifications were performed where necessary.

The Allentown Dolomite is composed almost entirely of dolomite rock, which is defined herein as a rock that contains 80 percent or more, by volume, of the mineral dolomite. Dolomite is practically the only carbonate mineral present in the Allentown. Calcite is rare to nonexistent except as a joint filling or coating. Most dolomite rocks in the unit are at least 85 percent dolomite and, in fact, most contain more than 95 percent dolomite. The only other volumetrically significant mineral constituent is detrital quartz and microcrystalline quartz (chert).

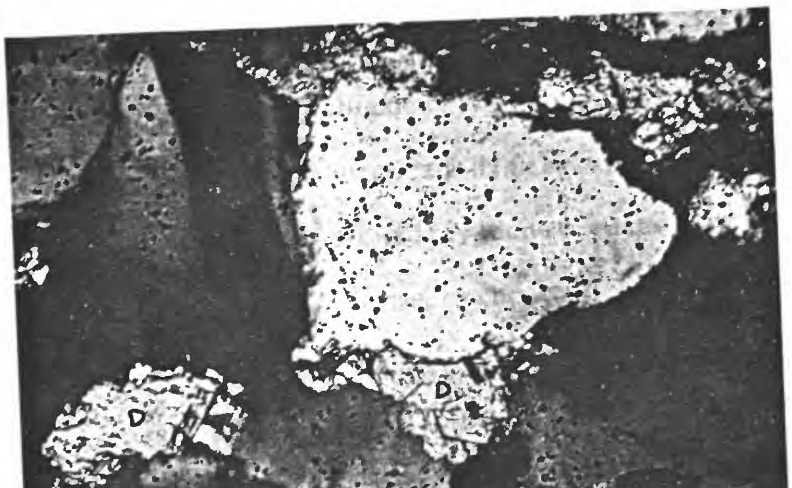
<u>Dolomite</u> occurs in subhedral to anhedral crystals that range in size (long diameter) from 0.001 mm to 1.0 mm. Two distinct petrographic varieties exist: 1) sparry dolomite and 2) aphanocrystalline dolomite.

Sparry dolomite (fig. 17) is by far the more common variety. It typically occurs in sizes that range from 0.35 mm to 0.05 mm, and constitutes the bulk of all carbonate lithotypes in the Allentown. It is clear, or nearly so, in transmitted light, but commonly has a somewhat dusty appearance owing to the presence of noncarbonate impurities that survived dolomitization. These dusty areas typically have regular shapes that clearly outline recognizable allochemical textural elements (oöids, lithic fragments, etc.) of the original carbonate (presumably aragonite) sediment, and are palimpsest relics, or "ghosts," thereof.

Figure 17. Photomicrographs of some petrographic features of the Allentown Dolomite. A, Sparry dolomite mosaic, finely crystalline dolomite. Plane light, 35X. B, Typical quartz sandstone interbed with interstitial dolomite (D). Note undulose extinction and angularity of grains. Polarized light, 100X.



A



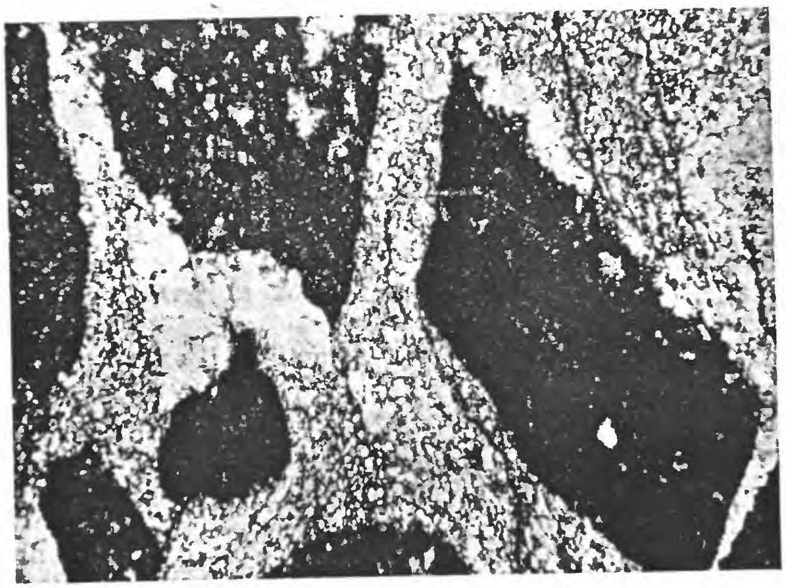
B

Aphanocrystalline dolomite ranges in size from 0.004 mm down to 0.001 mm, or less, and is dark gray and subtranslucent in transmitted light (fig. 18). There is no transition in size between this variety and the sparry dolomite. They represent two distinct size spectra. Beads, stringers, and irregular clots and swirls of sparry dolomite commonly occur within the aphenocrystalline material giving it a peculiar "birdseye" or mottled appearance (fig. 18). Other than these and indistinct laminations, no recognizable structures of any kind occur in this variety, nor do palimpsest relics. This type of dolomite typically occurs as angular to subrounded elongate to platy lithic fragments ranging in size from 2 mm to 20 mm in ruditic dolomites and edgewise (flat pebble) dolorudites, and is the principal large clast in these rocks (fig. 18), though other lithic types may predominate locally. These clasts also occur, though much less commonly, in the dolarenites where they tend to be smaller (0.5 mm-2.0 mm), more rounded, and rather compact.

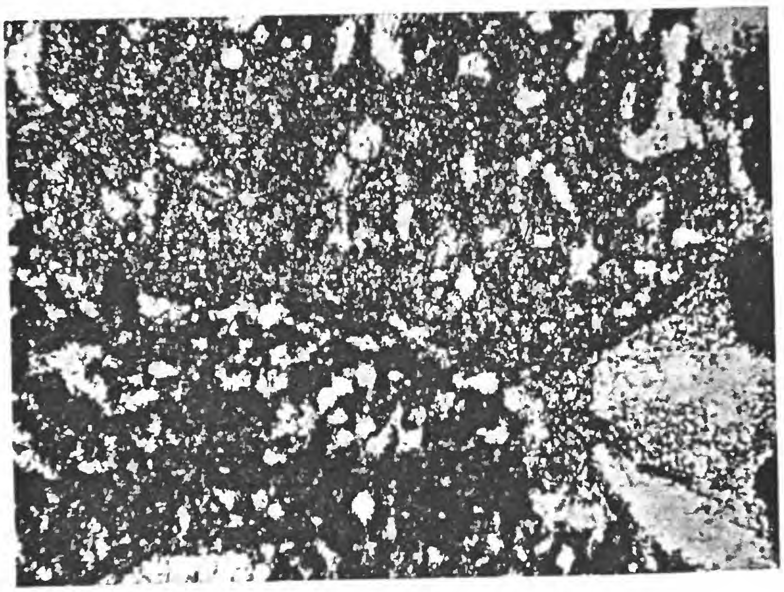
Quartz is widely distributed in highly variable proportions throughout the Allentown. It generally constitutes 5 to 10 percent of most dolomites, but locally forms orthoquartzites with less than 5 percent interstitial dolomite.

Quartz occurs mostly as single, rarely composite, grains ranging in size from 0.08 mm to 0.5 mm. The maximum range is 0.1 mm to 1.0 mm. Typically the grains are angular to rounded with slightly to strongly undulose extinction (fig. 17). The tendency to show undulose extinction is, of course, much decreased in the smaller grain sizes, as a smaller part of the deformed atomic lattice is

Figure 18. Photomicrographs of petrographic features of the Allentown Dolomite. A, Aphanocrystalline dolomite in clasts of desiccation dolorudite (edgewise conglomerate); plane light, 35X. B, "Birdseye" mottling in aphanocrystalline dolomite; plane light, 100X.



A



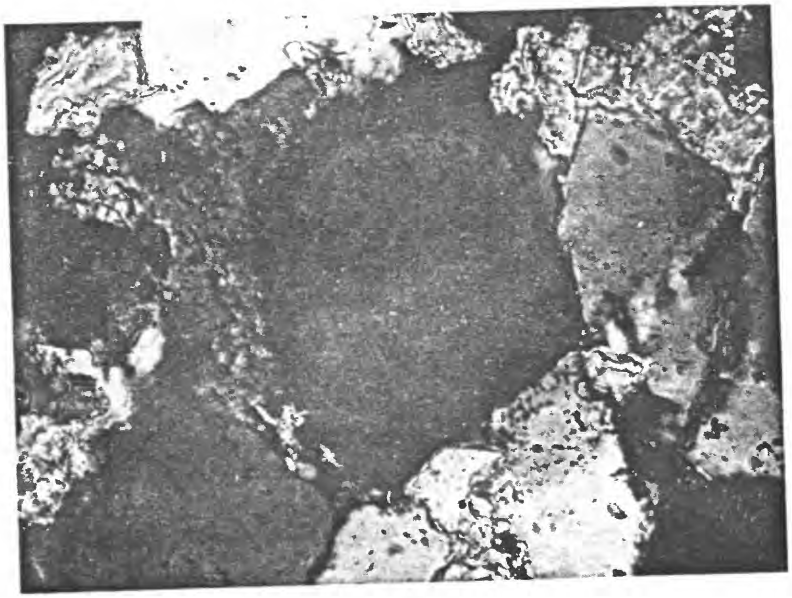
B

available to display it. Authigenic overgrowths in optical continuity, or quasi-continuity, with the host grain are common (fig. 19), and some show idiomorphic tendencies. There is no specific evidence of abraded overgrowths, but many grains, including overgrowths, have ragged edges owing to dolomite replacement. Roundness sorting is fair to good within thin sections but roundness varies widely between sections. Size sorting is fair to poor in dolorudites but fair to good in other dolomite lithotypes.

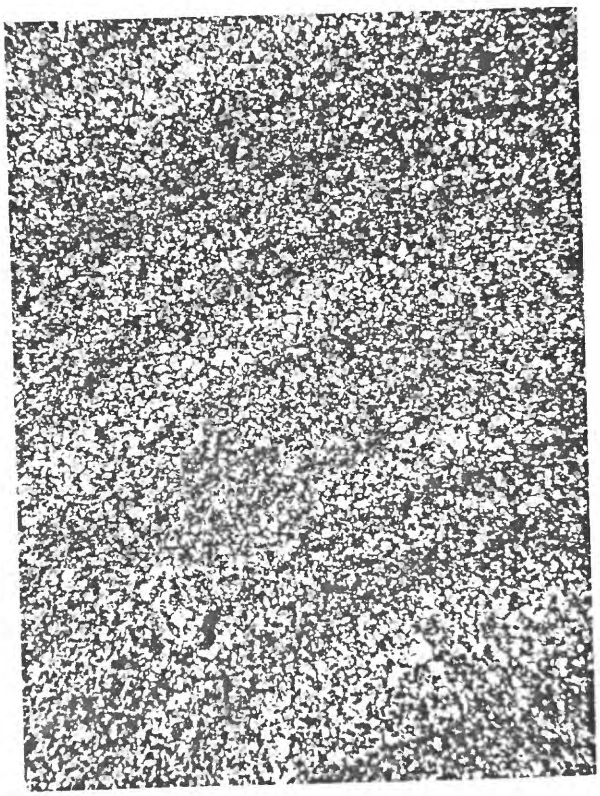
Bedded and nodular chert occurs mostly in the upper part of the Allentown, but thin sections show that chert is distributed throughout the unit. It forms small (0.1 mm, avg.) irregular authigenic patches and blebs within the dolomite mosaic, perhaps filling some of the void space created by dolomitization. Under the microscope all chert in the Allentown appears as a microcrystalline mosaic of anhedral quartz (fig. 20).

Feldspar is mostly microcline; mesoperthite is rare and plagiochase is very rare. Microcline generally is found wherever detrital quartz is concentrated, but never in amounts exceeding 1 percent (by volume) of the dolomite rocks. Where quartz is more abundant so is microcline. In orthoquartzitic rocks, microcline composes 5-10 percent of the rock. Detrital grains of microcline are angular to subround and range in size from 0.08 mm to 0.5 mm, averaging about 0.2 mm. Authigenic overgrowths are common and neither grains nor overgrowths show any tendency to be replaced by dolomite. Nearly all microcline is very fresh. The rare

Figure 19. Photomicrographs of petrographic features of the Allentown Dolomite. A, Authigenic overgrowth of detrital quartz grain. Polarized light, 100X. B, Petrographic detail of algal stromatolite in fine grained dolomite. Plane light, 35X.

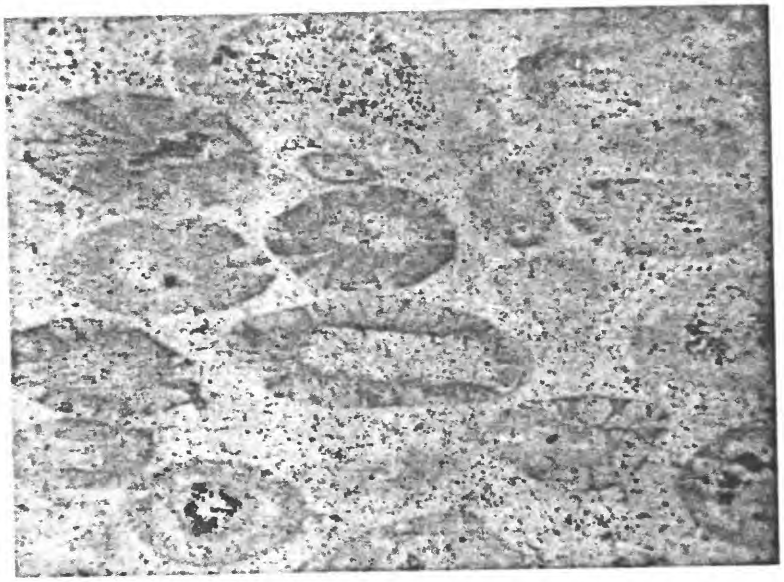


A

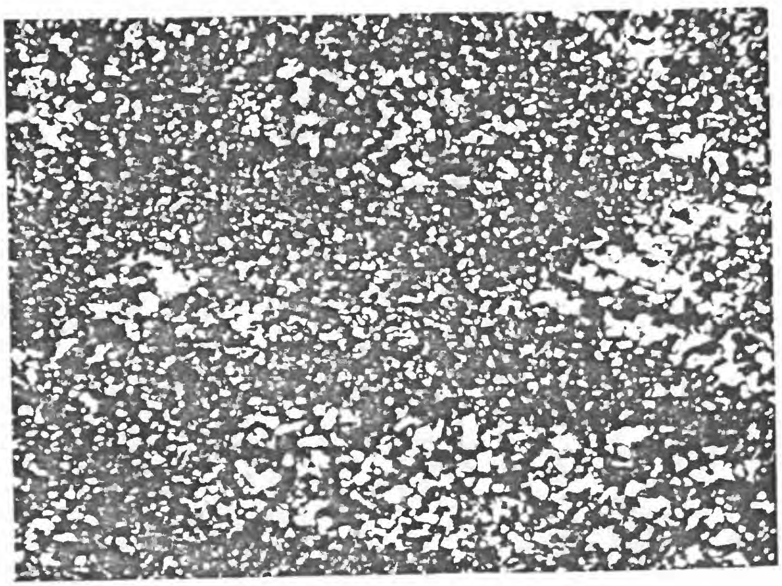


B

Figure 20. Photomicrographs of petrographic features of the Allentown Dolomite. A, Ghosts of deformed oöids in chert. Note radial and concentric structures; plane light, 35X. B, Same, under crossed polarizers. Field includes only the two oöids in center of above photo; 250X.



A



B

mesoperthite grains are only slightly altered differentially in the manner described previously.

Pyrite is widely scattered throughout all rock types, all samples, and all thin sections, but in all cases constitutes much less than 1 percent of the sample. It occurs as small irregular grains, cubes, or partial cubes that range in size from 0.002 mm to 0.2 mm. Smaller grains are somewhat altered to iron oxides but larger grains are pristine. All pyrite clearly is authigenic.

Monazite, tourmaline, and zircon occur rarely in thin sections. Where found, they are typically associated with concentrations of detrital quartz. All are fine (<0.06 mm), detrital, and subround to round. Some monazite is very angular, however.

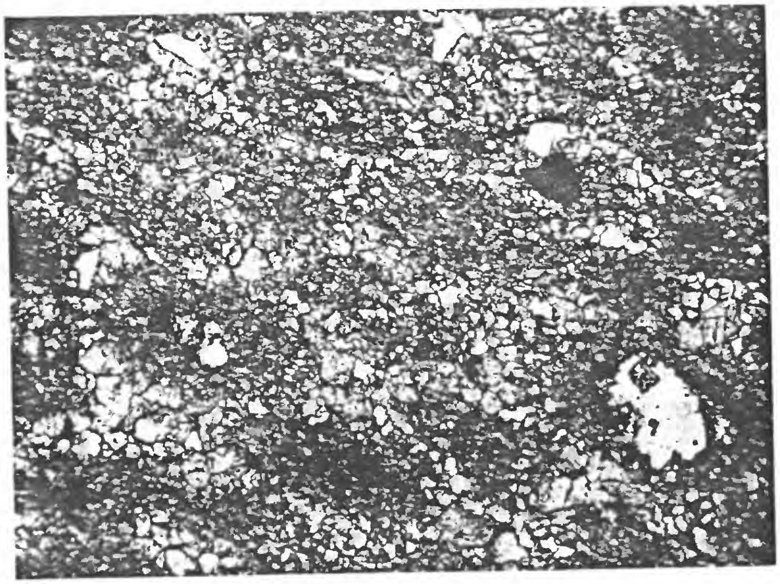
In addition to all of the above (except dolomite), the following minerals were observed in finer fractions (<0.06 mm) of insoluble residues: rutile, corundum, garnet, and topaz. All are detrital and are subangular to subround.

Texture

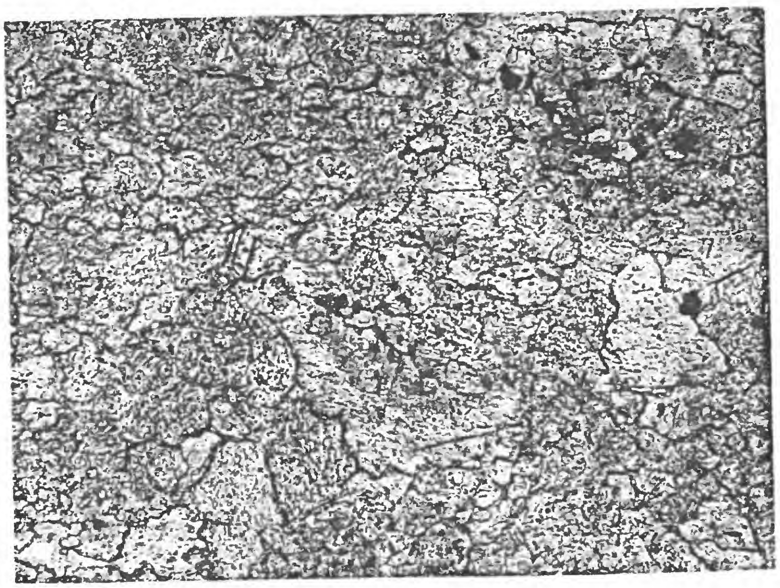
Texturally all dolomites of the Allentown are mosaics of anhedral to subhedral sparry and microcrystalline mineral dolomite (figs. 17 and 18). Regardless of the quantity and type of palimpsest relics present, petrographically these dolomites have crystallization fabrics that are equigranular xenotopic to hypidiotopic. The crystal size variation within a single thin section generally is small.

Features such as oöids, pellets, and intraclasts appear as "ghosts" in the dolomite mosaic (fig. 21). Similarly preserved are

Figure 21. Photomicrographs of some petrographic features of the Allentown Dolomite. A, Ghosts of deformed oöids in dolomite mosaic. Plane light, 35X. B, Dolomite crystal outlines transecting boundaries of allochem ghosts. Plane light, 100X.



A



B

ripple marks, graded bedding, and cross stratification. None of these features have any other petrographic expression where they are composed of dolomite. Algal stromatolites may show a fine lamination, but they are entirely dolomitic (fig. 19). Boundaries of oölites, pellets, and intraclasts commonly are transected by dolomite crystals and entire allochems may be completely enclosed within a single dolomite crystal (fig. 21). Only the very fine grained to aphanocrystalline dolomites and the algal stromatolite rocks lack the above allochemical and current derived features.

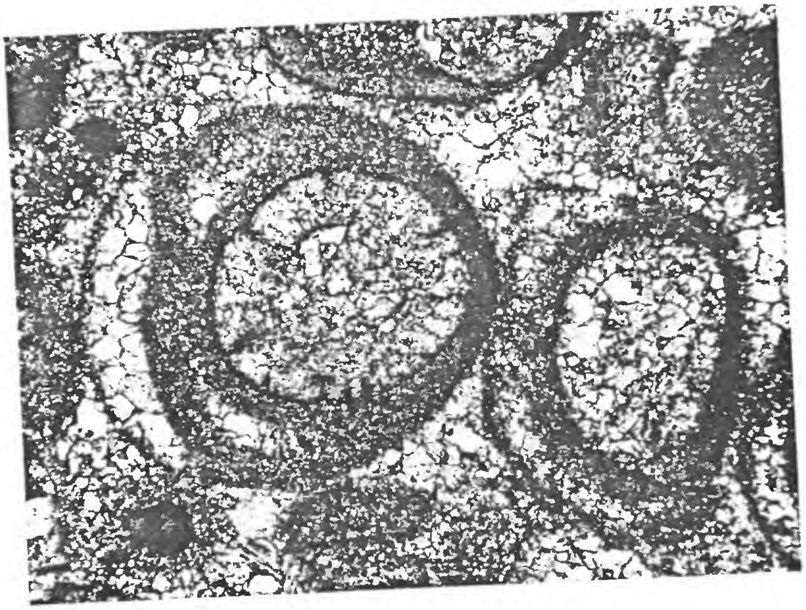
Eccentric Oöids

Petrography

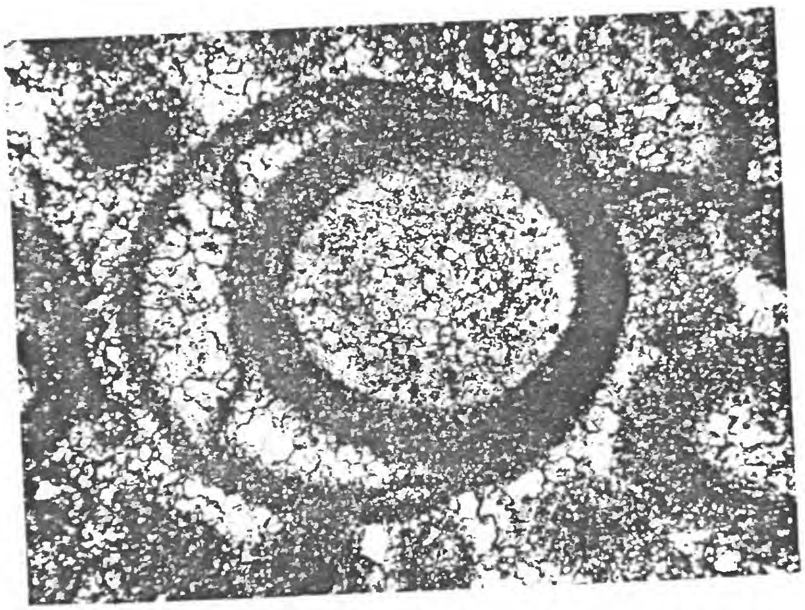
Some oölitic dolomites in the Allentown contain oöids with very unusual structural configurations. They are peculiar because the internal shells that compose them are eccentric relative to each other and to the core of the structure (figs. 22 and 23). In cross-sections perpendicular to bedding, the inner shells and the core are apparently displaced stratigraphically downward relative to the outer shell. The shells are separated by lunate areas (in thin section) that are convex toward the top of the bed. These peculiar features are herein termed eccentric oöids. In many of these oöids parts of the inner shells have collapsed toward the core of the structure.

Eccentric oöids are best observed under the microscope. However they are large and distinct and the eccentricity can be seen in hand specimens and outcrops as well, and may therefore be used in mapping as top and bottom indicators. Unfortunately they are much

Figure 22. Photomicrographs of eccentric oöids in the Allentown Dolomite. _A_, Oöid with wall of outer shell partially collapsed onto the core. Plane light, 35X. _B_, Oöid with most of outer shell collapsed onto the core. Note the thin wall of outer shell that remained intact and essentially spherical after rest of wall collapsed. Plane light, 35X. In both photos note the asymmetry of the thickness of the shells in the top and bottom of the structure.

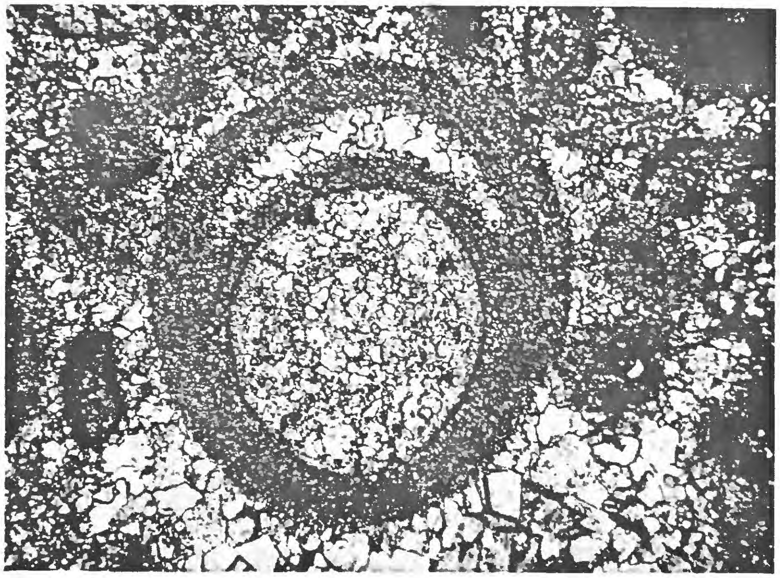


B



A

Figure 23. Photomicrographs of eccentric oöids in the Allentown Dolomite. A, Thick-shelled oöid showing eccentric structure and one lunate area at top of structure. Note that total shell thickness at the top is much greater than at the bottom.
B, Single-shelled eccentric oöids in which no lunate areas have developed. Eccentricity is expressed by the shell thickness, which is much greater at the top than at the bottom.



A

B

too rare to be of much practical value in this respect. They do, however, illuminate aspects of the depositional and diagenetic history of the dolomitic rocks containing them.

Eccentric oöids in the Allentown typically range from 1 to 2.5 mm in diameter and are among the largest oöids in the unit. They are composed entirely of the mineral dolomite that occurs in three basic textures, which systematically relate to three structural elements. These elements are 1) the core, 2) one to three inner shells and/or an outer shell, and 3) lunate areas that separate inner shells from each other and from the outer shell. Each element in the structure has a characteristic texture.

The core, as distinguished from the nucleus, is the innermost or central spheroid of the structure. The nucleus is the particle around which the core developed, commonly a dolomite crystal. The cores typically are spherical (circular in thin section) to slightly elongate in a direction perpendicular to bedding. In the elongate oöids, the long diameter is up to 1.2 times longer than the short diameter. The average diameter of the cores is about 0.8 mm, and most are in the range from 0.65 to 1.0 mm. Rarely are they squashed or ruptured. Texturally the cores are composed mostly of crystals of subhedral and anhedral somewhat cloudy dolomite that ranges from 0.02 to 0.06 mm in diameter. Ghosts of radial and concentric structures are clearly visible. In general the cores of the eccentric oöids are very similar in petrographic expression to the normal oöids in the Allentown.

Surrounding the core are from one to three shells composed of very finely crystalline to aphanocrystalline anhedral dolomite that contrasts sharply with the core, the lunate areas, and the enclosing matrix (figs. 22 and 23). The shells vary from 0.05 to 0.4 mm in thickness. In oöids with more than one shell, the aggregate thickness of shells around the top of the structure exceeds by 1.3 to 2 times the thickness of shell material at the bottom of the structure where the shells merge into a single unit. Many oöids that have only one shell surrounding the core are markedly asymmetrical from top to bottom (fig. 23). Shell thickness at the top of such oöids may exceed that at the bottom by as much as three times. Unlike the cores, no relic textures or structures are visible in the shells. In most oöids that have more than one shell the upper parts of some inner shells or the inner walls of such shells have collapsed toward the core (fig. 22). Part of the outer shell may be collapsed also, but the outer wall remains intact and essentially spherical.

The crescent-shaped areas between the shells are composed of sparry anhedral dolomite that averages 0.11 mm in size and is the coarsest dolomite in the structure. Its clarity contrasts rather sharply with the somewhat cloudy dolomite in the cores. No textural relics are visible in these areas. The number of such areas that can form obviously depends on the number of shells in the structure. No crescents form in oöids with only one shell surrounding the core. Where shells have collapsed, the shape of the coarse-grained crescentic areas may be quite complex because they are controlled by the form taken by the collapsed shell.

The matrix enclosing the eccentric oöids is a poorly sorted mixture of aphanocrystalline to very finely crystalline dolomite intraclasts, fine to medium crystalline dolomite intraclasts, and fine to medium crystalline anhedral and subhedral dolomite. Most intraclasts have a coating or shell of aphanocrystalline dolomite that is similar in all respects to the shell material of the eccentric oöids. This coating or shell is quite variable in thickness. Its generation clearly is a separate depositional event even on aphanocrystalline intraclasts of essentially similar texture. Oöids and intraclasts are fairly closely spaced in the matrix, but not so close that each particle is in contact with all of its neighbors. Many are not in contact with any other oöid or intraclast and are completely engulfed by matrix. Contacts between oöids and matrix typically are very sharp in terms of grain size, but locally may be somewhat fuzzy and indistinct.

Unlike other oöids in the Allentown, the eccentric oöids are true oöids, not palimpsest relics. They actually exist as distinct petrographic entities. The only relic element in the structure is the radial or concentric structure of the core itself.

Unusual oöids from the Allentown Dolomite previously have been reported on by Wherry (1916) and Carozzi (1963). Their descriptions concern what Wherry termed "half-moon oölites," which are somewhat different from though probably related to the eccentric oöids described herein.

Half-moon oöids, according to Wherry (1916), are circular in cross-section and are divided inside, parallel to bedding, into a light and dark portion. The dark portion occupies the lower part of the oöid and the light portion above generally is the larger of the areas (in cross-section). The dividing line between the two parts is convex toward the top, but in some cases may be straight across. The two parts are similar in grain size, both being coarser than the groundmass.

In Wherry's (1916) interpretation, the half-moon oöids originally were composed of concentric shells of aragonite that contained carbonaceous material of organic origin. Groundwater entering the oölitic rock along joints and bedding planes dissolved the aragonite and left behind the core and the carbonaceous material, both of which settled to the bottom of the spherical cavities. Secondary dolomite later filled the remaining space in the upper part of the cavities and now constitutes the lighter areas that are on top of the darker carbonaceous portions of the structures.

Carozzi (1963) described similar dolomitic oöids in a much better state of preservation in samples from the Carpentersville section, Warren County, New Jersey. He classified half-moon oöids according to structural complexity, based on the number of internal shells. All are of the same general type, however, and are characterized by 1) spherical outer shells of variable thickness that maintain the shape irrespective of internal configurations, 2) sets of internal concentric shells in various stages of collapse and deformation, and 3) cores that consist of the innermost parts of the concentric shell structure, which have settled to the bottom of the structure, and which are themselves flattened and deformed. The nuclei typically are small carbonate crystals (now dolomite). Generally shell thickness at the bottom of the structure is equal to that at the top, except, as Carozzi (1963) noted, in some cases where a given set of concentric rings appears thinned when traced underneath the flattened core as if the weight of the latter had partially squeezed it.

In Carozzi's (1963) view, half-moon oöids originated by the diagenetic alteration of normal concentrically-structured oöids. The perfect preservation of the internal structure of the collapsed rings (shells), he feels, indicates that the original mineral composition was calcite, plus another mineral that was dissolved in diagenesis. The removal of this other mineral, which he thinks was probably anhydrite or gypsum, created space that allowed the soft calcitic internal laminae to collapse and deform plastically. These changes took place, according to Carozzi, very soon after deposition, before compaction was completed and before dolomitization had started.

Interpretation

Questions raised by eccentric oöids in the Allentown Dolomite concern whether or not they are essentially depositional features or the result of diagenetic alteration, and whether they grew as normal oöids by rolling in agitated carbonate saturated marine waters, or whether they formed <u>in situ</u>. Other questions concern the nature and origin of the dolomite and possible other material that was present in the structure, and its effect on the development of the oöids.

Some conclusions are reasonably certain. The oöids must have developed their eccentric structure in place, not by rolling, because the eccentricity bears a consistent relationship to bedding. The medium-crystalline lunate areas between aphanocrystalline shells must at sometime have been open space (not filled with carbonate or other minerals) for otherwise it would not have been possible for inner shells and shell walls to collapse. Filling of this space by dolomite clearly occurred after the shell collapsed. The empty space must have represented material that was deposited in the structure originally and later removed in some fashion.

The eccentric oöids described herein differ from the half-moon oöids of Wherry (1916) and Carozzi (1963) chiefly in the nature and behavior of the core. In half-moon oöids the core is no more than the central part of the concentrically laminated structure and is similar in all discernible respects to the outer collapsed shells of the structure. The entire half-moon oöid represents an essentially continuous depositional episode. Typically, these cores are flattened parallel to bedding owing, in all probability, to its similarity in texture and composition to the outer shells. Both core and shells must have been soft and plastic, whereas the undeformed outermost shell had attained a measure of strength. The size of the core evidently was determined by the number and thickness of calcium carbonate shells that developed before the other material in the concentric structure, presumably gypsum or anhydrite, began to precipitate.

In contrast, the cores of the eccentric oöids are entirely different in texture from the surrounding shells and very rarely are they flattened or otherwise deformed. They remain generally spherical regardless of the number of surrounding shells or the complexity of shell collapse and deformation. Clearly these cores had developed some degree of strength at the time shell collapse occurred, and for this reason they behaved as rigid bodies. The size, size sorting, and microscopic expression of these cores strongly indicates that they are normal Allentown oöids around which the aphanocrystalline shells were deposited, but, unlike the cores of half-moon oöids, may represent an entirely different depositional episode than the shells. They are composed of replacement dolomite but the evidence is not clear as to whether replacement occurred before or after the shells formed.

The development of the eccentric structure can be viewed in three ways. One is that some material that originally was deposited in the eccentric structure of normal oöids was selectively removed at a later time and thus allowed the remaining shells to settle into contact at the bottom of the structure and created crescent-shaped spaces between the shells at the top of the structure. The second is that the oöids grew eccentrically in place with material being added to the structure in crescentic segments. Material in some segments was selectively removed to create the void space at the top, which allowed the upper shells or shell walls to collapse, and which was later filled with dolomite. The third is that the oöids never had another mineral in the structure, but consisted of a large rigid core surrounded by shells of aphanocrystalline dolomite. The core sank through the bottom of the shells before they were solidified and created space at the top.

The third possibility is most easily disposed of. In some oöids with only one or two shells the lunate space above is small enough and the discrepancy in thicknesses large enough to make such an explanation plausible. However in the larger and more complex structures, there is no way in which the space above can be accounted for by sinking of the core, assuming that such a phenomenon could have occurred.

It is difficult to choose between the first two hypotheses on the basis of the evidence at hand. If the eccentric oöids formed initially as normal concentrically-structured oöids, the shell thickness at the bottom of the structure where all shells are in contact with each other ought to be very nearly equal to the total shell thickness (excluding the lunate areas) at the top of the structure. Such is not the case, however, as the shell thickness at the bottom is consistently and significantly less. It is conceivable that such thinning may be the result of downward squeezing by the weight of the core and other shells, as Carozzi (1963) postulated in a few of his half-moon structures. However, he was able to trace individual shells around the bottom of the core where the shells thinned. This cannot be done with the eccentric oöids. In no case can a given shell, other than the innermost shell surrounding the core, be traced with any degree of certainty around the bottom of the core. Such downward sinking of the core would indeed contribute to the void space at the top of the structure, however. Eccentric oöids with a single shell around the core appear to argue against such sinking. In many of these the top of the shell is markedly thicker than the bottom (fig. 23), but it is clear that the core, if it sank, did not pull away from the inner wall of the shell at the top, as might be expected, nor is there any evidence of shell distortion at the top, bottom, or sides, of the structure to indicate that the upper part of the shell was pulled down with the core or that the lower part was pushed aside or squeezed. Indeed, in one of these oöids (fig. 23, bottom) the single shell does seem to be composed of two depositional increments, one of which is decidedly crescentic. Thus it would appear that eccentric growth *in situ* may have occurred in some oöids. The general application of this hypothesis is not without problems however.

In oöids that contain more than one inner shell the shell thicknesses do not vary as markedly as in the single shell structures. It is not clear why this should be so if the single shell and multishell oöids have the same origin, especially if they grew in place. Also it is difficult to conceive of an *in situ* process that could produce such nearly spherical, well-rounded shells.

These oöids are unusual for a reason other than their eccentric structure. These are the only true oöids in the Allentown Dolomite, and in all of the dolomitic rocks of the Lehigh Valley. All other oöids that the writer has observed are entirely of replacement origin and only palimpsest relics of the original radial and concentric structures remain in the replacement dolomite fabric.

As previously noted, the cores appear to be such replaced normal oöids. Petrographically the cores do not differ from other Allentown oöids. The very finely crystalline to aphanocrystalline shells do not contain textural relics, however, or any other evidence of replacement, and may represent deposition of primary dolomite or incorporation of syngenetically replaced dolomite in the oöid structure. Such an interpretation assumes that the fine grain size is not the result of degenerative recrystallization, for which there is no evidence in any of the dolomitic rocks of this study. Indeed, the presence of such essentially primary dolomite may be the reason that these features have the expression of true concentrically-structured (now eccentric) oöids. Dolomite that was deposited in the structure originally, could have been unaffected by subsequent dolomitization of the surrounding sediments.

The medium crystalline lunate areas represent material removed from the structure. What the material might have been is largely a matter of conjecture, as no traces of the mineral remain. However if the interpretation of the aphanocrystalline shells as essentially primary dolomite is correct, it is reasonable that the missing mineral was an evaporite more soluble than dolomite. Gypsum or anhydrite are obvious possibilities because they immediately follow dolomite in the normal precipitation sequence of marine evaporites. Slight fluctuations in water temperatures or salinity would have determined which mineral was deposited. The sequence apparently ended with dolomite deposition because all outer shells are dolomitic, but outer shells of gypsum or anhydrite could have been removed by solution.

Solution of shells or crescentic areas of calcium sulfate to create the void space now represented by medium crystalline dolomite must have occurred while the inner shell walls were still soft and permeable, but the outer shell walls were rigid enough to maintain sphericity. The plastic behavior of the inner shells and the invariable relation of eccentricity to bedding suggest that solution and shell collapse occurred very early in the history of the sediment, almost certainly in the depositional environment, and probably very soon after the oöids were formed.

It is difficult to compare in detail the eccentric oöids, described herein, and the half-moon oöids of Carozzi (1963), because he does not discuss or describe the dolomite textures of the oöids or the enclosing rock, except to point out that the rocks are oölitic dolarenites. Moreover, his photomicrographs were prepared in a way that admirably displays the shell structure but suppresses the effects of grain boundaries. However, his photomicrographs and his assertive remarks about the state of oöid preservation suggest that they too are rather unusual oöids for the Allentown, aside from their internal deformation. They do not appear to be ghosts in a replacement dolomite fabric. They may be true oöids that formed in the normal concentric fashion with shells of dolomite (primary or syngenetic replacement) and anhydrite or gypsum. Instead of forming around other oöids, as eccentric oöids appear to have done, half-moon oöids formed around nuclei of tiny carbonate crystals. The writer feels that the alleged perfect preservation of the internal structure of the collapsed shells does not necessarily indicate that these oöids originally were calcite, as Carozzi (1963) believes, and in fact such a degree of preservation, in the writer's view, would argue against such an origin. Such an origin implies that dolomite replaced the calcite after the oöids were formed. All other dolomitic oöids in the Allentown clearly are of such replacement origin, and none have the degree of preservation or petrographic expression of the half-moon oöids. The same would be true if the oöids were originally aragonitic. The unusual expression of both half-moon and eccentric oöids is much better explained if dolomite were incorporated into the structures initially. The presence of such essentially primary dolomite is wholly consistent with the deposition of anhydrite or gypsum in the structure, as both Carozzi (1963) and the writer believe, because both are sequential members of the normal marine evaporite sequence. Gypsum or anhydrite follow dolomite in the sequence, at higher salinities.

The question of the origin of the eccentricity of these oöids remains, for now, unanswered. The writer would favor the hypothesis of diagenetic alteration by solution of calcium sulfate shells in normal concentric oöids, were it not for the great differences in shell thickness, from top to bottom, of eccentric oöids. Perhaps this can be accounted for by sinking of the large, rigid core into the shells at the bottom of the structure, as has been previously discussed, but if so, the single-shell eccentric structures would seem to have a different origin, perhaps eccentric growth _in situ_.

Cyclicity

Reference previously has been made to the conspicuous cyclicity of the Allentown. To an extent, all of the lower Paleozoic carbonate units described herein have cyclic aspects at some level of generalization, as do related rocks throughout the Appalachians and other lower Paleozoic depositional areas on the periphery of the North American shield. Although the nature and significance of the cyclicity of some of these rocks are rather obscure, the character of the repetitive lithotypes in the Allentown is highly distinctive and bears important evidence concerning their origin.

The character and general sequence of rock types in Allentown cycles are discernible in most of the larger natural exposures and highway and railroad cuts. Zadnik (1960) studied two long sections along the Delaware River in New Jersey and was the first to systematize the lithic successions in terms of an ideal cycle. His work was based entirely on A. V. Carozzi's (1958) method of treating measurements of sizes and frequencies of detrital, authigenic, and organic components of each bed in the succession and interpreting the results in terms of bathymetry and current intensity. Zadnik measured the degree of crystallinity (the average size of the larger dolomite crystals), the clasticity (size) of reworked carbonate fragments ("general clasticity"), the average size of larger quartz grains and frequency of quartz grains, the average size of the larger oölites ("oölite clasticity") and frequency of oölites, and the frequency of pyrite grains. The resulting variations were expressed as continuous smooth curves on parallel scales and were interpreted on a similarly parallel scale by means of a bathymetrical curve expressing variation of the relative depth of deposition assigned to each lithotype in the ideal cycle.

According to Zadnik (1960) variations of the above parameters define and limit cyclic trends in the Allentown. In fact, however, most of the significant variations that appear in Zadnik's measurements and curves are inherent in the field classification of the rocks themselves and have little or no independent petrographic significance in defining cycles. For instance, it is quite reasonable to expect that oölite clasticity and frequency would be greatest in oölitic rocks, especially since some lithotypes contain few, if any oölites. Although many Allentown rocks contain oölites that can be detected only under the microscope, Zadnik's study clearly shows that the only significant concentrations are those that are easily observable in the field. Similarly, a flat pebble dolorudite ("edgewise" or "desiccation dolorudite") ought to have a significantly greater general clasticity than other rocks containing much smaller lithic fragments because otherwise it would not be a rudite. Moreover, one would expect that currents capable of transporting large intraclasts also could transport more and larger quartz grains if they were available. Unfortunately, Zadnik's study begins and ends with variation curves of parameters such as these and never confronts real questions of process, environment, and the immediate or ultimate origins of the rocks or the cycles. Undoubtedly Zadnik's work provides some objective criteria for describing these carbonate cycles, but bathymetry alone is a much too limited, and potentially misleading, basis on which to frame the results, however correct they may be.

In essentials, the writer agrees with Zadnik's (1960) elucidation of the succession of dolomite lithotypes that constitute the ideal cycle, and to a limited extent with the bathymetric conditions that they are presumed to represent. In Zadnik's view the ideal cycle consists of the following six lithotypes, from bottom to top: 1) dololutite, 2) dolarenite, 3) oölitic dolarenite, 4) dolorudite, 5) cryptozoan dolomite, and 6) desiccation dolorudite (tab. 9). These are taken to represent, from bottom to top, progressively shallowing depositional conditions. The term "shallowing" is strictly relative as all lithofacies probably originated in rather shallow water. The environment undoubtedly was marine, though Zadnik (1960) does not commit himself on that point.

Table 9. Succession of dolomite lithotypes in the ideal cycle, as interpreted by Zadnik (1960)

Desiccation dolorudite (top)
Cryptozoan dolomite
Dolorudite
Oölitic dolarenite
Dolarenite
Structureless dololutite (base)

The writer disagrees with Zadnik's conception of the cycle boundaries, particularly with the place of his desiccation dolorudite. In terms of geologic processes, the natural cycle boundary would seem to be not above the desiccation dolorudite, but below it. This lithotype does not represent the ultimate shallowing stage of the cycle but rather the destruction of that stage by submergence and exposure to stronger marine currents. The shallowing stage itself is represented by the flat, angular clasts of aphanocrystalline dolomite that commonly compose the dolorudite. The writer feels that this material is primary or syngenetically replaced dolomite that was precipitated under essentially evaporitic conditions that probably accompanied this stage.

Submergence after the evaporitic or desiccation stage typically was accompanied by a strong influx of quartz grains, sometimes coarse and sometimes fine. Such an influx of terrigenous material would not be expected under the regimen of extremely shallow water with restricted circulation that would be required to produce the hypersaline evaporitic conditions that generated what the writer feels is probably primary or syngenetically replaced dolomite that occurs as clasts in these edgewise dolorudites. Although influxes of quartz occur in other parts of the cycle, the association with a particular lithotype is not nearly as strong as with the edgewise ("desiccation") dolorudite. However, the size and quantity of quartz in these dolorudites is difficult to reconcile with the probably slight changes in slope and water depth that accompanied shifting depositional environments in shallow water. One possibility is that the shallowing evaporitic desiccation stage also exposed areas where quartzose sediments were being deposited. These could then have been affected by wind action that moved small amounts of quartz sand into areas of dominantly carbonate deposition, and which was subsequently incorporated in the dolorudites. In any case the regular introduction of quartz at this stage, especially in association with rudites seems inconsistent with the interpretation of this lithotype as the ultimate shallowing stage of the cycle.

The desiccation or flat-pebble dolorudites herein are interpreted as the base of the ideal cycle (fig. 24) and represent the disruption of sediment marking the shallowing evaporitic desiccation stage and the transition to the relatively deeper, quiter water represented by the superjacent dololutite. The base of this dololutite represents the most fundamental change in the driving energy of both the chemical and mechanical systems involved in the formation of these rocks, at least to the extent that these systems are manifested in observable rock properties. The dololutite is the deepest water facies in this interpretation, as it is in Zadnik's (1960). The succeeding facies no doubt formed in shallower water, but there is little specific evidence on which to base an assertation that the changes in bathymetry were necessarily progressively shallower, although they may have been.

Chemical Composition

Twelve new chemical and semiquantitative spectrographic analyses of the Allentown Dolomite are presented in Table 10.

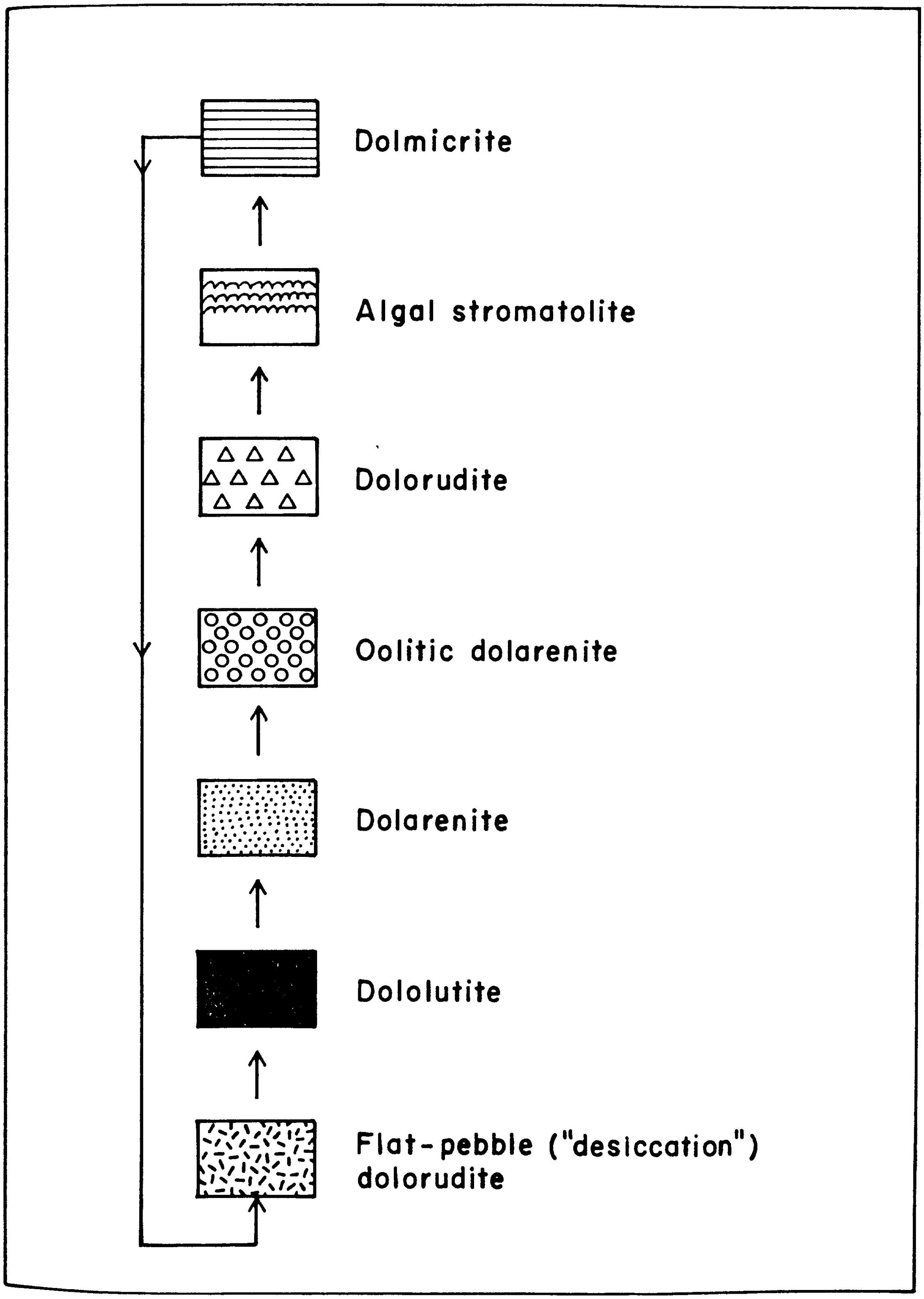


Figure 24. Column showing the succession of dolomite lithotypes in the ideal cycle in the Allentown Dolomite. The presence of dolmicrite at the top of cycle is inferred from clasts in the flat-pebble dolorudite, and does not occur as a discreet stratigraphic unit in the field.

Table 10. Chemical analyses of the Allentown Dolomite

	1	2	3	4	5	6
SiO_2	15.2	3.7	3.4	2.8	2.0	1.8
Al_2O_3	2.2	1.0	.41	.66	.23	.60
Fe_2O_3	--	--	--	.16	--	--
FeO	.60	.50	.40	.36	.65	.40
MgO	19.8	20.8	21.1	20.5	20.8	21.2
CaO	20.7	27.8	28.5	28.3	29.0	28.7
Na_2O	.21	.09	.07	.20	.08	.06
K_2O	1.8	.58	.33	.61	.20	.27
H_2O-	--	--	--	--	.02	--
H_2O+	.62	.74	.66	.56	.62	.74
TiO_2	.10	.04	.03	.04	--	.02
P_2O_5	.04	.02	--	.02	.02	.02
MnO	.03	.03	.03	.03	.02	.02
CO_2	38.3	44.5	45.1	45.4	46.1	45.9
S (as SO_3)	.38	--	--	--	--	--
C (organic)	.096	.063	.043	.069	.028	.10
	100.08	99.86	100.07	100.01	99.77	99.83

Table 10. Continued

	7	8	9	10	11	12
SiO_2	1.2	15.4	2.3	2.6	3.8	3.4
Al_2O_3	.46	2.3	.52	.70	.74	.56
Fe_2O_3	--	.36	--	--	.21	--
FeO	.46	.40	.42	.42	.32	.40
MgO	22.0	13.8	20.9	21.6	20.8	21.1
CaO	28.3	29.2	28.3	27.8	27.7	28.2
Na_2O	.02	.07	.15	.05	.15	.10
K_2O	.20	.19	.46	.40	.75	.35
H_2O-	--	--	.02	.02	.03	.02
H_2O+	.62	.74	.34	.40	.22	.62
TiO_2	.02	.17	.02	.05	.06	.02
P_2O_5	--	.06	.02	.02	.05	.03
MnO	.03	.03	.02	.03	.05	.03
CO_2	46.4	37.0	46.5	45.5	45.1	45.0
S (as SO_3)	--	.25	.11	--	--	--
C (organic)	.051	.050	.047	.061	.026	.058
	99.76	100.02	100.13	99.65	100.01	99.89

Samples in Table 10 are as follows:

1. Dololutite; field no. ACZ-1A.
2. Dololutite; field no. ACZ-1B.
3. Dolarenite; field no. ACZ-2A.
4. Dolarenite; field no. ACZ-2B.
5. Oölitic dolarenite; field no. ACZ-3B.
6. Oölitic dolarenite; field no. ACZ-3C.
7. Dolorudite; field no. ACZ-4A.
8. Dolorudite; field no. ACZ-4B.
9. Algal stromatolite; field no. ACZ-5A.
10. Algal stromatolite; field no. ACZ-5B.
11. Desiccation dolorudite; field no. ACZ-6A.
12. Desiccation dolorudite; field no. ACZ-6B.

All samples collected in exposures around the 13th street intersection of Rt. 22, city of Easton, Pa., and analyzed in laboratories of the U. S. Geological Survey, Washington, D. C. Analysts: Paul Elmore, L. Artis, G. Chloe, J. Kelsey, S. Botts, J. Glenn, H. Smith, Ellen Gray, and Philip Aruscavage.

Although published analyses of the Allentown are by no means uncommon, no previously published analyses were directed specifically at the cyclic aspect of these rocks. The table contains two analyses of each lithotype in the ideal cycle. These do not represent single complete cycles because single cycles that contain all six lithotypes in succession are extremely rare. These samples were collected as close together as possible, however.

The most striking feature of these analyses is the high degree of uniformity. In some the carbonate fraction approaches the composition of chemically pure dolomite. Figure 25 illustrates the point. The triangular diagram was constructed according to the method of Gault (1950) and Gray (1951), with modifications. In the diagram the upper angle of the triangle represents 100 percent calcium carbonate, the lower right angle 100 percent magnesium carbonate, and the lower left angle 100 percent noncarbonate material. Superimposed on the conventional ternary grid are lines denoting the percentage of dolomite that could be present in the carbonate fraction of the rock, based on the relative amounts of calcium and magnesium carbonate reported or computed from the analyses. The line labeled theoretical dolomite represents calcium and magnesium carbonate in the proportion in which they would occur in pure stoichiometric dolomite. Points falling on that line indicate that the entire ("100 percent") carbonate fraction of the rock analyzed may be pure dolomite. It is apparent that dolomites in the Allentown closely approach that composition, and where detrital impurities (noncarbonate

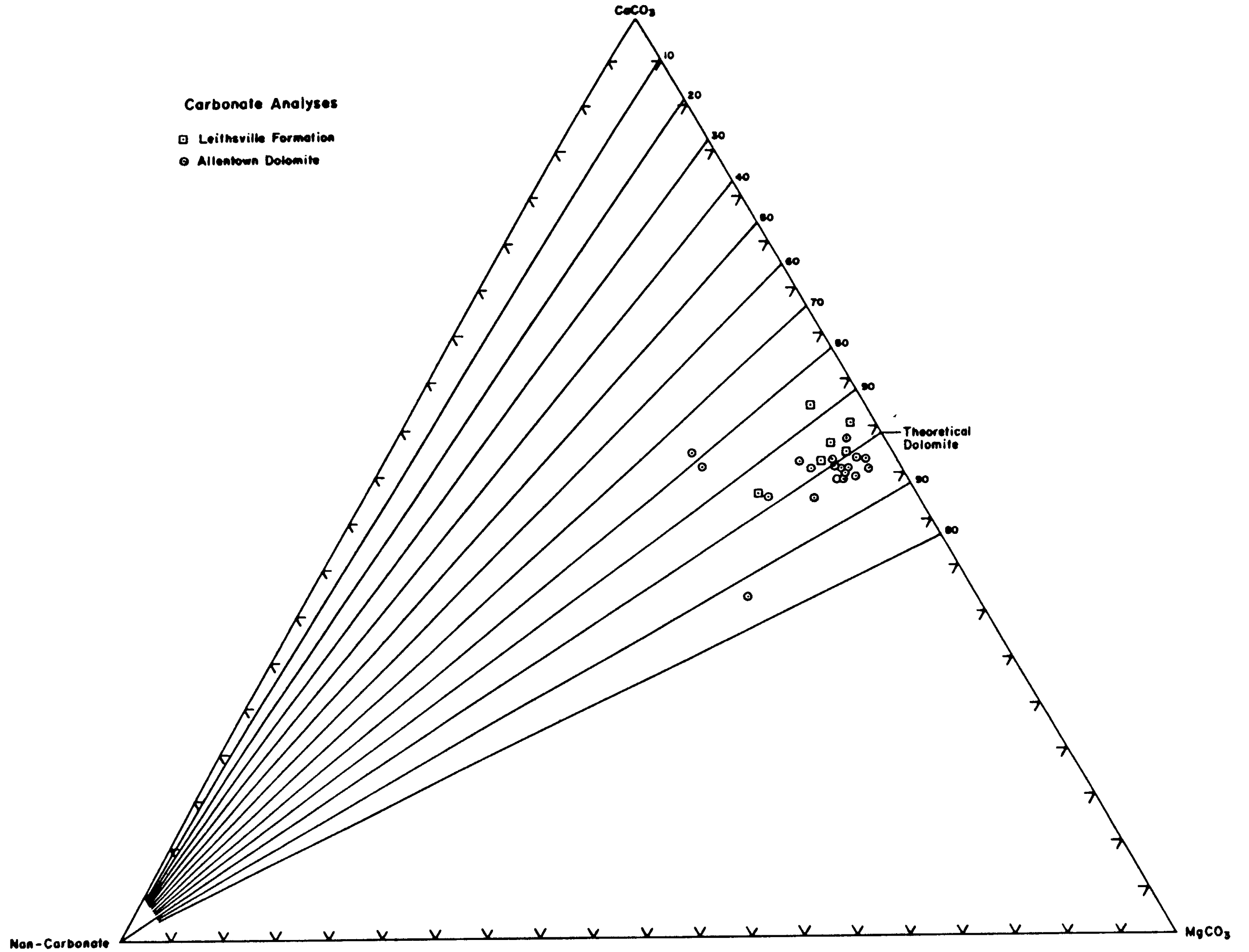


Figure 25. Triangular diagram showing chemical analyses of the Leithsville Formation and the Allentown Dolomite in relation to the composition of pure dolomite.

pole of the diagram) are few, the whole rock approaches that composition.

The abundance of silica is highly variable and mostly represents detrital quartz and feldspar. Feldspar, especially microcline, is present in minor amounts wherever quartz occurs, and the variation of Al_2O_3, Na_2O, and especially K_2O reflects this tendency. Where quartz is more abundant so is feldspar. The minor amounts of Al_2O_3, Na_2O, and K_2O that seem clearly to be associated with feldspars would suggest that clay minerals are relatively unimportant in these rocks.

There are, however, some subtle but discernible variations among the lithotypes of the ideal cycle. Figure 26 graphically shows the variation of organic carbon and ferrous oxide. Organic carbon was analyzed not only because it could conceivably show distinctive cyclic trends, but also because its variation could illuminate genetic aspects of the rocks and cycles. Under the mechanism proposed for the origin of rocks on the Allentown cycles, the latest stages of an ideal cycle would be characterized by increasingly hypersaline evaporitic conditions in shallow marine waters with restricted circulation. Such conditions would not be conducive to extensive organic activity, if any at all, and this fact might be reflected as a paucity of organic carbon in the resulting rocks. On the other hand such a deficiency also might originate in quite the opposite fashion. Abundant organic matter could easily be digested by bacteria that thrive in shallow, wave-agitated, well-oxygenated marine waters. Somewhat deeper poorly-oxygenated

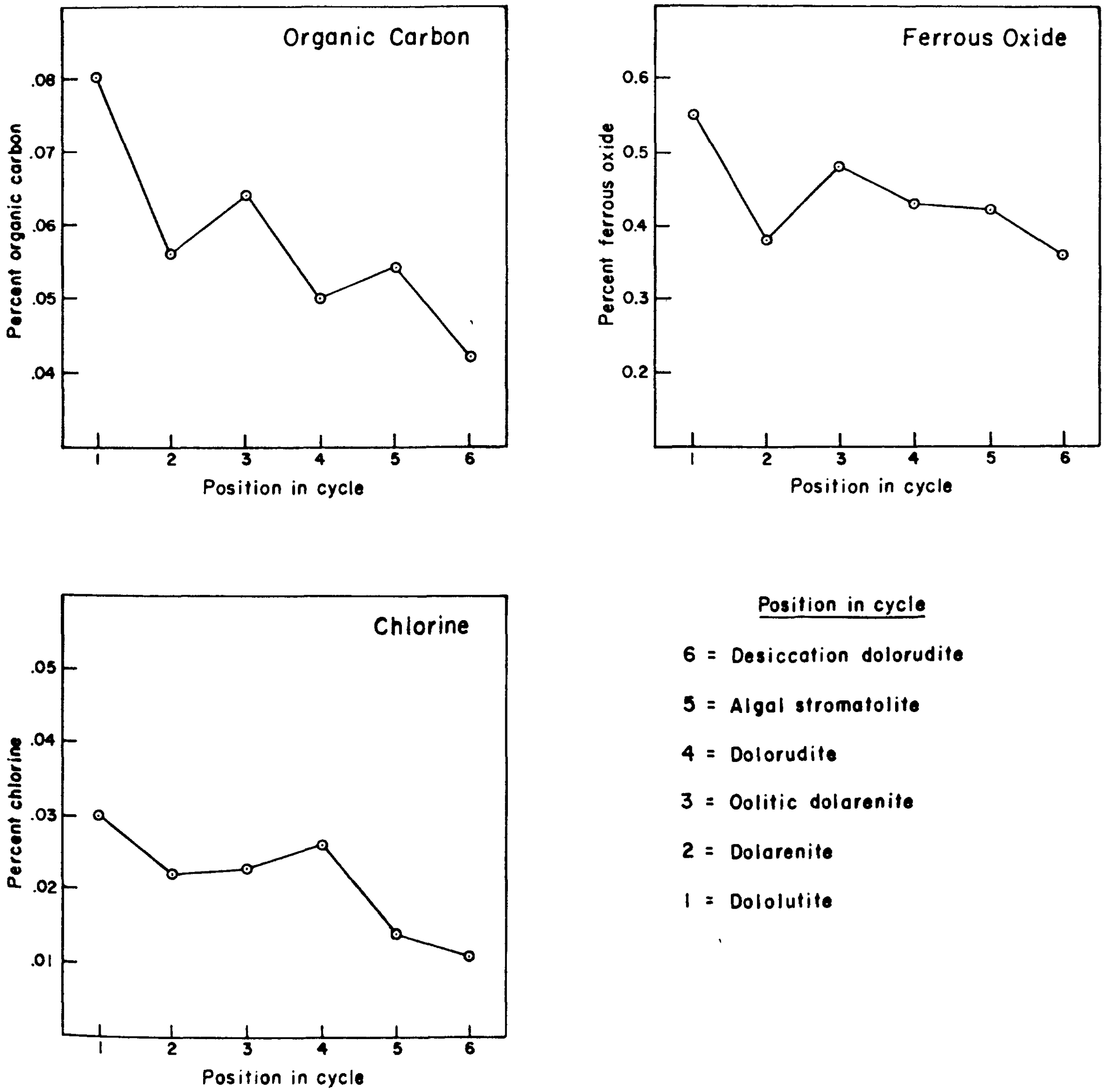


Figure 26. Graphs showing some chemical variations within rocks composing the ideal cycle in the Allentown Dolomite.

offshore marine waters, the depositional environment envisioned for the dololutite of the ideal cycle (fig. 25) could harbor reducing conditions that would allow organic matter to accumulate if the supply were great enough. Such conditions also would favor authigenic precipitation of pyrite.

Although two analyses of each lithotype hardly are definitive, they are, indeed, suggestive of the above trends. The average concentration of organic carbon is greatest in the dololutite and least in the desiccation dolorudite. Moreover, the average FeO content, indicative of the pyrite content, also is significantly greater in the dololutite than in any other rock type.

Although the graphs in Figure 26 superficially seem to suggest that the cycle boundary ought to be drawn above the desiccation dolorudite (position 1 in cycle), it must be borne in mind that analyses of that dolorudite are strongly influenced by the composition of the flat and angular aphanocrystalline dolomite clasts that compose it. Samples were chosen to reflect the composition of these particular clasts as much as possible, because logically they would seem to be the lithic manifestations of the terminal stage of the cycle. The analyses tend to support this concept and the interpretation of the rudites as the product of subsidence marking the beginning of a new cycle.

Organic carbon impurities commonly are cited as the coloring agent of dark colored limestones and dolomites, and differences of this sort have been used by some to explain the variegated nature of the Allentown in weathered exposures. However on freshly broken surfaces all dolomites in the Allentown are rather dark and the color differences are minor, to say the least.

Origin of the Allentown Dolomite

The relatively good exposure and distinctive lithologies of the Allentown Dolomite provide an excellent opportunity to illuminate genetic aspects of the unit. Its transitional contacts and close relation to the subjacent Leithsville Formation and superjacent Beekmantown Group should provide further understanding of these units as well. Pertinent aspects include the nature and extent of the depositional environment, nature of the pre-dolomitization sediment, chronology and processes of dolomitization, and the origin of significance of cyclicity.

The bulk of the Allentown consists of originally clastic calcium carbonate sediment that has been replaced by dolomite. The original carbonate textures occur only as palimpsest relics (ghosts) in the replacement dolomite mosaic, but their clastic heritage is clear. Replacement is amply demonstrated by the many typically calcium carbonate textures and structures that now are only ghosts in the replacement dolomite fabric. Dolomite crystals commonly transect the boundaries of these features.

Such dolomite would be considered "secondary" in the petrographic sense because dolomitization (replacement) occurred during or after consolidation of the sediment, and after calcium carbonate textures and structures (oöids, algal stromatolites, pellets, etc.) had formed. Conceptually, such replacement is a diagenetic process, even if it occurred penecontemporaneously in the basin of deposition.

Replacement appears to have been on a volume-for-volume basis. A 12 percent volume reduction would accompany molecule-for-molecule transformation of calcite to dolomite, and would increase porosity or cause brecciation of the rocks. No such phenomena occur in the Allentown.

Dolomite that possibly is essentially primary occurs in some oöids (eccentric variety) and as flat-pebble clasts in some dolorudites at the base of the Allentown cycle. The dolomite in these features typically is very-finely crystalline to aphanocrystalline (dolomicrite), may be finely laminated, may contain birdseye structures, and entirely lacks ghost structures and textures and other evidence of replacement. Its micritic nature is highly distinctive and contrasts sharply with coarser dolomite that clearly is of replacement origin. The above features, plus the nature of the occurrence of dolomite in eccentric oöids, suggest that such dolomite may be the result of either direct inorganic precipitation of dolomite or penecontemporaneous replacement of calcium carbonate crystals by dolomite. In either case the dolomite would be considered "first cycle" or "primary" in the petrographic sense, and suggests analogies with modern or Recent carbonate localities where micritic dolomite is forming or recently formed. Such dolomite is termed syngenetic (in the sense of Friedman and Sanders, 1967) and no distinction is drawn between inorganically precipitated and penecontemporaneously replaced varieties because currently there are no criteria to make the distinction in ancient dolomites like those in the Allentown.

The syngenetic dolomites in the Allentown are analogous to Recent syngenetic dolomite that is forming in such places as the Bahamas (Shinn and others, 1956), the Persian Gulf (Illing and others, 1965), and in several localities in south Florida (Shinn and Ginsburg, 1964). In each of these areas micritic dolomite with features similar to those described above in the Allentown is associated with evaporitic processes in hypersaline marine waters; and in each the nature of the occurrence is well documented. One would think that if such dolomite were preserved in ancient carbonates it would closely resemble the aphanocrystalline dolomite in the Allentown. It seems reasonable, therefore, to postulate such an origin for this dolomite.

The hypersaline brines that were instrumental in the formation of the syngenetic dolomites, in all probability, were largely responsible for the replacement dolomites also. Replacement is pervasive and complete, and it occurred on a regional basis.

There is no evidence that dolomitization is related to post-depositional structural elements or to stratigraphic discontinuities within the unit. Dolomite in cross-cutting relationships to stratigraphic units within the Allentown has not been observed either. Such widespread and thorough replacement is most likely if dolomitization occurred early in the history of the sediment, before consolidation was complete or even well along, when conditions of high porosity and permeability still prevailed. Under such conditions dolomitizing solutions had maximum access to the body of sediment, and the solutions could be constantly renewed and ionic concentrations easily maintained. It seems most likely, therefore, that dolomitization occurred within the depositional basin itself, perhaps in part penecontemporaneously (but diagenetically) and it probably accompanied the hypersaline conditions that favored deposition of syngenetic dolomite. Waters that were capable of depositing dolomite syngenetically should have had little difficulty in replacing pre-existing calcium carbonate sediment.

Relic textures (ghosts) and structures in the Allentown dolomites indicate that the original calcium carbonate and dolomite sediments were deposited in at least two, and probably three distinct marine environments. These are mostly sea marginal and include subtidal, intratidal, and supratidal zones.

Most of the original Allentown is intertidal. In terms of the lithologic cycle this environment, which comprises the area between mean high and mean low tide, is represented by the dolarenite, oölitic dolomite, dolorudite, algal stromatolite, and flat pebble (desiccation) dolorudite. Ripple marks and small-scale cross stratification, indicative of persistent currents of weak to moderate intensity, abound in these rocks. To the extent that modern Bahamian oöids are typical (Newell and others, 1960), these features appear to require relatively great agitation caused by waves breaking in warm shallow water saturated or supersaturated with respect to calcium carbonate. The dolorudites also imply that fairly strong currents were present in the environment, not only to break up the sediment but to transport it. Doubtless many rudites are due to subaerial exposure, desiccation, and polygonal cracking of the carbonate sediment at low tide, as numerous mud cracks are present. All of the above current related features are typical of the intertidal to very shallow subtidal environment, and, except for mud cracks, are not present at all in the supratidal zone.

Many organo-sedimentary structures associated with algal stromatolites are also formed in the intertidal zone. In particular, the laterally-linked hemispheroid structure that constitutes the collenia and cryptozoan form-species is very common in the Allentown. Such a structure characteristically develops in continuous mats and algal-bound sediments from the marine intertidal mud-flat environment, mainly in protected locations (Logan and others, 1964).

The finely crystalline dololutite (unit 2 of the Allentown cycle) almost surely is subtidal. The intraclasts are small and rounded, indicating a certain amount of transportation, and limiting to some extent the intensity of the current that finally dropped them. Also, this unit lacks the numerous primary features that would indicate persistent currents or periodic desiccation, so probably even a very shallow subtidal environment must be ruled out. Further evidence supporting this interpretation is found in the chemical analyses and thin sections, which show that pyrite, indicative of reducing conditions, and perhaps deeper water, is common in this lithology.

The supratidal environment, the area above mean high tide, is the most fleeting in terms of lithologic representation in the

Allentown. Only the syngenetic aphanocrystalline flat-pebble clasts of the desiccation dolorudite (stage 1 of the Allentown cycle), and possibly the eccentric oöids, are assigned to this environment. The evidence for this interpretation, based partly on analogy to dolomite occurrences in modern carbonate localities and partly on lithologic and chemical relationships among the Allentown lithotypes, is fairly strong, however, and may be pivotal to the understanding of the entire sequence of Cambrian and Ordovician dolomitic rocks in the Lehigh Valley.

The chief bits of evidence suggesting a supratidal origin for this dolomite are the very small crystal size, birdseye structures, very thinly laminated bedding, and lack of current features or other evidence of persistent currents or agitation. The presence of mud cracks and the lack of fossils tends to reinforce this interpretation, but these points could bear on the intertidal zone as well. By analogy with Pleistocene and Recent dolomites in south Florida and the Bahamas (Shinn and Ginsburg, 1964; Shinn and others, 1965) a supratidal origin is inferred. Birdseye structures (fig. 18), in particular, have been shown by Shinn (1968) to be reliable indicators of supratidal deposition, even if no other structures are present. Where they occur in micritic rocks that also contain mud cracks and laminated bedding the supratidal interpretation is greatly reinforced.

Dolomite may originate in the supratidal environment by means of essentially evaporitic processes. Since that environment ranges from a few inches to a few feet above mean high tide, carbonate deposition must depend upon periodic inundation by marine waters in periods of unusually high tides, which may be either seasonal or the result of storms. Fine calcium carbonate crystals may form or may be dropped from suspension as the water recedes or evaporates. As evaporation increases water salinity, and if ionic concentrations of calcium and magnesium are favorable, dolomite may begin to precipitate directly, or it may replace calcium carbonate penecontemporaneously. Further evaporation and concomitant salinity increases could result in deposition of anhydrite or gypsum, or other evaporite minerals. Since this mechanism is being proposed for the Allentown supratidal dolomite, one might wonder where are these other evaporites. None occur in the Allentown, and none are known to interfinger with the Allentown. Comparison of the Persian Gulf and Bahamas environments, however, suggests that relative humidity may be a critical factor. Gypsum accompanies dolomite in the Persian Gulf (Illing and others, 1965) but does not in the Bahamas and south Florida (Shinn and others, 1965). Except for the profound aridity of the Persian Gulf, the occurrences are similar.

The original sediments of the Allentown Dolomite have been interpreted as sea-marginal subtidal, intertidal, and supratidal deposits. Five of the seven lithotypes of the ideal Allentown cycle are assigned to the intertidal environment, one is subtidal, and one is supratidal (tab. 11). The cycle, as interpreted herein, represents, from bottom to top, a shift from subtidal, through intertidal, to the supratidal environment. The flat-pebble dolorudite is the bridge from the supratidal back to the subtidal environment but,

Table 11. Summary of depositional environments of dolomite lithotypes in the Allentown Dolomite, arranged in order of cyclic succession

Rock Type	Environment of Deposition
Dolomicrite	Supratidal
Algal stromatolite	Intertidal
Dolorudite	Intertidal
Oölitic dolarenite	Intertidal
Dolarenite	Intertidal
Dololutite	Subtidal
Flat-pebble dolorudite	Subtidal or Intertidal

conceptually, the writer feels that it belongs at the base of the ideal cycle and may be either intertidal or subtidal, or both. Doubtless, the shifting and migrating environments that are manifest in the vertical expression of the cycle also occurred laterally. Aside from tidal fluctuations that may be at least in part eustatic, there is no evidence of other causes of the cycles, cosmic or terrestrial, and probably no further explanation is needed.

Under the interpretation of the origin of the Allentown proposed herein one might predict that these rocks should interfinger with marine shelf limestones seaward (probably eastward) from the Allentown tidal complex. No such relations are known, but the evidence may not be available because of intervening Triassic or Coastal Plain sediments. On the other hand, on a grander stratigraphic scale, perhaps the overlying Beekmantown Group (upper part) and the

Jacksonburg Limestone should be considered in this context. The vertical sedimentary succession ought to reflect in a general way the lateral succession of the sediments and environments perpendicular to the depositional strike, especially in the relatively stable, broad, shallow epicontinental transgressive sea envisioned as the environmental plexus of the Appalachian Lower Paleozoic carbonate rocks. Although these overlying units are demonstrably slightly younger than the Allentown in the Lehigh Valley, their lateral extensions and equivalents down the regional depositional slope may represent marine shelf facies of Late Cambrian time.

Ordovician System

<u>Beekmantown Group</u>

Introduction

Beekmantown is a name extensively used throughout the middle Appalachians to designate Lower Ordovician carbonate rocks. In the Lehigh and Delaware Valleys of eastern Pennsylvania and western New Jersey, the Beekmantown Group comprises two mappable units, the Rickenbach Dolomite, below, and the Epler Formation, above. These formations principally are dolomite, but the Epler has numerous interbedded limestones. Despite the apparently clear criteria for subdividing the Beekmantown in this area, the rocks have many features in common. Because of this and because outcrops generally are scarce, especially away from major drainage, a great deal of experience and familiarity with these rocks are required to subdivide and map them consistently.

Rickenbach Dolomite

The Rickenbach Dolomite is almost entirely heterogeneous dolomite rock, and chert. The lower part of the Rickenbach includes laminated to thickly bedded, somewhat thickly parted, medium light gray to medium dark gray, finely to coarsely crystalline dolomite, and nodular or bedded dark gray chert. Dolomites weather light gray, light olive gray, and yellowish gray. Three lithotypes are most diagnostic: 1) a medium dark gray microfractured, thickly to very thickly bedded, very thickly parted, coarsely crystalline, slightly calcareous, apparently structureless dolomite, the cleavage faces of which produce a lustrous sparkle on freshly broken surfaces; 2) pseudo-dolorudites, the weathered surfaces of which display a peculiar patchy ruditic texture that is rarely, and then just barely, visible on fresh surfaces; and 3) true dolomitic rudites with angular or rounded intraclasts up to several millimeters in diameter. Figure 27 illustrates two of these lithologies. Although laminated beds, terrigenous material, and small-scale cross stratification occur in this part of the unit, they are relatively uncommon.

The upper part of the Rickenbach (about 200 to 300 feet) generally contains laminated to thinly bedded, aphanocrystalline to medium crystalline dolomites, commonly with thin lenses and stringers of quartz sand and silt. Sand and silt size quartz also is rather widely scattered throughout these dolomites. Small scale

Figure 27. Photographs showing typical outcrops of Rickenbach Dolomite. A, Very thickly bedded and parted coarsely crystalline dolomite. B, Typical coarse dolorudite.

A

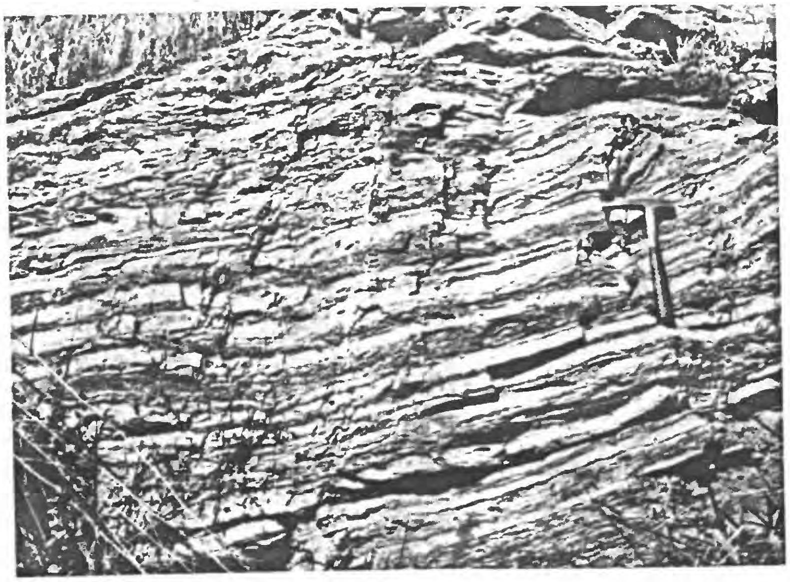
B

cross-stratification is fairly common, but other sedimentary structures have not been observed. Characteristically, this part of the Rickenbach is thinner-bedded and finer-grained than the lower part, but generally these units are not mappable as subdivisions.

Epler Formation

The Epler Formation is composed of interbedded limestone, argillaceous limestone, and dolomite (fig. 28), and bedded and nodular chert. Dolomites in the Epler are similar in many respects to those in the upper part of the Rickenbach. They are medium light gray to medium dark gray, and light olive gray, mostly thin-bedded to laminated, and aphanocrystalline to medium crystalline. They weather yellowish gray, light olive gray, and light brownish gray. Limestones are medium light gray to dark gray, thinly laminated to medium bedded, and aphanocrystalline to finely crystalline. Some beds, particularly in the upper half of the unit, consist principally of comminuted fossil fragments cemented with calcite. Many, probably most, limestones are in various incomplete stages of dolomitization and may be conspicuously mottled or streaked with light olive gray dolomite that contrasts sharply with medium gray limestone (fig. 29). Thin beds and laminae of quartz sand and silt, many with small-scale cross-laminae, occur throughout the limestones (fig. 28), and these beds commonly are the only clue to the location and structural position of the original compositional layering of the rock. Deformation resulting in extensive flowage and recrystallization has obliterated most other traces of bedding. Recrystallization and

Figure 28. Photographs of typical outcrops of the Epler Formation. A, Interbedded thin to medium bedded argillaceous limestone and dolomite. B, Silty banded limestone. Both outcrops near the village of Hecktown, Nazareth quadrangle.



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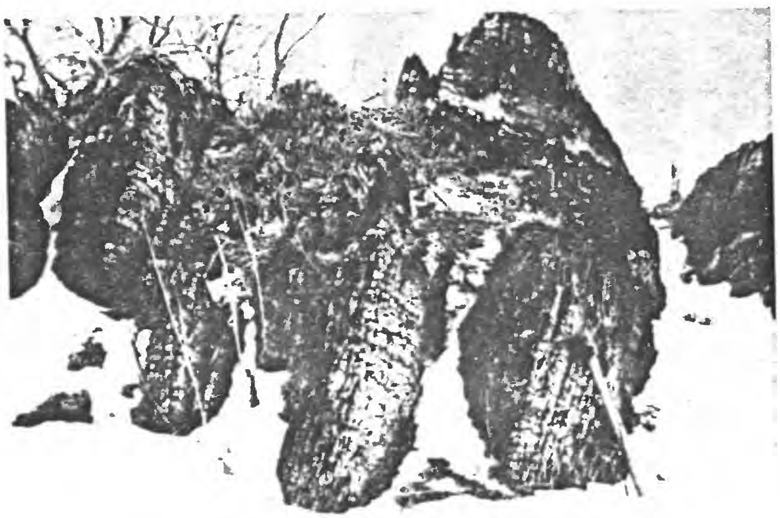
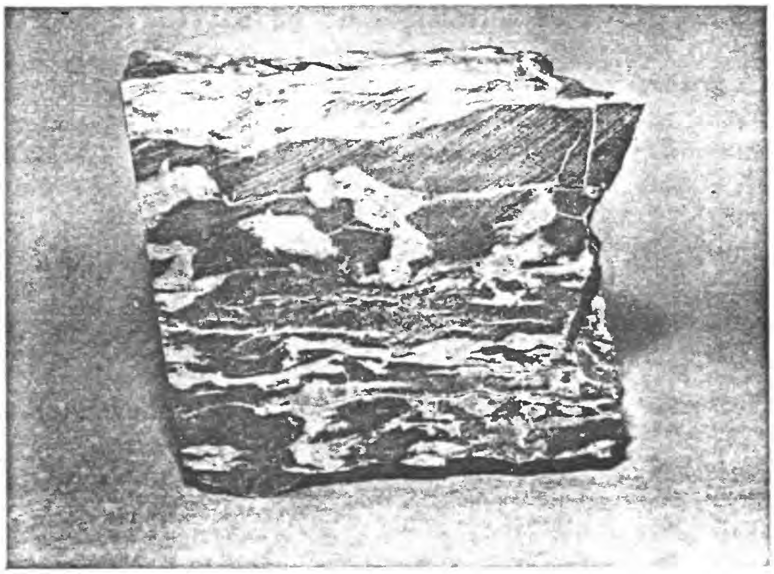


B

Figure 29. Photographs showing lithologic features of some Epler limestones. A, Mottling due to partial dolomitization. Lighter material is dolomite, darker is limestone. B, Laminations due to concentration of less soluble material on closely-spaced planes of flow cleavage, etched by differential weathering.



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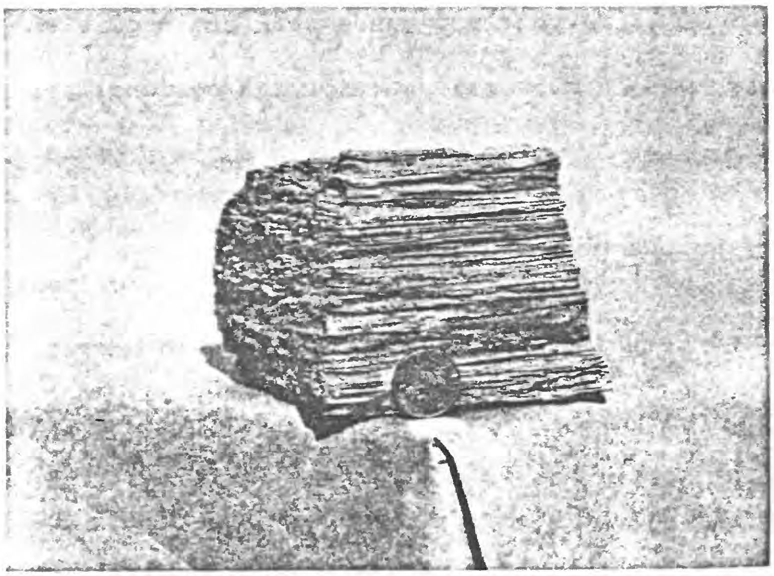


B

concentration of less soluble materials on cleavage surfaces sometimes has produced a secondary compositional layering (a false bedding) that might appear to be true bedding to the casual observer. This false bedding may or may not parallel true bedding, depending on the orientation of the flow cleavage. Some of the more conspicuous "thinly-laminated bedding" results from this process, especially where the rocks have undergone differential solution on weathered surfaces (fig. 29).

Bedded and nodular chert occurs throughout the Epler, interbedded with the limestones and dolomites. Near the base is a particularly chert-rich zone which has been useful in mapping the Rickenbach-Epler contact where Epler limestones are scarce.

Hard, pinkish gray to very pale orange orthoquartzite beds have been seen in the Epler by the writer at two localities in New Jersey (Belvidere and Bloomsbury quadrangles). To the writer's knowledge this lithotype has not been reported in the Epler in Pennsylvania.

Names and Distribution

Beekmantown is an old and deeply entrenched stratigraphic name for Lower Ordovician (Canadian) carbonate rocks in the middle Appalachians. It was first used by Clarke and Schuchert (1899) for rocks near the village of Beekmantown, New York.

Introduction of the name Beekmantown in eastern Pennsylvania apparently dates from the work of Miller (1911), who recognized faunal and lithic similarities both to the New York type section and to rocks of south-central Pennsylvania that Stose (1908, 1909) correlated with the type Beekmantown. Later reports by Miller (1925, 1934, 1939, 1941) firmly established the usage in the Lehigh and Delaware Valleys. Most subsequent workers have retained the name but have changed its status and meaning somewhat. Beekmantown replaced the terminology of Wherry (1909), who named these rocks Coplay Limestone after the town of Coplay, Lehigh County, Pennsylvania. Although, conventionally, only rocks of Canadian age are included in the Beekmantown, recent work by Willard (1958) and Hobson (1963) indicates that the upper part of the Beekmantown in eastern Pennsylvania includes Chazyan rocks. Willard (1958) argued that, in view of this age difference, Beekmantown should be abandoned and Wherry's name, Coplay, should be reinstated. This argument must be rejected, however, because paleontologic considerations, however valid, have no direct bearing on the definition of Beekmantown as a rock-stratigraphic unit, as it is herein used. Moreover, the name Beekmantown should be retained because of its extensive usage and its value as a unifying thread in the Appalachian stratigraphic maze.

Subdivision of the Beekmantown in eastern Pennsylvania derives from work in Berks County. Gray (1951, 1952a), in reports summarizing chemical data on Beekmantown and other rocks, suggested the possibility of subdividing the Beekmantown. Hobson's (1963) stratigraphic studies in central Berks County showed that the Beekmantown comprised four mappable subdivisions, which are, in ascending order, the Stonehenge Formation, Rickenbach Formation, Epler Formation, and the Ontelaunee Formation. A Beekmantown Group was thus established. As a result of reconnaissance studies Hobson (1963) suggested that rocks similar in nature and stratigraphic position to his Rickenbach and Epler Formations were present in the Lehigh and Delaware Valley areas.

A. A. Drake (1965) formally extended parts of Hobson's Beekmantown stratigraphy to the Delaware Valley. He recognized and mapped the Rickenbach Dolomite and the Epler Formation. The present study extends Drake's (1965) terminology westward into the Lehigh Valley, and Drake currently (1968) is applying it in quadrangles west and southwest of the Nazareth quadrangle. At present only the Rickenbach and Epler have been mapped. The Stonehenge Formation either pinches out or loses its identity along the regional strike at an undetermined point between the Schuylkill (Berks County) and Lehigh Rivers.

In the Nazareth quadrangle rocks of the Beekmantown Group occupy a broad, relatively flat belt between Route 22 and the town of Nazareth (pl. 1). There are good, though discontinuous, exposures along the loop of Monocacy Creek from the village of Steuben to the western quadrangle border at Pinetop, particularly around the village of Brodhead. Good exposures also occur along Bushkill Creek, near the village of Tatamy, and along Shoeneck Creek east and west of Coilton. Uppermost Epler is well exposed in several small quarrys, near the contact with the Jacksonburg Limestone, south and east of the town of Nazareth. The Trumbauer quarry, one mile east of Nazareth, is particularly noteworthy for both rocks (upper Epler) and structure. Elsewhere outcrops are unpredictable and scarce, owing to patches of glacial drift cover (pl. 1), and generally low relief.

Thickness

The Beekmantown reference section (Drake, 1965) near Carpentersville, New Jersey, provides the only reliable estimate of the thicknesses of the Rickenbach Dolomite and Epler Formation in the immediate vicinity. There the thickness of the Rickenbach is about 635 feet, and the Epler about 800 feet. These figures are consistent with estimates based on geologic mapping and construction of geologic cross-sections of the Nazareth quadrangle.

Miller (1939) estimated a possible range of thickness from 700 to 2000 feet for his Beekmantown Formation (undivided). His estimates and reconstructions are based on an entirely different structural picture, however.

Reconnaissance along the Lehigh River at Fullerton, Lehigh County, Pennsylvania, indicates that the total thickness of Beekmantown rocks there is about 1000 feet.

Stratigraphic Relations

The Rickenbach Dolomite conformably overlies the Allentown Dolomite. The contact is gradational and is characterized by a gradual appearance of coarsely crystalline, medium gray, thick-bedded dolomite, and shaly dolomite, and the disappearance of typical Allentown cycles, particularly beds of algal stromatolite, oölitic dolomite, and desiccation dolorudite.

The Rickenbach is transitional upward into the Epler Formation. In mapping and in measured sections the contact is drawn beneath the stratigraphically lowest limestone stratum.

The Epler is disconformably overlain by the Jacksonburg Limestone. This disconformity is recognized throughout eastern and southeastern Pennsylvania, and western New Jersey, regardless of whether the uppermost Beekmantown unit is Epler or Ontelaunee. East of the town of Nazareth the base of the Jacksonburg Limestone is marked by a few inches of dolomite-pebble conglomerate that thickens rapidly eastward into New Jersey. The contact is exposed at the top of an 80-foot face at the Trumbauer quarry which adjoins the workings of the Nazareth Cement Company east of the town of Nazareth. No angular discordance is apparent at the contact at this locality. The relations at the Trumbauer quarry are certainly equivocal however. The folding in the Epler is very complex and clearly disharmonic, and in no way does it match the structural style of the thick-bedded coarsely crystalline limestone of the superjacent Jacksonburg. The contrast in mechanical properties of the interbedded dolomites and limestones of the Epler and the thick coarsely crystalline limestone of the Jacksonburg probably accounts for this apparent discordance.

The lithology of the Epler adjacent to the disconformity is highly variable along strike, especially east of the town of Nazareth. This probably is related to the development of the dolomite pebble conglomerate at the base of the Jacksonburg, which, it will be noted, appears and thickens rapidly from about the same point. The variability thus is due at least in part to erosion.

In mapping, the Epler-Jacksonburg contact is drawn above the stratigraphically highest dolomite bed. Although some thickness of Epler limestone might occur above this level, in float it may be so similar to the lower unit of the Jacksonburg that the two are virtually indistinguishable. In measured sections the contact is drawn beneath the dolomite pebble conglomerate, if present, or beneath the first thickly bedded coarsely to very coarsely crystalline limestone of the Jacksonburg.

Age

To the writer's knowledge no fossils of any kind have been reported from the Rickenbach Dolomite in eastern Pennsylvania or western New Jersey. Hobson (1963) reported a few low-spired gastropods and a single orthocerid cephalopod in central Berks County, but extensive dolomitization apparently prevented further identification. He reported no fossils at all from the upper Rickenbach. It seems reasonable, however, to consider the Rickenbach as Early Ordovician (Canadian) in age because of its stratigraphic position between the Epler (see below) and, in Berks County and southwestward, the Early Ordovician Stonehenge Formation.

Willard (1958) reviewed the paleontoligic data from the upper Beekmantown (Epler of present usage) in eastern Pennsylvania. He indicates that, although much of the unit undoubtedly is Early Ordovician, there is evidence that the upper part contains rocks of probable early Middle Ordovician (Chazyan) age. Hobson (1963) described a similar relationship in upper Beekmantown rocks of central Berks County, as did Neuman (1951) and Sando (1957) in the Cumberland Valley of south-central Pennsylvania.

As noted previously, the Beekmantown conventionally includes only carbonate rocks of Early Ordovician (Canadian) age. From the above discussion, however, it is clear that the Beekmantown as used throughout the Great Valley in Pennsylvania includes rocks of a somewhat greater age range.

Correlation

Pertinent regional correlations of Beekmantown rocks in the Great Valley of Pennsylvania are summarized in Table 12.

Northeastward from eastern Pennsylvania the Beekmantown Group is correlative with the upper part of the Kittatinny Limestone of New Jersey (Lewis and Kümmel, 1940) and the Stockbridge Group of New York (Broughton and others, 1962). Both units are undivided carbonate sequences that comprise Lower Cambrian through Lower Ordovician rocks.

In the Lebanon Valley area, southwest of the Lehigh-Delaware Valleys, the Beekmantown Group includes all four formations represented in Hobson's (1963) subdivision (Field Conference of Pennsylvania Geologists, 1966; MacLachlan, 1967). The Rickenbach of Delaware-Lehigh Valleys probably occupies the stratigraphic interval of both Stonehenge Limestone and Rickenbach Dolomite of Lebanon Valley. There the Rickenbach thins and disappears southwestward as the subjacent Stonehenge correspondingly thicknes. The Epler Formations of the two areas are equivalent.

Table 12. Stratigraphy and correlation of Ordovician rocks in the Appalachian Valley, Pennsylvania, New Jersey, and New York.

Age	Cumberland Valley, Pa. (Field Conf. Pa. Geol., 1966) Autochthonous	Lebanon Valley, Pa. (Field Conf. Pa. Geol., 1966) Allochthonous	Nazareth quadrangle, Lehigh and Delaware Valleys. (This report; Drake, 1965; Drake and Epstein, 1967) Allochthonous	Kittatinny Valley, N.J. (Lewis and Kümmel, 1940) Allochthonous	Eastern New York (Broughton and others, 1962) Allochthonous (?)
Late Ordovician	Martinsburg Formation 1,000 feet	Martinsburg Formation 3,000 - 5,000 feet	Martinsburg Formation: Pen Argyl Member 3,000 - 6,000 feet	Martinsburg Shale 3,000 feet	Snake Hill Shale 3,000 feet
Late (?) Ordovician			Ramseyburg Member 2,800 feet		
Middle and Late Ordovician			Bushkill Member 4,000 feet		
Middle Ordovician	Chambersburg Formation 750 feet St. Paul Group 1,000 feet Beekmantown Group: Pinesburg Station Dolomite 450 feet	Hershey Formation 0 - 1,000 feet Meyerstown Formation 200 feet Annville Limestone 120 feet Beekmantown Group: Ontelaunee Formation 600 - 800 feet	Jacksonburg Limestone: Cement rock facies 300 - 1,000 feet Cement limestone facies 200 - 400 feet	Jacksonburg Limestone 135 - 150 feet	Balmville Limestone 100 feet
Early Ordovician	Rockdale Run Formation 2,500 feet Stonehenge Limestone 775 feet Staufferstown Formation 260 feet	Epler Formation 800 - 1,300 feet Rickenbach Dolomite Stonehenge Limestone 800 - 1,500 feet	Beekmantown Group: Epler Formation 800 feet Rickenbach Dolomite 635 feet	Kittatinny Limestone 2,500 - 3,000 feet (to Early Cambrian Tab. 4)	Stockbridge Group 4,000 feet (to Early Cambrian Tab. 4)

In the Cumberland Valley (Field Conference of Pennsylvania Geologists, 1966; MacLachlan, 1967) the Stonehenge Limestone and the Staufferstown Formation are correlatives of the Rickenbach, whereas the lower part of the Rockdale Run Formation is the Epler equivalent. Both the Lebanon Valley and Cumberland Valley Beekmantown sections are markedly thicker than that of the Lehigh and Delaware Valleys. It should be borne in mind that the Cumberland Valley sequence is autochthonous, whereas the Lebanon-Lehigh-Delaware Valley sequences are allochthonous.

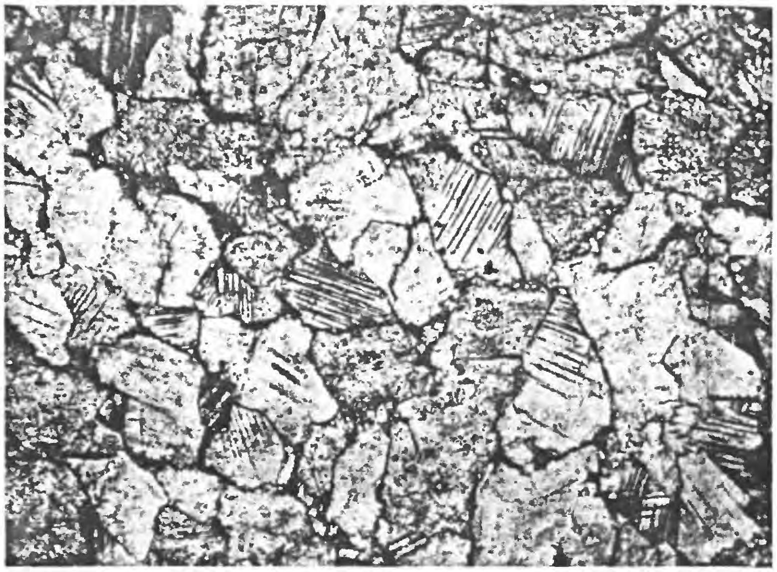
Petrography

Mineral Composition

The Beekmantown rocks are the most variable of the Lehigh Valley carbonates, in terms of mineral composition and texture. The principal lithologies, as previously mentioned, are limestone and dolomite. Most of the rocks are dolomite, but the Epler Formation contains numerous limestones that range in composition from mostly pure to extensively dolomitized. Thin sections, sawed and etched slabs, and staining techniques were used to study the Beekmantown lithologies.

Dolomite is the most abundant mineral. It occurs as sparry subhedral to anhedral crystals that range in size from 0.008 to 1.5 mm (fig. 30). Most is rather uniformly cloudy in transmitted light, but little in the way of palimpsest relics are apparent. The entire range of crystal sizes occurs in the Rickenbach dolomites, but the coarsely to very coarsely crystalline sizes are especially

Figure 30. Photomicrographs of petrographic features of Beekmantown Group rocks. A, Coarsely crystalline sparry dolomite mosaic, Rickenbach Dolomite. Plane light, 25X. B, Planar calcite fabric in Epler limestone. Plane light, 25X.



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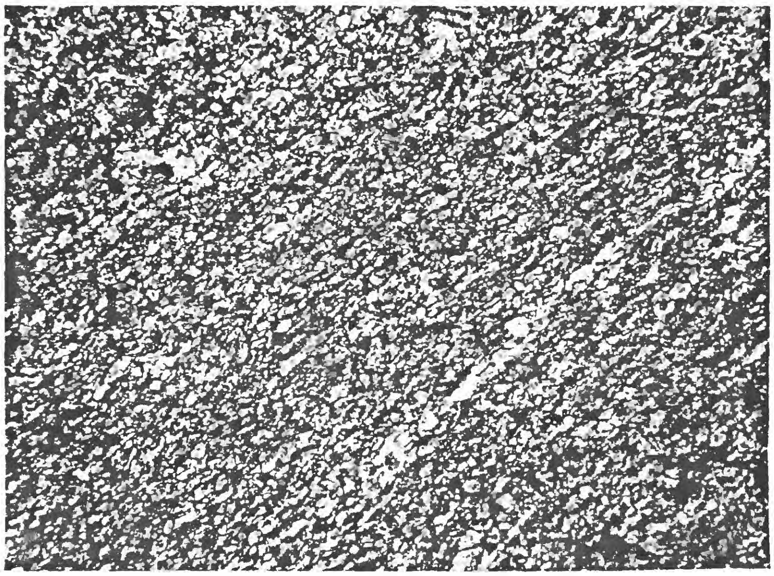

B

characteristic (fig. 30). Dolomite in the Epler is finer, typically very finely crystalline to medium crystalline. In both units, the finest grain sizes (0.004 to 0.025 mm) typically are associated with very finely laminated or essentially structureless dolomites.

Calcite is, of course, the principal mineral in the Epler limestones and dolomitic limestones. It occurs typically in elongate anhedral crystals that range in size (long diameter) from about 0.010 to 0.062 mm. Length to width ratios range from about 1.2 to 8. In all limestones examined these elongate grains clearly are oriented parallel to planes of rock cleavage (fig. 30). This phenomenon is apparent at every level of sampling, from outcrop to thin section. The less dolomite the rock contains, the more pronounced is this tendency. Calcite composes up to 95 percent of some Epler limestones, but the proportion varies greatly. All proportions to pure dolomite occur.

Quartz occurs in megacrystalline form as detrital grains, and in microcrystalline form as chert. Chert occurs in discrete beds and nodules, and as an interstitial filling. Detrital quartz grains typically are in the silt to fine sand size range and are mostly angular to subangular. Some angularity results from replacement by calcite, but some appears to result from physical sedimentation processes. The petrographic distinction is fairly clear. Ragged, embayed contacts are ascribed to replacement, whereas smooth or straight contacts are believed to result from physical processes. In many cases no such distinction can be made, however. Detrital quartz occurs as scattered grains, laminae, and thin beds mostly in the Epler Formation, and mostly in limy rocks (limestones and incompletely dolomitized limestones). It is relatively rare in the Rickenbach dolomites, and in the purer laminated or structureless dolomites of the Epler.

Sericite (muscovite) and chlorite occur in impure limestones of the Epler Formation. Both are in the size range of coarse silt to clay, and both are strongly oriented parallel to planes of flow cleavage (fig. 31). Cleavage typically is parallel or subparallel to bedding in essentially isoclinal folds, but transects bedding at high angles in fold hinges (fig. 31). The phyllosilicates are oriented similarly, thus it seems likely that their orientation is tectonically controlled. The minerals themselves could be crystallized and reoriented detrital micas or may have formed from detrital clay minerals present in the original sediment. Most sericite and chlorite is associated with quartz in limy rocks, and very little occurs in the purer dolomites. The abundance of these minerals is quite variable within outcrops, hand specimens and thin sections. Up to 30 percent occurs in some limestones near the top of the Epler.

Trace amounts of microcline, zircon, tourmaline, and iron oxides occur associated with quartzose layers. All are subangular to subround and seem clearly to be detrital. Authigenic pyrite also occurs in trace amounts throughout the Beekmantown.

Figure 31. Photomicrographs of petrographic features of rocks of the Beekmantown Group. A, Microfolds in silty argillaceous limestone of the Epler. Note grain elongation parallel to cleavage and sheared-off fold limbs. Plane light, 25X. B, Very finely crystalline to aphanocrystalline structureless dolomite, Rickenbach Dolomite. Plane light, 40X.

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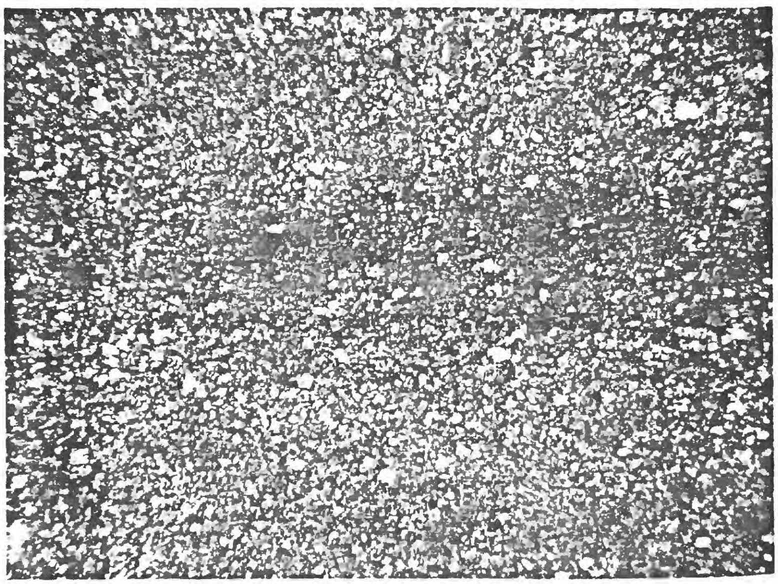

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Texture

The textures of the Beekmantown rocks depend entirely on mineral composition. Dolomites, regardless of grain size, are mosaics of anhedral to subhedral dolomite crystals (fig. 30). The fabrics typically are equigranular xenotopic to hypidiotopic, but some are inequigranular. Relics of oöids, stromatolites, and pellets are notably lacking in the Beekmantown carbonates, as are cross-stratification and graded bedding. Both of the latter occur in the quartzose layers, however, and are invaluable top and bottom indicators. Only fossil fragments and subangular to round carbonate intraclasts occur as ghosts in the dolomite mosaic. These range in size from about 0.05 mm to more than 64 mm. The larger intraclasts occur in sedimentary conglomerates that are especially characteristic of the Rickenbach Dolomite. In many thin sections the relic intraclasts are hardly recognizable in transmitted light, but are clearly visible in reflected light under low magnification. Opaque iron oxides are associated with some relic intraclasts, however, and clearly, delimit their extent. The very finely crystalline to aphanocrystalline dolomites (fig. 31), which typically are structureless or finely laminated, contain no vestiges of textural relics. All relics occur in medium to coarsely crystalline dolomites.

Limestones and calcareous dolomites in the Epler Formation typically exhibit a foliated character that results from the elongation of calcite crystals parallel to flow cleavage (fig. 31). Impure limestones typically contain sericite and fine elongate quartz that are similarly oriented, and thus also contribute to the foliation. Only in areas or zones of most intense deformation do dolomites exhibit a similar foliation.

Many Epler limestones show a degree of dolomitization that ranges from incipient to almost complete replacement. This can rather easily be seen in weathered exposures because the dolomite typically weathers to shades of orange and brown, whereas the limestone fresh or weathered, typically ranges in shades of gray. Figure 29 illustrates this contrast. In the fresher part of the same hand specimen the contrast between dolomite and limestone is barely apparent. The contacts between limestone and dolomitic mottles range from sharp to gradational. Gradational contacts are marked by dolomite crystals that are isolated in limestone just a fraction of a millimeter outside the purely dolomitic part of the mottle. To the naked eye, however, all mottle contacts appear sharp.

Chemical Composition

The Beekmantown is the most variable of the Lehigh Valley carbonates in terms of chemical composition. Chiefly this results from the presence of both relatively pure limestones and very pure dolomites in the unit, as well as varying degrees of replacement of limestone by dolomite. Representative chemical analyses of Beekmantown rocks from the Nazareth quadrangle are presented in Table 13.

It has been noted previously that some limestones in the Epler are very similar to those of the Jacksonburg Limestone, which are quarried extensively for use in the manufacture of portland

Table 13. Representative chemical analyses of limestone and dolomite from Beekmantown Group rocks in the Nazareth quadrangle

	1	2	3	4
$CaCO_3$	68.5	71.4	50.7	45.0
$MgCO_3$	20.3	15.3	36.8	40.6
SiO_2	7.6	9.4	8.1	12.0
Al_2O_3	2.6	--	--	--
Fe_2O_3	0.9	--	--	--
R_2O_3	--	3.7	4.4	2.2
Total	99.9	99.8	100.0	99.8

1. Epler Formation. Dolomitic limestone from the Trumbauer Quarry, 1/2 mile east of the town of Nazareth (O'Neill, 1964).

2. Epler Formation. Impure limestone from the Trumbauer Quarry, 1/2 mile east of the town of Nazareth (Miller, 1939).

3. Epler Formation. Dolomite from quarry 3/8 mile of village of Georgetown (Miller, 1939).

4. Rickenbach Dolomite. Structureless dolomite from Bushkill Creek, south of the village of Tatamy. Sample 284-3A. Analysis by Paul Elmore, L. Artis, G. Chloe, J. Kelsey, S. Botts, J. Glenn, and H. Smith, U. S. Geological Survey Analytical Laboratories Branch.

cement. The Epler limestones repeatedly have been considered for such purposes, but rarely have they been successfully so used. The chief problem is that the MgO content is too high, owing to partial dolomitization and to the presence of dolomite intraclasts. The very pure Epler limestones occur too sporadically and are typically too small to justify exploitation.

Origin of the Beekmantown Rocks

The Beekmantown Group in eastern Pennsylvania represents the transition from largely dolomitic rocks of the Leithsville-Allentown depositional epoch to the calcareous and argillaceous rocks of the Jacksonburg-Martinsburg epoch. As such, it has characteristics that are common to both the underlying and overlying sequence.

The dolomitic rocks of the Rickenbach Dolomite and the Epler Formation most probably owe their origin to processes and environments similar to those described for the Leithsville and Allentown. The dolomites probably comprise both diagenetic replacement and syngenetic types. However, the Beekmantown dolomites are notably lacking in relics of the variety of primary structures that typify the Leithsville and Allentown, thus evidence for assigning rocks to one process or the other is considerably less clear-cut. The principal criterion is grain size, in conjunction with other petrographic and field relations.

The dolomites of least questionable origin are those that occur as mottles in incompletely replaced limestones of the Epler Formation, such as is pictured in Figure 29. In these dolomites the grain size is considerably coarser than is that of the limestone host rock, and the dolomite contains textural relics of intraclasts and shelly fragments that are similar to such undolomitized features in the enclosing limestone. Clearly this dolomite is the result of diagenetic replacement of the calcium carbonate host. By inference, other Beekmantown dolomites with medium to coarse crystal sizes, relics of intraclasts, and variously shaped, mostly rounded and lenticular shadows that suggest relics of shelly fragments are also believed to result from such replacement.

Many medium to coarsely crystalline Beekmantown dolomites entirely lack relic intraclasts, ghosts that suggest shelly fossil fragments, or any other evidence of a pre-dolomitization texture. These rocks also are believed to result from diagenetic replacement. Although there are no specific petrographic criteria to support this ascertion, the coarser grain size is at least suggestive. The dolomitic mottles in incompletely dolomitized Epler limestones show, without exception, that dolomitization in these rocks is accompanied by crystal growth to sizes that range from medium fine to very coarse. The dolomitization of primary calcium carbonate structures in the underlying Leithsville Formation and Allentown Dolomite clearly followed a similar trend. The problem with this interpretation of the Beekmantown medium to coarsely crystalline structureless dolomites is that at present there is no tangible trace of a pre-replacement calcium carbonate fabric. All such evidence apparently was destroyed during dolomitization, or perhaps the original fabric was not sufficiently organized to survive replacement.

Many Beekmantown dolomities not only lack any trace of a previous sedimentary history but are exceedingly fine grained. Such rocks typically display finely laminated bedding, especially on weathered surfaces, or are visibly structureless. These dolomites are believed to be of syngenetic origin, either as direct inorganic chemical precipitates or as syngenetic replacements of calcium carbonate crystals and fragments prior to the development of a calcium carbonate depositional fabric. In either case the dolomite is essentially primary in the petrographic sense. Again, specific petrographic evidence to support this interpretation is lacking. The diminutive crystal size (very finely crystalline to aphanocrystalline) is only suggestive, but it is reinforced by the trend to larger crystal sizes in dolomites that clearly are of replacement origin, and by the fact that syngenetic dolomite in Pleistocene and Holocene occurrences (Illing and Wells, 1964; Shinn and Ginsburg, 1964; Shinn and others, 1965; Illing and others, 1965, among others) typically occurs as a micrite or as very fine grained crystals. The laminated character of the bedding of these fine grained dolomites indicate that depositional environment was relatively peaceful.

Limestones are not known to occur in the Rickenbach Dolomite, the lower member of the Beekmantown Group in the Lehigh Valley, but they are increasingly common toward the top of the Epler Formation. The stratigraphically lowest limestones that occur in the Lehigh-Delaware Valley lower Paleozoic carbonate sequence occur in the Epler. These facts, along with the sporadic incomplete dolomitization of some Epler limestones, reflect the waning of basin conditions favorable to the development of dolomite and herald the onset of the essentially calcium carbonate depositional epoch represented by the Jacksonburg Limestone. Indeed, many Epler limestones are virtually indistinguishable from those of the Jacksonburg. Petrographic studies indicate that most Epler limestones are clastic in one sense or another, as carbonate intraclasts, comminuted shelly debris, quartz grains, clays and micromicas commonly are major components.

Interpretation of the Beekmantown dolomites and limestones in terms of depositional environments is hindered by the paucity of primary features, such as desiccation cracks, oöids, algal stromatolites, and the like, that typically can be related to specific environments or environmental complexes. Perhaps such paucity is itself a clue to the depositional environment. However, one element that is central to any such interpretation is the fact that dolomite is basically an evaporitic mineral that is deposited from hypersaline solutions. This concept applies equally to syngenetic, diagenetic, and epigenetic dolomites representing vastly different geologic occurrences. Numerous recent studies of Pleistocene and Holocene dolomites as well as dolomite occurrences throughout the geologic record support this concept. Thus any interpretation of the Beekmantown depositional environment must provide for marine waters of sufficiently high salinity to favor the formation of dolomite.

Consideration of the possible Beekmantown depositional environments may take two directions. One direction assumes that the typical structures of calcium carbonate sediments that are so common and useful in the Allentown also were present in the Beekmantown sediments but were destroyed during the course of dolomitization. If this were so, little else can be said because the keys to understanding the ancient environment have been lost. This approach probably is invalid, however, because the other dolomitic rocks of the section show a remarkable tendency to preserve even the fine details of pre-dolomitization calcium carbonate structures, especially at the level of hand specimens and outcrops. The Beekmantown dolomites are similar in all essential lithologic and petrographic aspects to the Leithsville and Allentown dolomites, except that they are higher in the stratigraphic section and they contain few primary structures that are typical of calcium carbonate sediments. Thus it seems reasonable to assume that these structures simply were never an important part of the Beekmantown of this area, and that this fact in itself reflects in some way the dynamics of the depositional environment.

The near-shore tidal complex that seems to have been the locus of Allentown sedimentation, and probably the Leithsville also, must be rejected as a plausible depositional environment for the Beekmantown rocks because they lack the many current, high energy, and organo-sedimentary structures that typify this environment. In modern dolomite occurrences intense evaporation of sea water in the in-shore supratidal zone is the chief process by which both syngenetic and diagenetic dolomite are formed. This zone does not appear to be the principal agent of Beekmantown dolomite sedimentation, however.

The lithologic and petrographic character of the Beekmantown suggests a relatively quiet environment that was little influenced by tides and currents. The areal distribution of Beekmantown rocks similar to those described herein suggest that the environment was fairly widespread, on the order of about 50 miles along the regional strike and perhaps 5 to 10 miles perpendicular to the strike. A possible depositional environment consistent with the characteristics of the Beekmantown and the need to develop hypersaline brines would be shoreward of a barrier reef complex or other topographic obstruction. The barrier would hinder circulation of marine water of normal salinity, and would provide a buttress to intercept or attenuate marine and tidal currents. On the landward side of the barrier, under favorable conditions, hypersaline waters could develop by evaporation of waters in which circulation is restricted. Dolomite could then be precipitated syngenetically, and could form as a diagenetic replacement of earlier formed calcium carbonate in and underlying the basin of evaporation. Moreover, the denser Mg-rich solutions concentrated by evaporation could percolate downward through loosely consolidated calcium carbonate sediments, flow along bedding planes and through intergranular openings to replace sediments well outside the immediate area of the basin of evaporation.

Studies in localities in which modern dolomite is precipitating syngenetically or forming by replacement (Teodorovich, 1946; Alderman and Skinner, 1957; Jones, 1961; Deffeyes and others, 1964; Shinn, 1964; Illing and others, 1965; Shinn and others, 1965; Friedman, 1966) indicate that conditions of hypersalinity sufficient to initiate dolomite formation are accompanied by a rise in pH to between 8 and 10, a rise in the Mg/Ca ratio to between 10 and 40 to 1, or more, and an increase in salinity to more than 10 percent. By contrast, in the open marine system, normal pH is about 7, the Mg/Ca ratio is 5 to 1, and the salinity is 3.5 percent. As salinity increases beyond the point required for dolomite formation, more soluble members of the evaporite sequence, such as anhydrite, gypsum, or later, halite, begin to precipitate. In Coorong Lagoon, South Australia (Alderman and Skinner, 1957), halite begins to precipitate at 27.4 percent salinity and plants begin to die.

There are no sulfates or halides associated with the Beekmantown, but as was noted in the discussion of the origin of the Allentown, this may have been a function of the relative humidity. Interestingly, however, B. L. Miller (1937) reported finding casts of halite crystals in the Beekmantown in Lehigh County, and suggested that shallow water, salt-marsh-like conditions prevailed locally in Ordovician time. For many years this occurrence was treated as a mere curiosity, but it now takes on added importance as the evaporitic nature of dolomite becomes increasingly established.

Fossiliferous and argillaceous limy sediments would accumulate on the seaward side of the reef or topographic barrier, and would represent deposits of the shallow open sea of normal salinity. This presumably is the origin of the limestones that become increasingly common toward the top of the Epler. Thus, under this interpretation, the Beekmantown laminated aphanocrystalline dolomites are near shore syngenetic dolomites that formed in quiet water in a shallow basin behind a reef or topographic barrier. The coarser dolomites are of replacement origin, but the replacing solutions probably were generated by evaporation in the above-mentioned evaporite basin. The limestones are offshore shallow marine deposits that almost surely pass laterally into the near shore dolomites. Such relations of limestones and dolomites to ancient shorelines are not uncommon in the geologic record and several workers have reported on similar situations (Van Tuyl, 1918; Cloud and Barnes, 1948; Fairbridge, 1957; Krynine, 1957).

Presumably the shoreline lay northwest of the depositional area, and gradually it was being pushed further northwest by transgressing seas. As a result, offshore limestones, the lateral equivalents of the near shore syngenetic dolomites, untimately interfingered with and then succeeded the dolomites in the local vertical section as the evaporitic complex migrated northwestward out of the sea. However, in the Lehigh and Delaware Valleys there is little in the Beekmantown rocks, as they are currently known, that would provide a firm clue to the location of the ancient shoreline. Only regional stratigraphic and tectonic considerations suggest that it was northwestward.

The Beekmantown is a viable Appalachian stratigraphic interval that extends at least from New York State to southwestern Virginia. The interpretation presented herein applies only to the Beekmantown of the Lehigh and Delaware Valleys. Obviously in such an extensive carbonate depositional complex many different factors and subenvironments must have influenced the sediments, and rock textures and structures may, therefore, vary enormously over relatively short distances. Detailed studies in modern carbonate depositional systems repeatedly have shown this to be true. For example the Beekmantown of the Reading area (Hobson, 1963), Pennsylvania, approximately 45 miles southwest of the Nazareth quadrangle, is lithologically more varied than the rocks described herein. Included in the Group are rocks bearing algal stromatolites, flat-pebble conglomerates, mud cracks, crossbedding, and pellets, all of which are indicative of a tidal-flat complex, but all of which are rare to absent in the Beekmantown of the Lehigh-Delaware Valleys. Also, according to Hobsons (1963) descriptions, the Reading area lithologies are conspicuously more cyclic than the Lehigh Valley rocks, and this is another indication of a probable tidal flat origin for at least some of the Beekmantown Group rocks in that area.

Jacksonburg Limestone

Introduction

The Jacksonburg Limestone is closely associated with the principal mineral industry of the Lehigh Valley--the manufacture of portland cement. The upper part of the Jacksonburg, when properly treated, is a natural hydraulic cement. The lower part is a pure, high-calcium limestone that can be mixed with other ingredients to make cement.

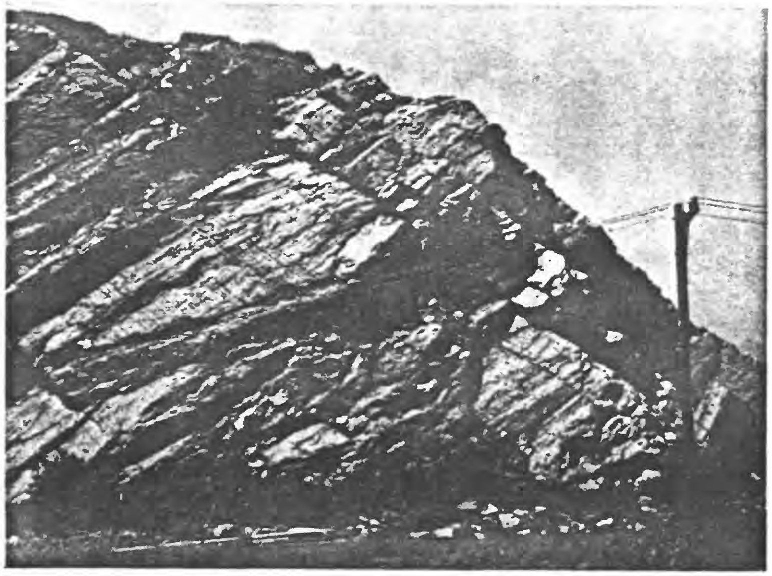
The cement industry in the Lehigh Valley is one of the two earliest of such industries in the United States, and dates from 1825 when the Lehigh Coal and Navigation Company dug a canal along the Lehigh and Delaware Rivers, from Mauch Chunk, Pennsylvania, to Philadelphia (Miller and others, 1939). Portland cement, much superior in properties to natural cement, first was produced in this country in the Lehigh Valley in 1875, and still is the backbone of the industry.

The Jacksonburg has been the subject of intense interest to geologists for many years because of its almost ideal composition for cement production. Some of the earliest geologists recognized a bipartite lithologic division of the Jacksonburg in Pennsylvania, and these parts have come to be known as the "cement limestone facies" (below) and the "cement rock facies" (above) because of their economic properties. These parts are viable stratigraphic units as well.

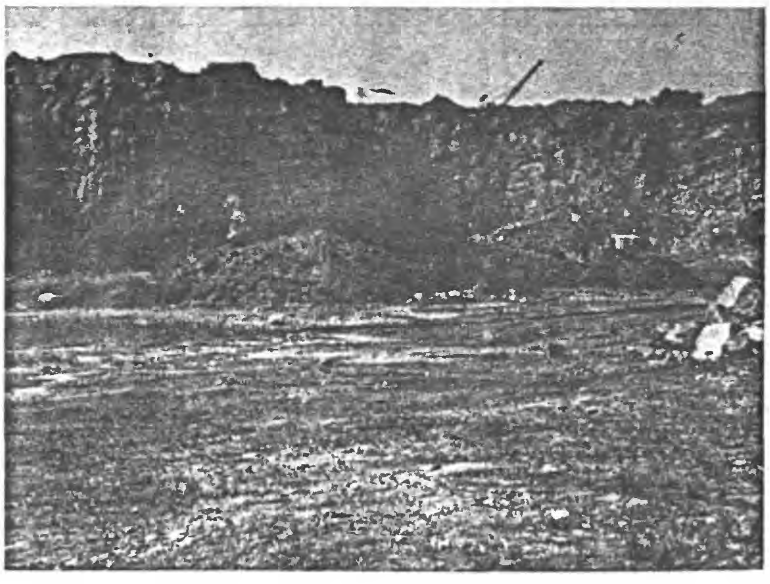
Cement Limestone Facies

The cement limestone facies of the Jacksonburg comprises medium to dark gray, medium to coarsely crystalline intrasparite and biosparite. In direct light freshly broken surfaces sparkle conspicuously owing to reflections from large sparry calcite grains. The rock weathers medium light gray to light gray, yellowish gray, and light brown. Stratification (fig. 32) characteristically is thick to very thick (up to 6 feet). However, many of such apparently thick beds actually include thin to very thin beds of dark gray

Figure 32. Quarry exposures of the Jacksonburg Limestone. A, Thickly bedded limestones of the cement limestone facies, Nazareth Cement Co., Nazareth, Pa. B, Cement rock facies, Nazareth Cement Co., Nazareth, Pa.



A



B

argillaceous material that are from a few inches to about 18 inches apart. These are virtually impossible to see except on weathered surfaces.

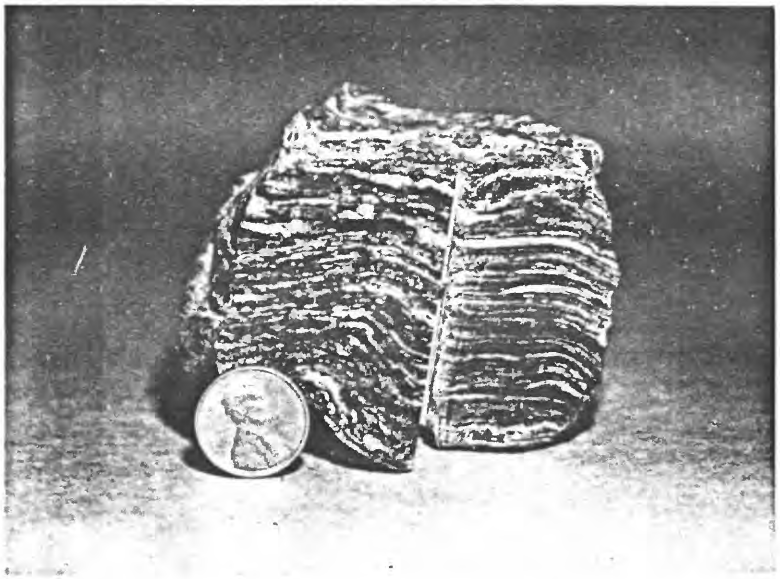
East of the town of Nazareth the base of the cement limestone facies is marked by a dolomite pebble conglomerate that is only a few inches thick, at most, within the Nazareth quadrangle, but which thickens markedly eastward into New Jersey. The dolomite pebbles are recognizable mostly as Beekmantown-type lithotypes.

Bentonites occur in this unit, and in the superjacent cement rock facies. Typically they are quite thin (2 inches or less) but a bed 1 foot thick occurs in the quarry of the Nazareth Cement Company, on the east side of the town. These bentonites appear to have no particular value as stratigraphic markers because they are so thin and difficult to locate, and, up to now at least, it has been impossible to show consistent lithostratigraphic correlation between occurrences.

Cement Rock Facies

The Jacksonburg cement rock facies is a dark gray to black, finely crystalline to aphanocrystalline argillaceous limestone that weathers medium light gray to light gray and pale yellowish brown. Brown stains are evident where pyrite is abundant. In contrast to the sparkling surfaces of the cement limestone, the cement rock has a soft velvety sheen on freshly broken surfaces. Veins of calcite or quartz are conspicuous locally (fig. 33). Owing to extreme deformation, the original compositional stratification is almost

Figure 33. Photographs of some lithologic features of the Jacksonburg Limestone. A, Prominent calcite veining related to intense deformation in the cement rock facies. B, Fence posts hewn from slabs of the cement rock facies. Smooth faces bounding posts are cleavage planes on which traces of bedding are clearly visible.



A

B

completely obliterated in this unit. Its response to deformation was quite different from that of the cement limestone facies or of the dolomites lower in the Paleozoic section.

The strong planar layering that characterizes the cement rock facies invariably is slaty cleavage. Despite the superficial resemblance of highly cleaved cement rock to the overlying Martinsburg Formation, even highly weathered cleavage fragments of cement rock retain enough calcium carbonate to effervesce briskly in hydrochloric acid.

Two medium to coarsely crystalline limestones, principally intrasparite and biosparite very similar to those in the cement limestone facies, occur within the cement rock facies. Within the Nazareth quadrangle the upper limestone is mappable only near the town of Nazareth, whereas the lower one is fairly persistent along the entire Jacksonburg outcrop belt. The lower limestone is about 50 feet thick; the upper one is about 35 feet thick. Both units attain somewhat greater thicknesses eastward, near the Delaware River. Sherwood (1964) reported that at the Lehigh Portland Cement Company quarry at Mud Run the lower crystalline limestone is 80 and the upper 105 feet thick.

Name and Distribution

The limestone that now bears the name Jacksonburg was well known to early workers in the area because of its economic importance in the cement industry. The stratigraphic nomenclature evolved somewhat independently in New Jersey and Pennsylvania, however. Rogers

(1858) and Prime (1878, 1883) recognized middle Ordovician fossils and correlated the limestone with the Trenton of New York State, as did Cook (1863, 1868) in New Jersey. Trenton was widely used in both states until Kümmel (in Spencer and others, 1908), building on Stuart Weller's (1903) detailed description of the section at the village of Jacksonburg, Warren County, New Jersey, used the name Jacksonburg to include all the New Jersey limestones that previously has been called Trenton in both New Jersey and Pennsylvania.

In Pennsylvania, however, Wherry (1909) and Miller (1911) recognized lithologic differences between the lower and upper parts of the Trenton cement limestones and used the names Nisky and Nazareth for these parts, respectively. Peck (1911), on the other hand, used the names Nazareth and Lehigh for approximately the same divisions, respectively. Later, Miller (1925) abandoned all of these names and used simply "cement limestone" and "cement rock" for the lower and upper parts, respectively, and adopted the name Jacksonburg for the entire unit.

R. L. Miller (1937) studied in detail the stratigraphy and paleontology of the cement limestones (Trenton-Jacksonburg) in eastern Pennsylvania in an effort to better determine their relation both to the Jacksonburg of New Jersey and to rocks of similar age and character, but of different names, west of the Lehigh Valley. Although he recognized the bipartite lithologic division of previous workers, he opposed elevating these divisions to member or formational status for the following reasons: 1) paleontologic evidence does not permit establishment of the time equivalence of the respective divisions, 2) the lithologic change is not contemporaneous throughout the area, and 3) intercalation of lithotypes representative of the upper and lower parts precludes mapping the contact with an acceptable level of precision. Miller therefore introduced the terms "cement limestone facies" and "cement rock facies" of the Jacksonburg Limestone.

Prouty (1959), working eastward into the Lehigh and Delaware Valleys from Dauphin, Lebanon, and Berks counties, used the names Meyerstown and Hershey for the lower and upper parts of the Jacksonburg, respectively. For what he considered to be cogent, though largely negative, paleontologic evidence, he did not extend these terms to the Jacksonburg of New Jersey, although some of the rocks there are lithologically similar and structurally related to the Pennsylvania equivalents. The Geologic Map of Pennsylvania (Gray and others, 1960) shows Prouty's stratigraphic nomenclature in the Lehigh and Delaware Valley areas.

In the present study of the Nazareth quadrangle Prouty's nomenclature has been rejected for several reasons. Application of Meyerstown-Hershey to the Jacksonburg of eastern Pennsylvania and not to laterally continuous, lithologically identical, structurally-related rocks in western New Jersey, on negative paleontologic evidence, obscures the obvious geologic affinities of the rocks in the two areas. In this case the name Jacksonburg clearly has precedence. Moreover, Prouty (1959) by no means established the rock-stratigraphic equivalence of his Meyerstown-Hershey interval and the Jacksonburg of Lehigh and Delaware Valleys. Although his paleontology clearly shows that the two sections are at least approximate temporal equivalents, time is irrelevant to the definition of these rock-stratigraphic units, and to statements of their lithostratigraphic equivalence. The Jacksonburg belt of Lehigh and Delaware Valleys and the Meyerstown-Hershey belt of Dauphin, Lebanon, and western Berks counties are not laterally continuous, and Prouty's (1959) published descriptions of his type sections, especially of the Meyerstown, raise enough questions concerning mutual lithologic similarities to warrant rejection of his nomenclature in the Lehigh and Delaware Valleys, at least for the present. It is perhaps of more than passing interest that although the Hershey Formation of Prouty (1959) and the cement-rock facies of the Jacksonburg are essentially similar lithologically, the Hershey, in the main belt that includes the type section, is not usable in the production of portland cement. There are subtle chemical and lithologic differences which account for this, and these are touched upon in the section on chemical composition.

It is with considerable reluctance that the nomenclature of R. L. Miller (1937) is adopted in this report. The principal reasons for doing so are that the terms "cement-limestone facies" and "cement-rock facies" are stratigraphically meaningful, they accurately reflect the economic significance of the unit, and they are widely used and understood by geologists, quarrymen, and laymen alike throughout the Jacksonburg cement-producing belt. These divisions, however, are rock-stratigraphic units in any modern sense of the term, and fully qualify for formational or member status. All of Miller's (1937) reasons for opposing such status are invalid in terms of the modern stratigraphic code. Two are based on notions of time equivalence and are irrelevant to definitions or ranking of rock-stratigraphic units, however accurate his points may be. His third objection has been met by the proven mappability of these units at scales of 1:24,000 or less by several different geologists over a period of years.

Undoubtedly any attempt to change the status of the Jacksonburg or its subdivisions would be met with colossal indifference, and would amount to little more than a scholarly exercise. Any attempt at redefinition would have to consider reinstating Wherry's (1909) and Miller's (1911) Nazareth and Nisky Formations because these directly corresponded to the present subdivisions and were adequately defined. The writer feels, however, that in spite of the rule of priority, the present nomenclature ought to be retained in some form because it is well-established, descriptive, and clearly meaningful in both stratigraphic and economic terms. This is consistent with Article 12, part b, of the modern Code of Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature, 1961).

The Jacksonburg mainly occurs in a narrow, continuous outcrop belt that extends over 24 miles from east-central Lehigh County, Pennsylvania, across central Northampton County, Pennsylvania, and Warren County, New Jersey. Isolated outcrop belts occur in the Paulins Kill Valley of New Jersey and its Pennsylvania extension, in some intermontane valleys of southern and southeastern Warren County, New Jersey, and in western Lehigh and eastern Berks Counties, Pennsylvania. These occurrences belong to different tectonic units, however, and it is not clear at this time how they relate to each other.

The problem of where to place the boundary between Jacksonburg rocks and Prouty's (1959) Meyerstown-Hershey Formations further west probably is best resolved by the rocks themselves. The Jacksonburg subdivisions are based on sound economic properties as well as lithologic character. The rocks are no longer "cement-limestone facies" and "cement-rock facies" in central Berks County, Pennsylvania, just east of the Schuylkill River, where the portland cement industry on the Jacksonburg belt ceases to exist, or never existed. Prouty's terminology should apply only west of the Schuylkill River.

The Jacksonburg crops out in a narrow belt across the northern part of the Nazareth quadrangle and in a small area centering about 0.8 miles northwest of the village of Brodhead (pl. 1). Outcrops are, of course, superb in the numerous active and abandoned quarries of the various cement companies. Such exposures are by no means easy to work with, however, because weathering processes have had little or no chance to accentuate the subtleties of composition, texture, or bedding character, and the complexity of the available structural detail usually is overwhelming. Owing to the extreme solubility of these rocks, natural exposures are almost nonexistent.

Thickness

The cement rock facies is about 700 to 800 feet thick throughout the mapped area. The cement limestone facies is about 350 feet thick. These figures are believed to be reasonably valid because they are based on the relatively good exposures in quarries. Away from the quarried areas accurate estimates of thickness are hindered by structural complexities and poor exposures. The thickness of 700 to 800 feet given for the cement rock facies probably is somewhat less precise than the figure given for the cement limestone facies. The upper part of the cement rock facies is less suitable for cement production, hence quarries do not normally expose the contact with the overlying Martinsburg.

Stratigraphic Relations

The Jacksonburg (cement limestone facies) disconformably overlies the Epler Formation of the Beekmantown Group. The youngest Beekmantown rocks are Chazyan (early Middle Ordovician), whereas the oldest Jacksonburg is believed to be of Rockland (late Middle Ordovician) age. Thus the disconformity probably represents part of Chazyan and all of Black River time.

The Cement Limestone Facies is overlain, apparently conformably by the Cement Rock Facies. The contact is difficult to map where exposures are poor. Prouty (1959) on the basis of rather sketchy paleontologic evidence and a dolomite pebble and boulder conglomerate that occurs at the base of the Hershey Limestone near Harrisburg, believed that his Hershey-Myerstown contact is a disconformity of regional significance. R. L. Miller (1937) found similar evidence of a disconformity in beds of upper Jacksonburg at some localities in New Jersey. This hiatus, according to Miller (1937), is suggested by an abrupt faunal change above the lower 58 feet (Leperditia zone) of the unit at the type locality, and by a few small conglomeratic pebbles at Portland. He does not associate this hiatus with lithologic changes characteristic of the Cement Limestone-Cement Rock contact, and fossils typical of the lower 58 foot zone itself are absent from most other Jacksonburg localities in New Jersey. Neither lithologic nor faunal evidence suggests such a disconformity within or between units of the Jacksonburg in the Nazareth area.

The Jacksonburg cement rock facies is overlain conformably by the Martinsburg Formation. The contact is transitional and is marked by a gradual upward increase in the proportion of insoluble material and a corresponding decrease in the proportion of calcium carbonate. R. L. Miller (1937) reported an angular divergence of 15 degrees just north of the Nazareth quadrangle boundary. The writer finds no such divergence, however, Disconformable relations were reported by Miller (1937) near the village of Stephensburg, New Jersey, which is considerably east of the area of this report, but the writer has no personal familiarity with the exposure.

A particularly enlightening exposure of the contact occurs 0.8 miles northwest of the village of Stockertown, just off the northern border of Nazareth quadrangle, on a steep bank above Bushkill Creek. There the transitional nature of the contact in this area is clearly shown. Sherwood (1964) reported results of insoluble residue determinations on samples taken across this contact. These data are presented in Table 14, along with similar data from a core obtained by the Universal Atlas Cement Company of Northampton, west of the area of this investigation.

The Jacksonburg-Martinsburg contact is fairly easily drawn in the field. The difference in the proportions of calcium carbonate and insoluble material typically is discernible in the manner in which the rocks react in hydrochloric acid. The cement rock reaction is an even, moderate to robust effervescence, whereas the Martinsburg reaction is weak and uneven. Also, weathered cleavage surfaces of the cement rock facies rarely show traces of bedding, but cleavage surfaces in the Martinsburg typically do. An adjunct criterion that is a good first approximation but which is by no means infallible is the manner in which thin (1/2 inch or less) cleavage plates of these two units break when struck by the pick-end of a rock hammer. Jacksonburg cement rock facies typically punctures, but does not shatter, and exhibits a well-defined square to rectangular hole that reflects the shape and depth of penetration of the pick. Cleavage plates of the Martinsburg, on the other hand, shatter completely, leaving many smaller chips but no hole. This property prompted the use of larger slabs of Jacksonburg cement rock as fence posts (fig. 33) because neat holes to support the cross-timbers could be hewn easily. These posts are still common on many of the older farms in the area, particularly those near the Jacksonburg

Table 14. Insoluble residue variation across the Jacksonburg-Martinsburg contact at two localities in eastern Pennsylvania (from Sherwood, 1964)

Bushkill Creek		Dragon Quarry, Northampton	
Outcrop Sample 5-foot intervals	Percent Insoluble	Hole Depth in Feet	Percent Insoluble
10	87.4	870-890	28.7
9	86.0	890-920	28.2
8	85.1	920-932	28.8
7	90.6	932-940	32.4
6	86.4	940-950	32.7
5	86.6	950-970	39.3
4	80.1	970-990	41.0
		990-1000	34.7
Martinsburg		1000-1032	59.7
....			
Jacksonburg		Jacksonburg	
		....	
3	57.0	Martinsburg	
2	48.2		
1	42.9	1032-1062	72.6
		1062-1083	75.9
Section upright		1083-1103	75.4
		1103-1139	78.7
		1139-1150	77.0
		1150-1180	88.1
		1180-1210	85.2
		1210-1230	86.6
		1230-1250	88.1
		Section overturned	

outcrop belt, and many can be found well up into the Precambrian terrane of the Reading Prong.

Age

Parts of the Jacksonburg in easternmost Pennsylvania and western New Jersey are very fossiliferous, but fossils are scarce west of Nazareth, Pennsylvania. Most paleontologists are in essential agreement on the interpretation of the fossil data, but there are slight differences in details. In general, the Jacksonburg is considered to be late Middle Ordovician in age.

Fossil occurrences and assemblages in the Jacksonburg were described by Weller (1903) and Miller (1937). The principal point of disagreement concerns the age of <u>Leperditia fabulites</u> (Conrad) in the lowermost 58 feet of Jacksonburg at the type locality. Weller (1903) thought these basal beds were of Black River age (tab. 15), whereas Miller (1937) assigned these lowermost beds a Trenton age, equivalent to the Rockland Formation of the Middle Ordovician, Mohawkian, standard section of New York. This reference section is presented in Table 15. Although in this case the distinction is a fine one, Miller (1937) seems to have the better case. He points out that although this fossil is found in the Black River, it also is abundant in lower Trenton rocks in the Mohawk Valley and Ontario. Moreover, it had recently (1937) been shown that another typically Black River fossil cited by Weller (1903) was of Rockland (Trenton) age. The remainder of the Jacksonburg at the type section is equivalent to the Kirkfield (Hull of the now obsolete terminology

Table 15. Mohawkian (upper Middle Ordovician) standard sections referred to by authors cited in this report

<table>
<tr><th colspan="2">Stage</th><th>R. L. Miller (1936)</th><th>Twenhofel and others (1954)</th><th>Kay (1964)</th></tr>
<tr><td rowspan="6">Trentonian</td><td rowspan="10">Mohawkian</td><td>Gloucester</td><td rowspan="2">Cobourg</td><td>Collingwood</td></tr>
<tr><td>Collingwood</td><td>Cobourg</td></tr>
<tr><td>Cobourg</td><td rowspan="2">Sherman Falls</td><td>Denmark</td></tr>
<tr><td>Sherman Fall</td><td>Shoreham</td></tr>
<tr><td>Hull</td><td>Kirkfield</td><td>Kirkfield</td></tr>
<tr><td>Rockland</td><td>Rockland</td><td>Rockland</td></tr>
<tr><td rowspan="4">Black River</td><td>Chaumont</td><td>Chaumont</td><td>Chaumont</td></tr>
<tr><td>Lowville</td><td>Lowville</td><td>Lowville</td></tr>
<tr><td>Pamelia</td><td>Pamelia</td><td>Pamelia</td></tr>
<tr><td>(not given)</td><td>(not given)</td><td>(absent)</td></tr>
<tr><td colspan="5">Chazyan (lower middle Ordovician)</td></tr>
</table>

of Miller) and Sherman Fall of the standard section, as is the entire unit at all other localities in New Jersey and Pennsylvania.

Barnett (1965) studied conodonts from the Jacksonburg and found them to be Trentonian in age, probably older than upper Shoreham of the standard section (tab. 15), possibly correlative with the Rockland, Kirkfield, and lower Shoreham (lower Sherman Fall). Conodonts of this age are dominant throughout the Jacksonburg and belong to the North American mid-continent province.

Curiously, however, Barnett (1965) found conodonts in one isolated area, near Clinton, New Jersey, that are characteristic of the northern European province. Although mid-continent forms occur also, the fauna is definitely northern European in aspect and appears to be slightly younger (late Middle Ordovician to early Late Ordovician) than conodonts in the rest of the Jacksonburg. In the same area, the stratigraphic and structural affinities of the Martinsburg Formation, overlying the Jacksonburg, for years has been an enigma to geologists familiar with the area. Recent work on graptolites from the Martinsburg in this particular New Jersey area shows that it is allochthonous, as some graptolites are equivalent in age to the lower Beekmantown Group (A. A. Drake, written communication, 1970).

Correlation

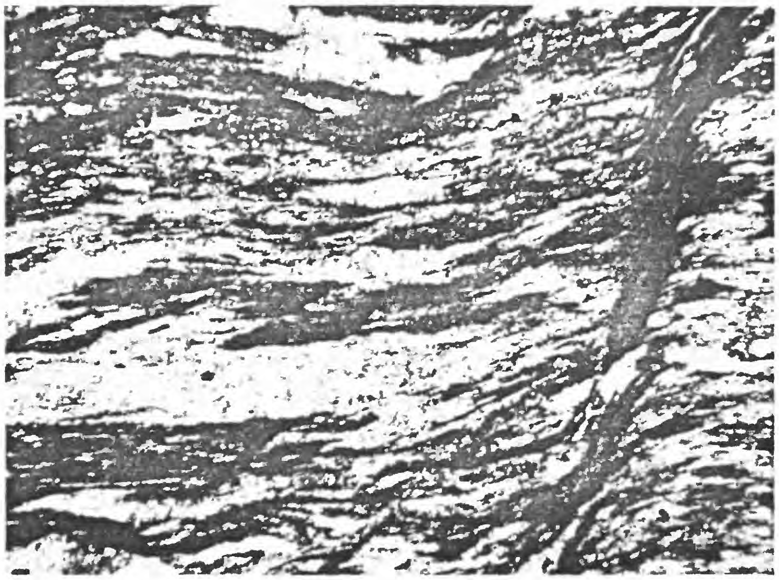
In New York State the Balmville Limestone is the equivalent of the Jacksonburg (Broughton and others, 1962). Southwestward from the Lehigh Valley, in the Lebanon Valley section of the Great Valley, the Meyerstown and Hershey Formations (Prouty, 1959) are temporal equivalents of the Jacksonburg cement limestone facies and cement rock facies, respectively. In the Cumberland Valley of south-central Pennsylvania, the upper part of the Chambersburg Limestone (Field Conference of Pennsylvania Geologists, 1966) is the Jacksonburg equivalent. These regional relations are summarized in Table 12.

Petrography

Mineral Composition

Calcite is the principal mineral constituent of both the cement limestone and cement rock facies of the Jacksonburg. The occurrence is quite different in each, however. In the cement limestone facies calcite occurs typically as anhedral crystals that are mostly equidimensional and prominently twinned. Twinning is especially conspicuous in larger grains. The grains range from about 0.008 to 1.5 mm in diameter, but average about 0.6 mm. Calcite in the upper crystalline limestones is similar but tends to be slightly coarser. Calcite in the cement rock facies occurs in single anhedral grains or polycrystalline aggregates that are strongly elongate and oriented parallel to each other and to flow cleavage as shown in Figure 34. Grain widths range from 0.01 to 0.15 mm, but lengths vary from 0.03 to 0.3 mm. The average length:width ratio is about 2.5 to 1. Calcite generally constitutes between 80 and 90 percent of the cement limestone facies, and between 50 and 75 percent of the cement rock facies.

Figure 34. Photomicrographs of cement rock facies, Jacksonburg Limestone. A, Extreme internal deformation of argillaceous limestone. Note shredded microfolds and strong orientation of all elements. Flow cleavage (S_2) runs left to right and is transected by fracture (crenulation) cleavage (S_3) in right side of photo. Plane light, 35X. B, Argillaceous limestone. Note orientation of textural elements. Plane light, 35X.



A

B

Quartz composes from 3 to 10 percent of the cement limestone facies, and as much as 20 percent of the cement rock facies. Its occurrence, like that of calcite, strongly differs in the two units. In the cement limestone quartz is in small, widely scattered equidimensional grains with normal to slightly undulose extinction. The size ranges from 0.005 to 0.05 mm in diameter, but averages about 0.03 mm. Grain boundaries and contacts range from very straight to very ragged. Ragged boundaries appear to result from replacement by calcite. Some quartz is detrital, but most is syngenetic or diagenetic. This is suggested by its occurrence as very irregular slivers and blobs with the microcrystalline texture of chert that fill irregular voids between calcite crystals. The fabric is strongly suggestive of growth in place. Detrital grains of microcrystalline quartz (chert) are present but these are essentially equidimensional and somewhat rounded. In the cement rock facies quartz occurs typically as grains and polycrystalline aggregates that are strongly elongate and have undulose extinction. Grain or aggregate widths range from 0.01 to 0.05, whereas lengths vary from about 0.01 mm to 0.2 mm. The average length:width ratio is 5:1. Whatever its origin in the rock, quartz in the cement rock facies very much reflects the tectonic processes that produced the strong foliation. The same is true for other minerals as well.

Micas and clay minerals are unimportant in the cement limestone facies but constitute an important fraction (up to 10 percent) of the cement rock facies. Ray and Gault (1961) investigated these minerals in drill cores of the Jacksonburg by X-ray diffraction and differential thermal analysis techniques. They found several poly-morphic forms of micas and clays, including 1M and 2M muscovite or illite, chlorite, and montmorillonite. None, except perhaps the montmorillonite in bentonites, now appear to be purely detrital, although originally most of it undoubtedly was originally detrital. All are strongly oriented in the plane of foliation (flow cleavage), or in the plane of slip cleavage, and, in fact, are partly responsible for the foliation. All except montmorillonite seem to have been recrystallized during diagenesis or metamorphism because they maintain a consistently parallel relation to the cleavage even where the cleavage intersects bedding at high angles in fold hinges. Montmorillonite probably is derived from volcanic material and very likely was airborne before reaching the site of deposition. Montmorillonite minerals commonly are a conspicuous constituent of bentonites--volcanic ash deposits. As described previously, bentonites occur in both the cement limestone facies and cement rock facies of the Jacksonburg. Although montmorillonite is not commonly found in rocks this old, Weaver (1953) documented its occurrence in K-bentonites of approximately the same age in central Pennsylvania.

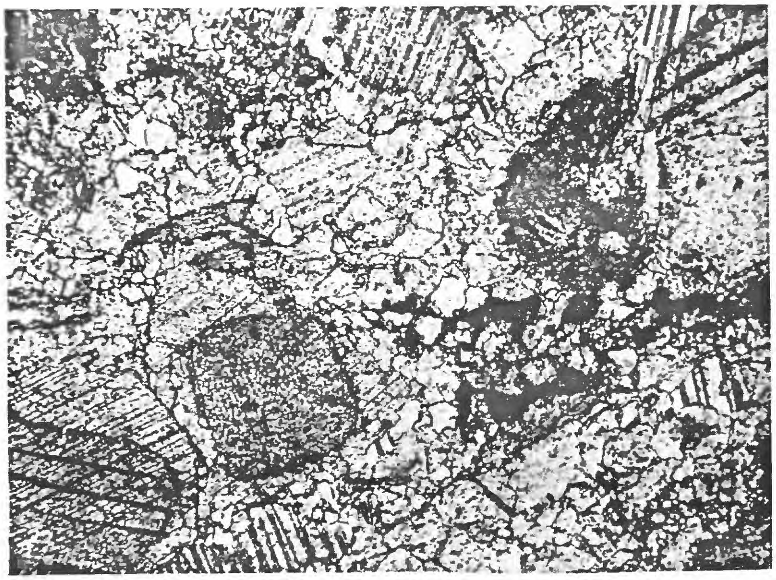
Dolomite is present in very minor amounts, generally about 2-4 percent, in both facies of the Jacksonburg. It occurs as sporadically subhedral to anhedral grains that vary in size from 0.06 to 0.10 mm in diameter. It occurs in intraclasts of dolomitic composition rather than as a replacement or syngenetic precipitate within the limestones.

Feldspar, apparently albite, occurs in rare scattered grains. Pyrite is present in widely scattered cubes and irregular blebs in both facies. Sizes are variable but generally are very fine, about 0.01 on the average.

Texture

As indicated by the foregoing mineral descriptions, the texture of the cement limestone facies and the upper crystalline limestones, is vastly different from that of the cement rock facies. Clearly visible in the coarse calcite mosaic of the cement limestone facies is a highly clastic subtexture. Allochems include comminuted fossil material, mostly brachiopods, corals, crinoids, and particularly bryozoans, and carbonate intraclasts that range up to 2 mm in diameter (fig. 35). The calcite texture is xenomorphic inequigranular, and is largely the result of recrystallization and grain growth. The clastic subtexture is a sedimentary feature rather than cataclastic, as Sherwood (1964) suggests. Cataclastic textures normally involve post-consolidation mechanical breakdown, crushing, or grinding in response to directed stress. That such directed stress affected the Jacksonburg is evidenced by the flow and slip cleavages that are prominent in the argillaceous cement rock facies, which, by virtue of its composition, is a very sensitive indicator of such stresses. However there is nothing in the subtexture of the cement limestone to suggest cataclasis, as such. The carbonate intraclasts and fossil fragments are subangular to well rounded and much of the debris is floating in calcite cement. The fragments

Figure 35. Photomicrographs of cement limestone facies, Jacksonburg Limestone. A, Intraclasts and bryozoan fragment in recrystallized limestone. Note that intraclasts are completely enclosed within optically continuous areas of twinned calcite; plane light, 35X. B, Bioclastic hash in cement limestone; plane light, 35X.



A

B

themselves are whole and complete as fragments and show no evidence of rebreakage after consolidation. Nor does one see fragments or trails of smaller fragments that could have been displaced from larger fragments during cataclasis. If cataclasis were strong enough to completely dislocate the various parts of the whole, one would expect much internal deformation accompanied by streaky banding. In fact, these rocks show no such features. The calcite crystals are extensively twinned but it is abundantly clear that translation on these planes is no more than a tiny fraction of a millimeter. Cataclastic textures are restricted to shear zones in these rocks.

Chemical Composition

Typical chemical analyses of the cement limestone and cement rock facies in the Nazareth quadrangle are presented in Table 16. The lithologic differences in the Jacksonburg facies are clearly reflected in the analyses. The proportion of $MgCO_3$ remains fairly constant, but that of $CaCO_3$ varies inversely with the quantity of insolubles. The markedly lower percentage of $CaCO_3$ in analysis Number 3 reflects the proximity of the Martinsburg contact.

In general the percentage of $CaCO_3$ in the cement limestone facies varies from about 80 to 95 percent. In the cement rock facies $CaCO_3$ varies from about 65 to 75 percent, except near the Martinsburg contact zone. Where $CaCO_3$ is deficient, the cement rock is upgraded or "sweetened" with cement limestone.

In the production of portland cement the raw material consists of about 75 percent $CaCO_3$, 20 percent SiO_2, Al_2O_3, and Fe_2O_3,

Table 16. Typical chemical analyses, in percent, of cement limestone facies and cement rock facies of the Jacksonburg limestone, Nazareth quadrangle

	Cement Limestone Facies		Cement Rock Facies	
	1	2	3	4
$CaCO_3$	83.55	90.0	54.62	74.00
$MgCO_3$	5.64	3.2	4.18	4.36
SiO_2	8.24	4.7	29.68	13.52
R_2O_3[a]	2.28	1.9	--	7.04
Al_2O_3	--	--	6.63	--
Fe_2O_3	--	--	2.23	--
	99.71	99.8	97.34	98.92

[a] $R_2O_3 = Al_2O_3+Fe_2O_3$

1. Drill hole sample, Lone Star Cement Corporation, Nazareth, Pa. (Miller and others, 1939, p. 362).

2. Penn Allen Cement Co., Nazareth, Pa. (O'Neill, 1964, p. 34).

3. Lone Star Cement Corporation, Nazareth, Pa. Uppermost cement rock, near Martinsburg contact (Miller and others, 1939).

4. Lone Star Cement Corporation, Nazareth, Pa. Lower beds of cement rock (Miller and others, 1939).

and 5 percent MgO and alkalies. The more desirable limestones used in cement manufacture closely approach this composition naturally, as does much of the cement rock facies (compare analysis 4, tab. 16). If the $CaCO_3$ content is too high, silica and argillaceous materials can be added to achieve the desired proportions, as is done with the cement limestone facies. The most critical property of limestones in the manufacture of portland cement is the content of MgO, which must not exceed 5 percent (8 percent $MgCO_3$). Some limestones in the Epler Formation meet these requirements, but interbedded dolomites and partially dolomitized limestones make economic recovery impossible.

Although bulk chemical composition is not normally a mappable distinction, these analyses provide a tangible basis for extending the Jacksonburg stratigraphic terminology only as far as the Schuylkill River, northeast of Reading, Pennsylvania, and for not extending Prouty's (1959) Meyerstown and Hershey Formations into the Lehigh Valley. It has been pointed out that cement production on the Meyerstown-Hershey-Jacksonburg belt ceases just east of the Schuylkill River. The reason is evident in chemical analyses published by Gray (1951) of Jacksonburg-equivalent rocks west of the Schuylkill, which show that these rocks (here called Meyerstown and Hershey Formations) contain either too much silica, too much magnesia, or too much of both to be suitable for manufacturing portland cement.

Origin of the Jacksonburg Limestone

The Jacksonburg Limestone probably originated in a shallow marine to sea-marginal marine environment. The clastic fossil debris that composes much of the cement limestone facies suggests that a complex of subenvironments may have been involved. The bulk of the fossil material (bryozoans, crinoids, brachiopods, corals) are bottom-dwellers that typically inhabit warm, clear, shallow water that is protected from strong waves or currents, such as would typify lagoons or shallow marine platforms. On the other hand, since most fossil material consists of well-sorted comminuted shells and carbonate fragments, considerable turbulence must have affected these sediments as well. The paucity of terrigenous material indicates that the mainland was either considerably removed from the depositional site or that terrigenous sediment was by-passing the site of deposition. In sum, the cement limestone facies probably represents first-cycle biogenic-calcium carbonate sediments that originally formed on a warm shallow shelf or in a marine lagoonal complex. These environments subsequently must have been altered, probably by a transgressing sea with stronger currents, wave action, and turbid waters. The organisms were overwhelmed, torn-apart, broken and abraded, and redeposited. Aside from the typical environpreferences of the organisms themselves, there is no other evidence of the original growth environment of the fossils.

Shallow banks and marine shelves, such as are suggested by the Jacksonburg cement limestone facies, can produce, under favorable conditions, considerable thicknesses of limestones of high purity.

Modern analogs of such environments are the Campeche Bank, off the north coast of the Yucatan Peninsula of Mexico, and the Bahama Bank southeast of Florida.

The Jacksonburg cement rock facies is a mixture of carbonate and terrigenous clastic sediment. There is little specific evidence with which to speculate about a depositional environment, except to say that it must have been a marine setting capable of carbonate production, and having some access to sources of terrigenous sediment. It is possible that the cement rock is an offshore marine carbonate that is mixed with terrigenous sediment that may have come from some comparatively remote source, or which may have by-passed nearer shore limestones. If the vertical succession of lithotypes, pure limestones composed of intrasparite and biosparite overlain by argillaceous limestones, also reflect the lateral distribution of these facies in a transgressing sea, then the above-mentioned possibility is quite plausible. The contact between the cement-rock facies and cement limestone facies, under this interpretation, must be time-transgressive. R. L. Miller (1937), on paleontologic evidence, offered a similar interpretation of the contact, though in somewhat different terms. He also noted that the argillaceous facies clearly overlaps the limestone facies in a northwest direction in New Jersey and felt that the same was true in Pennsylvania. The same relation probably holds for the Jacksonburg-Martinsburg contact as well. The Martinsburg, to be discussed subsequently, would represent, under this interpretation, a farther offshore, deeper water facies largely beyond the influence of near shore carbonates.

Martinsburg Formation

Introduction

Middle and Late Ordovician pelitic rocks occur in the Appalachian Valley from New York to Tennessee. Throughout most of the outcrop belt, nearly 500 miles long, these rocks are named Martinsburg, though stratigraphic rank varies from area to area.

The Martinsburg in eastern Pennsylvania for years has been the subject of controversy because of disagreements concerning its subdivision and the structural implications that necessarily follow from each proposal. Drake and Epstein (1967) have recently summarized the detailed field studies of several geologists of the United States Geological Survey in ten 7 1/2-minute quadrangles, including Nazareth, in eastern Pennsylvania and western New Jersey. They divide the Martinsburg into three mappable units, which are, in ascending order, the Bushkill, Ramseyburg, and Pen Argyl Members. Only the lower part of the Bushkill Member crops out in the Nazareth quadrangle.

Bushkill Member

The Bushkill Member is dark medium gray to dark gray, thin-bedded, banded claystone slate that weathers medium gray to very light gray or yellowish brown. Thin interbeds of quartzose and low-rank graywacke sandstone and siltstone, and carbonaceous slate typically occur in a cyclic fashion, giving the rock a "ribboned" appearance. Stratification units are thinly laminated to thinly bedded (1/10 in. to about 4 in.) but the great majority are much less than 1 inch thick.

Slaty cleavage is very prominent throughout the Bushkill Member and invariably is the most conspicuous planar element seen in outcrops. Bedding appears on cleavage planes as color-bands, but textural and compositional differences are discernible as well.

The Bushkill Member is equivalent to the "hard" slate member of Behre (1927). Commercial slate of somewhat inferior quality has been quarried from this member in the past, mostly for roofing purposes. Much superior slate presently is quarried from the two overlying members of the Martinsburg, however.

An estimate of the thickness of the Bushkill Member is not possible in the Nazareth quadrangle because only the lower few hundred feet are present. However, Drake and Epstein (1967) report a thickness of about 4000 feet for the member in nearby areas.

Name and Distribution

Geiger and Keith (1891) first introduced the name Martinsburg for exposures of shale near Martinsburg, West Virginia, but Keith (1894) first described the occurrence adequately. From there the name was spread throughout the Appalachian Valley because of the great similarity of lithology, age, and stratigraphic position of these pelitic rocks. Stose (1909) first used the name in

Pennsylvania for black shale and sandstone in the Mercersburg-Chambersburg district, south-central Pennsylvania.

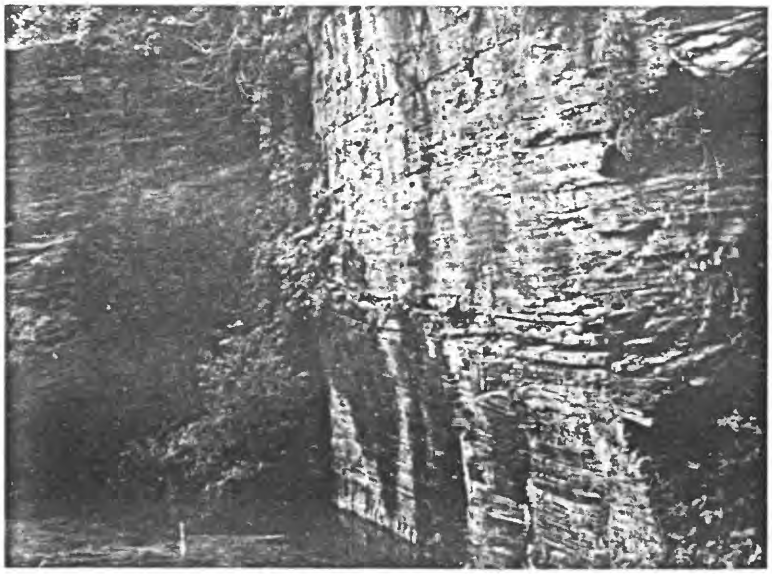
Rogers (1858) originally designated these pelitic rocks as Formation III of his Matinal Series. Lesley and others (1892), in a summary of the Second Geological Survey of Pennsylvania, noted the similarities between these rocks and the Hudson River Slate and Utica Black Slate of New York State and applied these names to the unit.

The present terminology is an outgrowth of the controversy over subdivision of the Martinsburg. The argument centers on whether the division should be two-fold or three-fold. Peck (1908) apparently was the first to recognize three subdivisions, whereas all previous workers and most subsequent workers recognized only two. Behre's (1927) work on the slate districts of Northampton County is the first modern exposition of a tripartite subdivision. He thought that the Martinsburg comprised a lower "hard" slate member, a middle sandy member, and an upper "soft" slate member. Stose (1930) extended his terminology from south-central Pennsylvania into the area, advocated a bipartite subdivision, and dismissed Behre's upper slate member as the lower slate member repeated by folding. Willard and Cleaves (1939) supported Stose and, later, Willard (1943) attached names to the two subdivisions and elevated the Martinsburg to group status. The present terminology (Drake and Epstein, 1967) is based on lithologic subdivisions that essentially correspond to those of Behre (1927). It is interesting to note that the tripartite subdivision almost invariably is favored by those who evolved and tested their ideas by detailed field mapping, e.g. Peck (1908), Behre (1927), and Drake and Epstein (1967). On the other hand, geologists whose work largely was of a reconnaissance nature, principally Stose (1930), Willard and Cleaves (1939), and Willard (1943), usually advanced a bipartite subdivision.

In the Nazareth quadrangle the Bushkill Member crops out only in the extreme northwestern part of the area (pl. 1). The terrain underlain by this member is significantly higher in elevation and relief and is topographically distinct from the dominantly carbonate terrane to the south. Outcrops are neither large nor plentiful, but they are generally adequate for mapping at a scale of 1:24,000. Most exposures occur in road cuts and numerous small quarries. Good exposures are available on almost all the north-south roads and deeper valleys through the outcrop belt. Typical exposures are illustrated in Figure 36.

On a regional scale, similar but subtly different lithologies can be traced from the type area across Pennsylvania into the Lehigh and Delaware Valleys. Work currently in progress (A. A. Drake, Jr., unpublished data) suggests that somewhat different Martinsburg sequences are in tectonic contact west of the Lehigh River, and that a bipartite Martinsburg as proposed by Stose (1930) and Willard (1943) is valid for areas to the west of the eastern Pennsylvania slate belt. The Lehigh-Delaware Valley Martinsburg can be traced into New York State near the Hudson River, where it is covered by rocks of the Taconic sequence.

Figure 36. Photographs showing typical outcrops of the Bushkill Member, Martinsburg Formation. A, Small abandoned slate quarry in member. Sub-horizontal planar structure is flow cleavage (S_2). B, Roadside outcrop. Hammer head rests on cleavage plane. Both exposures are along Route 946, northwestern part of Nazareth quadrangle.



A

B

Stratigraphic Relations

The Martinsburg Formation, Bushkill Member, conformably overlies the Jacksonburg Limestone. The contact is transitional. The Bushkill Member is conformably overlain by the Ramseyburg Member. The contact is a transition from banded claystone slate to a series of prominent graywacke sandstones and siltstones (Drake and Epstein, 1967). This contact is not present in the Nazareth quadrangle, however. It occurs nearly two miles north of the quadrangle boundary, in the Wind Gap quadrangle.

The Martinsburg Formation is unconformably overlain by the Shawangunk Conglomerate of Silurian age. This contact probably is also a zone of décollement and will be discussed in a subsequent chapter.

Age

Fossils in the Martinsburg are quite scarce in eastern Pennsylvania and western New Jersey. The transitional contact with the subjacent cement rock facies of the Jacksonburg Limestone implies that the base of the Bushkill Member represents late Middle Ordovician, Middle Trentonian (Sherman Fall), time. Farther west, Willard (1943) reported shelly fossils of early Late Ordovician (Edenian) age from shales equivalent to the Bushkill, and Late Ordovician (Maysvillian) fossils from sandy rocks equivalent to the overlying Ramseyburg Member. Thus, the Bushkill Member appears to range in age from late Middle Ordovician to early Late Ordovician.

The top of the Martinsburg is Late Ordovician (Maysvillian or younger) in age.

In the type area, in West Virginia, and in the Cumberland Valley of Pennsylvania the Martinsburg ranges from late Middle Ordovician (Sherman Fall) to Late Ordovician (Maysvillian) in age (Woodward, 1951; Bassler, 1919). This range is similar to that of the Martinsburg in eastern Pennsylvania. Until very recently the Trenton (Sherman Fall) age of the lower part of the Bushkill Member was inferred from the transitional contact with the subjacent Jacksonburg. However, Berry (1970) recently found that graptolites in the Pen Argyl Member, the uppermost unit of the Martinsburg in eastern Pennsylvania, are from a graptolite zone equivalent to the upper Middle Ordovician Trenton stage, which includes Sherman Fall. Aldrich (1967) reported graptolites in the lower Martinsburg near Seemsville, Pennsylvania which confirm a Middle Ordovician age for these beds. Thus, graptolites, in contrast to the shelly fauna, suggest that the bulk of the Martinsburg is of Trenton age.

It should be noted that all so-called Martinsburg rocks in eastern Pennsylvania do not fit the age range described above. West of Lehigh Valley in the region north of Harrisburg, Pennsylvania, Martinsburg-like rocks are interbedded with variegated red and green shales, conglomerates, bedded cherts, basaltic volcanic rocks, and carbonate rocks. These rocks contain graptolites of Early to Middle Ordovician (Deepkill) age, equivalent in age to the late Canadian to Chazyan stages, and are older than any known "normal" Martinsburg. Similar rocks with graptolites of similar age also occur in the

Clinton-Jutland area, New Jersey, where so-called Martinsburg is older than the subjacent Jacksonburg Limestone. The stratigraphic and structural significance of these peculiar rocks is not adequately known at the present time. Clearly the resolution of the problem is fundamental to the interpretation of the stratigraphic and tectonic history of the area.

Correlation

The Martinsburg Formation of this report is correlative with pelitic rocks mapped as Martinsburg throughout the Appalachian Valley from northwestern New Jersey to the type area. In New York State the Martinsburg equivalent is the Snake Hill Shale. Exotic pelitic rocks of the Hamburg area, Pennsylvania, and the Clinton-Jutland area, New Jersey, are excluded from this discussion because their affinity to "normal" Martinsburg is questionable.

Petrography

Mineral Composition

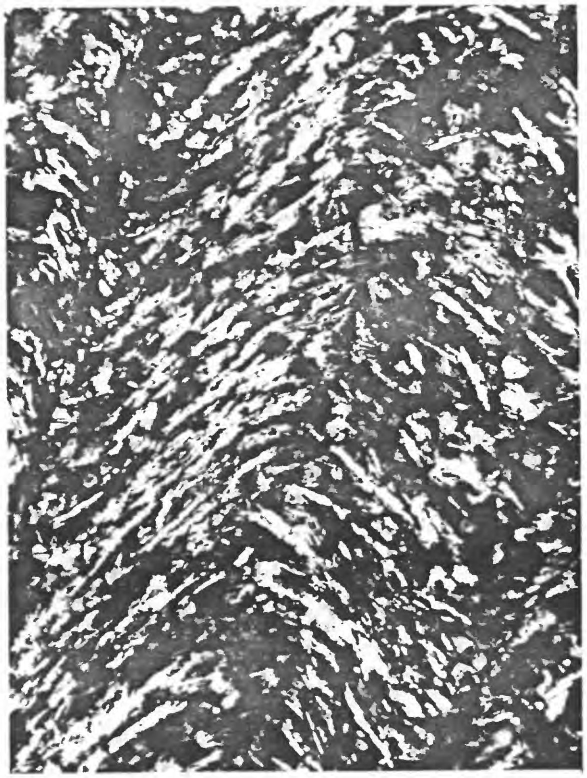
Quartz is one of the two most abundant minerals in the Bushkill Member, composing on the average about 40 percent of the slate. Two petrographically distinct varieties are present. One is essentially angular, equant to strongly elongate detrital quartz with slightly to strongly undulose extinction that is apparent even in the finer grains. The size ranges downward from 0.025 mm to less than 0.001 mm, thus almost all of this quartz is in the medium silt to clay size range. The average size range is 0.005-0.01 mm. Many grains contain minute needle-like inclusions of what appears to be rutile. The other variety is much less common and occurs in equant to elongate grains that are characterized by one or more very straight well defined boundaries and contacts. Extinction is normal to slightly undulose. This variety averages 0.05 mm in size, but the size range is much less than that of the detrital variety. This type of quartz appears to be the result of incipient metamorphic recrystallization.

The principal phyllosilicate in the Bushkill Member is sericite (muscovite), though not all workers agree on this point. Behre (1927), whose study of the Martinsburg was most exhaustive, thought that the mica is muscovite and this is supported by X-ray studies by McBride (1962), J. B. Epstein (written communication, 1969), and A. A. Drake, Jr. (written communication, 1970) all of whom found that the principal mica is the 2M polymorph of muscovite. Bates' (1947) chemical analyses and thermal tests indicated that a member of the illite group was the dominant phyllosilicate, but that sericite also occurs and is very similar to the illite mineral. Maxwell (1962) also believes that illite is the principal phyllosilicate.

Sericite occurs in irregular, elongate, subparallel, doubly tapered plates, 0.005 to 0.015 mm in length, that are interspersed among quartz grains (fig. 37). It constitutes from 35 to 40 percent of the slate.

Chlorite is a minor micaceous mineral constituent (about 5 percent by volume) that occurs with sericite as a fibrous

Figure 37. Photomicrographs of slate in Bushkill Member of Martinsburg Formation. A, Flow cleavage (S_2) is defined by strongly oriented sericite, and crenulated by slip cleavage (S_3), in left side of photo; crossed polarizers, 100X. B, Flow cleavage (S_2) is transected by crenulation ("slip") cleavage (S_3) which is also marked by recrystallized sericite. Medium gray mineral with sericite in both photos is quartz; crossed polarizers, 250X.



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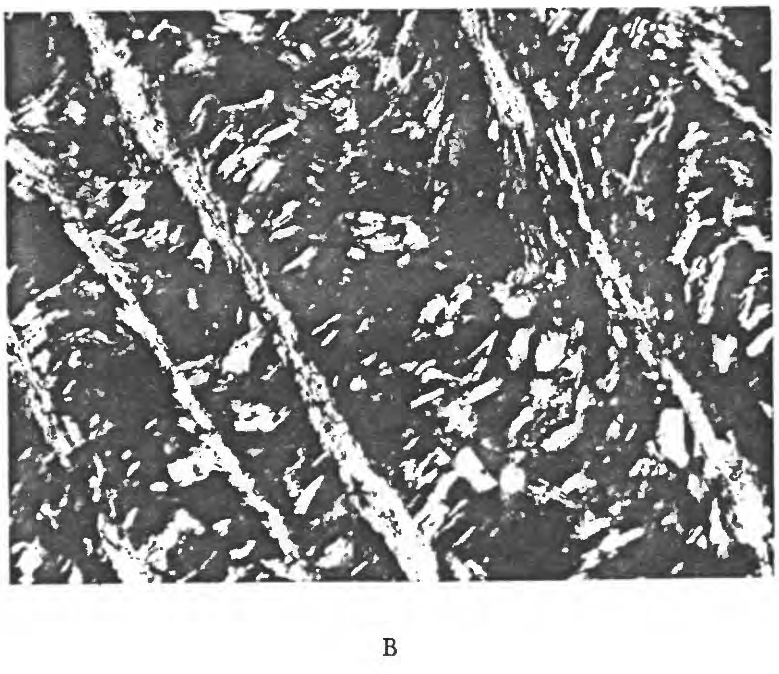


B

intergrowth, and as an alteration product of what apparently was detrital biotite.

Rutile occurs as subangular to subround detrital silt grains, as inclusions in detrital quartz grains, and as long (up to 0.01 mm) slender needles that parallel sericite plates. Magnetite, relatively fresh plagioclase, and very much rounded zircon also occur in the Bushkill Member but are of minor importance. Locally, calcite replaces quartz in some horizons and also occurs as an interstitial filling. Carbonaceous(?) material occurs as black rounded or lenticular masses ranging in size from 0.01 mm to less than 0.001 mm.

Texture

The orientation of sericite is the most significant textural element in slates of the Martinsburg Bushkill Member. Sericite may be oriented either in a single direction, which is the more common case, or in two directions, depending on the local style of deformation. Both directions represent planes of cleavage. In the more common case sericite is strongly oriented in a plane that is parallel, subparallel or at a low angle to the original compositional layering. This is the plane of the older and principal rock cleavage, a flow cleavage (S_2). In many cases this principal flow cleavage is deformed by a younger cleavage (S_3). Typically this younger cleavage is a crenulation cleavage that warps, folds, or actually displaces the older flow cleavage along a plane of dislocation. Petrographically this crenulation cleavage is manifested either by the planes of dislocation themselves or by the axial planes of the microfolds, which are parallel to the planes of dislocation. In some cases, however, depending on the local structural style, the crenulation cleavage passes into a second flow cleavage marked by very strong orientation (i.e., recrystallization) of sericite in planes parallel to planes of dislocation and crenulation axial planes, and transecting the older flow cleavage. Photomicrographs in Figure 37 illustrate these relations. The second cleavage is termed crenulation cleavage even where strong recrystallization of micas clearly has occurred on dislocation planes and the cleavage has the typical aspects of a flow cleavage. This is done for convenience in identifying and separating the respective cleavage directions in the field. In the great majority of cases, the effect of the second cleavage is to crenulate or displace an older, better developed flow cleavage, regardless of whether the second cleavage is itself, in fact, a flow cleavage marked by strong reorientation or merely a plane of dislocation with no apparent reorientation. This distinction is one that must be made under the microscope in most cases, and is of little value in the field. Chlorite, where it is intergrown with sericite, also is oriented in the plane of cleavage, and thin sections cut parallel to cleavage planes show that both chlorite and sericite are elongated perpendicular to fold axes. All visible sericite and related chlorite in the Bushkill Member is oriented in one or the other of the above cleavage directions.

Other minerals, notably quartz and rutile, have a definite orientation and elongation that is related to that of sericite and recrystallization, reorientation, and flowage along planes of cleavage especially in areas of severe deformation. Limestones in the Epler and Jacksonburg especially show these features.

Most workers (Behre, 1927, 1933; Bates, 1947; McBride, 1962; Epstein, personal communication, 1970) believe that slate in the Martinsburg is the result of metamorphic recrystallization, even though they do not agree as to whether the principal metamorphic minerals are sericite (muscovite) or illite, and chlorite. In contrast, Maxwell (1962), believes that the chief phyllosilicate is illite and that it developed and was oriented, along with the cleavage, during diagenesis. He documents his agrument with numerous references to possible low temperature-low pressure stability relations of illite and chlorite. Such possibilities would tend to give Maxwell's hypothesis an element of credibility if other evidence of metamorphism under directed stress were not present in these rocks. Curiously, however, Maxwell divorces the Martinsburg from the regional tectonic and stratigraphic setting, and asserts that the underlying carbonates contain no evidence of plastic deformation or recrystallization. This is patently untrue because the carbonates exhibit clear evidence of extensive flowage, recrystallization and reorientation; in fact, so much so that the rocks are locally foliated (figs. 34 and 37). Moreover, the fact that chlorite and illite, as well as the flow cleavage, could form diagenetically, does not imply that they did, in fact, do so. The bulk of the evidence cited above seems to favor low grade metamorphism under directed stress.

chlorite. Grains of quartz, intercalated between sheets of mica, are flattened in the plane of cleavage and elongated perpendicular to fold axes. Rutile needles also are elongated perpendicular to fold axes and are parallel to cleavage.

In thin sections it is clear that sericite flakes are not simply oriented detrital micas, but probably are the result of either diagenetic or metamorphic recrystallization of detrital clays or micromicas under directed stress. The strong association of sericite, chlorite, quartz, and rutile orientation and elongation with rock cleavage, and its relation to major structural features, strongly suggests that recrystallization was not merely diagenetic.

Cleavage characterized by strong reorientation and recrystallization of phyllosilicates commonly is parallel or subparallel to bedding in the Martinsburg, thus reorientation phenomena might conceivably result from compaction and diagenesis. However folding typically is isoclinal or nearly so, and cleavage at high angles to bedding can be seen in crests and troughs of folds in many places. The texture appears, therefore, to be very much the result of incipient to moderate low-grade metamorphic recrystallization.

The orientation of minerals in the Martinsburg and the composition of the phyllosilicates, as discussed above and in a foregoing section, are the principal source of information about the metamorphic history of Paleozoic rocks in the Lehigh Valley. The Hardyston Quartzite and the various carbonate rocks are not notably. revealing in respect to mineralogic changes related to metamorphism, but they do exhibit abundant evidence of plastic deformation,

Chemical Composition

Typical chemical analyses of the Bushkill Member of the Martinsburg Formation are presented in Table 17. The analyses are quite consistent and probably are indicative of very uniform depositional conditions. Moreover, the chemical composition of the Bushkill Member is very similar to that of the "average shale" (tab. 17). Although the rocks of the Bushkill Member have undergone slight to moderate low-grade metamorphic recrystallization and, strictly speaking, are not shales, the similarity to the composition of average shale suggests that metamorphism was essentially isochemical.

The abundance of most oxide components can be related directly to the detrital or metamorphic mineralogy. Silica is dominant, of course, because it is the dominant component of sericite and also occurs freely as quartz. Al_2O_3 is a principal constituent of sericite, as is K_2O. MgO is associated with chlorite and Na_2O with plagioclase.

Origin of the Martinsburg Formation

A few general remarks about the origin of the Martinsburg must suffice here because this report can contribute few data to the subject. Only the lower Bushkill Member crops out in the Nazareth quadrangle.

The Martinsburg clearly is of marine origin. The sediments were deposited in an open and, perhaps in part, a fairly deep sea. This is indicated by the graptolites. Radiolarian cherts occur

Table 17. Typical chemical analyses of slates from the Bushkill Member, Martinsburg Formation, and composition of "average shale"

	1	2	3	4
SiO_2	64.7	59.1	60.8	58.1
Al_2O_3	15.2	13.8	14.6	15.4
Fe_2O_3	2.2	2.8	1.6	4.0
FeO	3.8	3.0	4.6	2.5
MgO	3.0	2.8	3.8	2.4
CaO	.36	3.4	2.5	3.1
Na_2O	1.5	1.1	1.0	1.3
K_2O	3.7	3.5	3.4	3.2
H_2O+	3.6	2.9	3.6	5.0
H_2O	.38	.48	.21	
TiO_2	.70	.66	.68	.7
P_2O_5	.15	.17	.15	.2
MnO	.11	.11	.12	tr
CO_2	.15	2.9	2.3	2.6
S as SO_3	--	3.6	--	.6
C (organic)	--	--	--	.8
	99.55	100.32	99.36	99.9

Analysis 1. Upper part of Bushkill Member from outcrops along Delaware, Lackawanna, and Western Railroad, just above U. S. Highway 46, 0.4 mile north of Manunka Chunk, Belvidere quadrangle, N. J. (Drake and others, 1965).

Analysis 2. Bushkill Member, lowermost part, from outcrops along U. S. Highway 46, about 0.8 mile south of Manunka Chunk, Belvidere quadrangle, N. J. (Drake and others, 1965).

Analysis 3. Bushkill Member from cut along unnumbered road about 1.4 airline miles S63°W from Manunka Chunk, Belvidere quadrangle, Pa. (Drake and others, 1965).

Analysis 4. Average shale, based on the average of a composite analysis of 51 Paleozoic shales and a composite analysis of 27 Mesozoic and Cenozoic shales, each reported by Clarke (1924).

locally, associated with pillow basalts, siliceous shale, conglomerates, and graywackes in the "exotic" so-called Martinsburg of the Hamburg area, Pennsylvania. If rocks of this area prove to be related to normal Martinsburg, this would support the above interpretation as well.

Sedimentary features, such as graded bedding, small-scale cross stratification, convolute laminations and sole marks abound in parts of the Martinsburg, and suggest that turbidity currents have been important factors in sedimentation. McBride (1962) studied these features, flute casts, and substratal lineations in the Martinsburg between southeastern New York and central Virginia. He believed that the present outcrop belt is essentially parallel to the longitudinal direction of the depositional (presumably geosynclinal) basin, and that paleocurrents flowed parallel, transverse, and oblique to this northeast-southwest direction. He concluded that transverse turbidity currents originated on the shelf edge of a landmass located southeast of the present outcrop belt, and traveled down the sub-sea slope toward the axis of the basin, where many changed course and flowed longitudinally. Behre (1933) also believes that cross-stratification in slate belt sandstones indicates westward moving currents.

The original Martinsburg sediments were mixtures of quartz silt and sand, clays, micas, and accessory minerals. Lithic fragments of claystone, carbonates, graywacke, and chert are common in the low-rank graywackes of the Ramseyburg Member. This member does not occur in the Nazareth quadrangle, however, and the very small sample of Martinsburg that has been studied there does not justify sweeping statements about the origin of the detritus. Suffice it to say that the source area appears to have been a terrane of varied lithology, including carbonates, low-grade metamorphic rocks, and probably acid plutonic rocks. Many lithic fragments (claystone, graywacke) may have been derived from the Martinsburg itself.

The apparent fact that many Martinsburg paleocurrents trend westward is very significant in terms of Appalachian paleogeography and tectonic history. Earlier, in the discussion of the Early Cambrian Hardyston Quartzite, it was noted that in Hardyston-equivalent units the paleocurrents trend eastward or southeastward. Hardyston deposition was followed by a period of carbonate deposition spanning Middle Cambrian to Late Ordovician time. The argillaceous cement rock facies of the Jacksonburg signals a return to the deposition of dominantly terrigenous sediments and a shift from sea-marginal tidal and shallow marine shelf environments, presumably related to the Appalachian craton, to offshore, principally shallow open-sea to relatively deep-sea environments related to the main geosynclinal basin.

Thus, the Martinsburg, and probably to some extent the upper cement rock facies of the Jacksonburg, represent, as far as the Paleozoic of eastern Pennsylvania is concerned, the beginning of Appalachian Paleozoic sedimentation from an eastward source. This source presumably was activated by tectonic forces that either raised it above the sea for the first time or in some way moved a previously existing landmass laterally into such a geographic position as to affect Appalachian sedimentation. It need not have been particularly close to the depositional area in Martinsburg time because turbidity currents could have moved the sediment a considerable distance, perhaps several tens of miles. Earthquakes associated with tectonic activity could have acted as triggering agents for the currents.

Several studies (Pelletier, 1958; Yeakel, 1962; Meckel, 1967) directed at important Paleozoic stratigraphic intervals in the central Appalachians have shown that this eastern source area persisted throughout Paleozoic time. The nature of the sediments involved suggest, like the Martinsburg, a terrane of varied sedimentary, acid plutonic, and low-grade metamorphic rocks. In view of the mineralogy, thickness and lateral persistence of the Martinsburg throughout the central Appalachian basin, as well as the nature and volume of sediment throughout the Appalachian Paleozoic section that appears to have an eastern source, the source could not have been simply a chain of tectonically-active volcanic islands. Undoubtedly, volcanic activity was a factor, as is apparent from the bentonites that occur in the Jacksonburg and other Ordovician Appalachian rocks, but a major landmass is required as well.

Quaternary System

Glacial Deposits

Patches of glacial drift cap bedrock throughout the Nazareth quadrangle north of the Lehigh River. These deposits consist of dark yellowish orange, moderate yellowish brown, and grayish orange pebbly to bouldery silty clay and clayey silt. They are poorly consolidated, poorly sorted, deeply weathered, and lack stratification. Angular to round pebbles and boulders up to 500 mm in diameter are abundant throughout the deposits and are common on the surface as float. Most boulders are weathered, iron-stained sandstones, siltstones and quartzites that are similar to lithologies in the Silurian Shawangunk Conglomerate and Bloomsburg Red Beds which crop out eight miles north of the quadrangle boundary and from which the pebbles and boulders presumably were derived.

Topographic relations indicate that glacial drift is a maximum of 50 feet thick in the Nazareth quadrangle, but mostly much less. Topography on drift deposits typically is flat to gently rolling, and is not particularly distinctive as a mapping tool. No distinctive constructional topographic forms (eskers, kames, etc.) that are typically associated with glacial deposits have been observed in this drift.

The deposits described above occur sporadically for several miles south of the terminal moraine and other well-developed glacial features associated with the Wisconsin ice sheet in Pennsylvania and New Jersey, and presumably pre-date these features. These deposits of older drift south of the Wisconsin terminal moraine have long been recognized (Salisbury, 1892, 1902; Leverett, 1934). Leverett (1934) thought that most of these extra-morainic deposits result from the Illinoian glacial advance and that some may date from the Kansan advance. The deposits in the Nazareth quadrangle would be included in what Leverett (1934) and other workers considered to be Illinoian in age.

The deposits of glacial drift in the Nazareth quadrangle are considered herein as pre-Wisconsin, without assignment to a specific glacial advance. The correlation with the type Illinoian of the upper Mississippi Valley is very uncertain and use of that name for these deposits does not seem justified by the data available at this time.

Holocene System

Alluvium

Alluvial deposits occur in and along the Lehigh River and in some smaller streams in the area. The deposits generally are small and thin. They are largely composed of unconsolidated silt and sand, with lesser gravel.

Chapter IV

STRUCTURAL GEOLOGY

Introduction

The structural relations within and between the Precambrian and lower Paleozoic terranes have been debated for generations. Resolution of the problems has been hindered by the complexity of the geology and by the general scarcity of rock exposure that leaves much room for argument. Historically the lower Paleozoic rocks of the Great Valley have been thought to represent a regional synclinorium, whereas the Precambrian rocks of the Reading Prong have been considered to be the core of a first-order anticlinorium.

As noted in the first chapter, the Paleozoic rocks could have been subjected to the Taconic, Acadian, and Appalachian orogenies. If so, their structural characteristics, and to some extent the lithologies themselves, bear the cumulative effects of these events. Perhaps the greatest problem in these rocks is to determine the individual and cumulative results of each of these orogenies.

Internal Structure of the Precambrian Rocks

Foliation

Most of the Precambrian rocks of the Nazareth quadrangle are marked by a conspicuous foliation that results from parallel layering of minerals or mineral aggregates, and (or) orientation of platy or elongate minerals, commonly micas, hornblende, or pyroxene. The foliation typically is parallel or nearly parallel to lithologic boundaries.

The foliation may represent compositional differences in the original sedimentary or igneous rocks, it may be relic cleavage, or it may represent both relic cleavage and relic bedding if the rocks are isoclinally folded, as many clearly are. Some foliation results from injection of alaskitic material.

In other areas of the Prong, it has been observed by numerous geologists that foliation and lithologic layering are parallel. Most geologists feel that the compositional layering is relic bedding. Drake (1969) has observed in some places that layering and foliation are not quite parallel, and he feels that although foliation is mostly parallel to bedding, it probably also parallels cleavage, and is essentially transposed bedding. Moreover, he feels that the lithologic units are the limbs of sheared-out isoclinal folds. This aspect of the Precambrian rocks cannot be tested with the limited exposures available for study in the Nazareth quadrangle.

Lineation

Lineation commonly is expressed as a subparallel alignment of elongate minerals such as hornblende, sillimanite, and quartz, and as mineral streaks, and crinkled or fluted rock surfaces. It is a common feature of gneissic rocks in the Nazareth quadrangle, where it is oriented generally parallel to macroscopic fold axes.

Regionally, lineation plunges gently to moderately northeastward (Hotz, 1953; Sims, 1958; Drake, 1969) and most are parallel to the axes of both major and minor folds. Thus they represent the tectonic _b_ axis.

Joints

Joints are conspicuous features of most exposures of Precambrian rocks. Two joint sets are prominent. The best-developed set is transverse to the local and regional foliation and fold axes, in the _a-c_ tectonic plane. This set strikes northwestward. Dips typically are steep, ranging between 75° and 90°. Joint surfaces are smooth or locally slickensided.

Longitudinal joints are parallel to local and regional fold axes, and to the mineral lineations. This set thus strikes northeastward. A poorly-developed diagonal joint set is present in some exposures, but it has no well-defined regional trend.

Joint spacing is variable. Many joints, especially the transverse set, are filled with pegmatite, epidote, or a very thin film of chlorite.

Faults

Poor exposure of rocks in the Precambrian terrane limits recognition of faults. Sheared rocks are attributed to faulting, and shear zones, presumably fault zones, are mappable where shearing is extensive and traceable. One such zone was mapped on Morgan Hill in the Easton quadrangle (B. L. Miller, in Miller and others, 1939; Drake, 1967a) and was recognized and mapped by the writer in the Nazareth quadrangle. The faults within the Precambrian terrane are recognizable largely because they affect the lower Paleozoic rocks, and the Precambrian-lower Paleozoic contact. Relative displacement can be assigned only where lower Paleozoic rocks are involved.

Folds

Attitudes of foliation define folds of various magnitudes throughout the Precambrian terrane, and many such folds can be mapped where exposures are adequate. Folds are also mappable where float is distinctive and abundant. Amplitudes and wave lengths vary from a few inches to a few miles. Sims (1958) reported a mappable fold nearly seven miles long and more than a mile wide in the Precambrian rocks of the Dover district, New Jersey.

Most folds plunge northeastward parallel to mineral lineation. A variety of fold types is present in the Prong, including upright asymmetrical, isoclinal, and recumbent. Folds asymmetrical or overturned to the northwest appear to be the most common type.

The only fold mappable at the quadrangle scale in Precambrian rocks of the Nazareth quadrangle (pl. 1) involves microperthite alaskite in an apparently phacolithic relation to the enclosing gneisses. It is only mappable on float however.

Chronology of Deformation in Precambrian Rocks

Structural relations within the Precambrian rocks indicate that only one period of plastic deformation affected these rocks. Most observable macroscopic structural features appear to be related to the same Precambrian event. In some cases the structures are related to Paleozoic deformation. The Precambrian event must predate the deposition of the younger Precambrian sequence of weakly metamorphosed arkoses, ferruginous quartzites, and purple volcanic rocks, because the metamorphic grade and apparent structural style of these rocks throughout the region are inconsistent with the granulite facies rocks that constitute the bulk of the Precambrian. At this time the younger sequence is too poorly known to postulate a second Precambrian deformation.

In contrast to the structural evidence for a single Precambrian deformation is the previously cited rather clear radiometric evidence for two periods of metamorphism, at about 1150 m.y. and 835 m.y. B.P. Presumably the foliation and plastic folding dates from the younger event because this event is very persistent throughout the New Jersey-New York Highlands, and has been detected at the southwestern extremity of the Reading Prong as well. Intrusive rocks are believed by most to have been syntectonically emplaced at this time. The radiometric age determinations from which this date is derived were mostly done on biotite, a major contributer to the foliated fabric. If there were one or more earlier Precambrian deformations, it appears that the structural evidence for it has been destroyed by the latest deformation.

On the other hand, in New Jersey, Drake (1969) has observed a sheet of metasomatic albite-oligoclase granite that contains skialiths of isoclinally folded amphibolite. The granitic body is not dated, but there is no evidence of very young metasomatic activity in the Reading Prong-New Jersey Highlands terrane, so presumably the granitic body dates from the 835 m.y. event. Clearly the amphibolite was folded prior to replacement, but the time of deformation is not known.

Faulting in the Precambrian rocks occurred in both Precambrian and Paleozoic time. Faults that displace the Precambrian-lower Paleozoic contact or otherwise involve the lower Paleozoic rocks obviously must be Paleozoic or younger. The faults that involve only Precambrian rocks are indeterminate with regards to age, and may be Precambrian or perhaps younger. As no Paleozoic rocks are directly involved, it is impossible to assign relative age of such faults at this time.

Structure of the Paleozoic Rocks

Introduction

The structures of the lower Paleozoic rocks of the Lehigh Valley are many and varied. They occur on at least four different scales, but all are interrelated both geometrically and kinematically. These structures may be observed and discussed in terms of what can be seen in thin-sections, single outcrops or road-cuts, what is mappable at the quadrangle scale, and what pattern emerges on a multi-quadrangle, regional scale. The discussion that follows will begin with the description of smaller scale structures that occur at the outcrop and quadrangle scales. Subsequently the regional structural relations will be discussed.

Folds

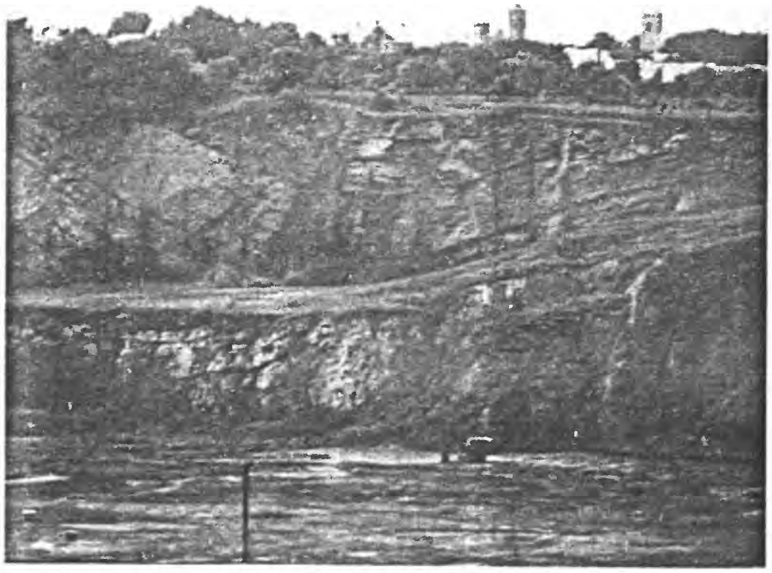
A broad spectrum of fold types and styles is present in the lower Paleozoic rocks of the Nazareth quadrangle and elsewhere in the Lehigh Valley. These range in size from tiny crinkles that can be observed only under the microscope to large folds that are areally prominent at the quadrangle scale, for example the prominent syncline that underlies the town of Nazareth.

The folds are difficult to characterize in a general way because most are hybrid types that geometrically are neither purely concentric, purely similar, nor purely disharmonic. Similarly, the mechanical processes by which such folds formed probably are mostly combinations of bedding slip and flow. Perhaps the most important aspect of folding in the Lehigh Valley is the role that lithology apparently has played in determining fold geometry and, by inference, in governing the balance of processes that formed the folds.

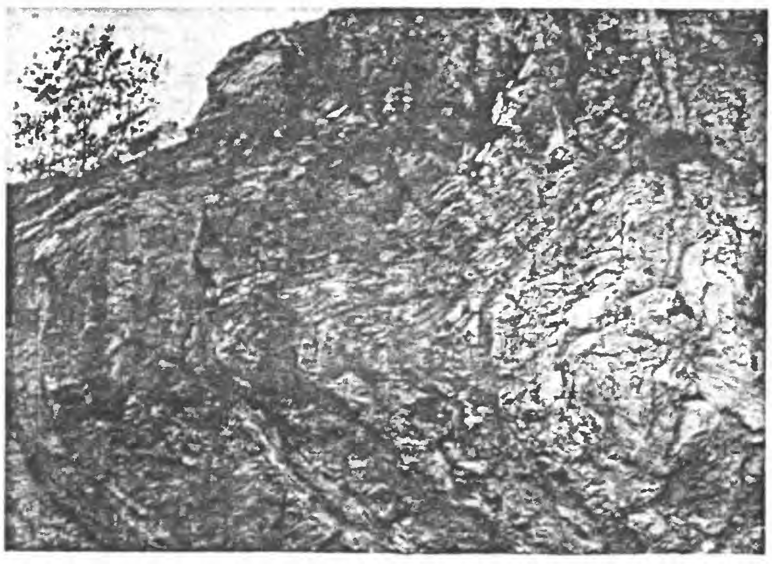
Concentric folds occur in the thicker, very competent dolomitic rocks of the Leithsville, Allentown, and Lower Beekmantown, and in the thickly bedded coarsely crystalline limestones of the Jacksonburg cement limestone facies. These folds range from symmetrical to slightly overturned, and typically are open. Wave-lengths vary from a few inches to perhaps 1,000 feet. Most folds of this type are observable only in road and quarry cuts, and are at the smaller end of the size range. Slickensides and grooves are common on bedding surfaces and are oriented perpendicular to fold axes. These indicate that flexural slip was the most important process in the development of these folds. Figure 38A illustrates a concentric fold in the Nazareth quadrangle.

Similar folds are the most common fold type in the Lehigh Valley area. These folds occur in thin to medium bedded Cambrian and Ordovician dolomitic rocks, and in some interbedded limestones and dolomites of the Epler Formation. Wave-lengths vary from a fraction of an inch to several hundred feet. Typically these folds are tight, overturned or recumbent, and many are isoclinal. These folds characteristically show some degree of peaking and thickening at the hinges, and a fracture cleavage or, less commonly, a flow cleavage may occur in some relatively less competent beds. This family of similar folds would correspond to the flexural-flow folds of Donath and Parker (1964), and result from a process of flow within

Figure 38. Photographs illustrating fold types in the Nazareth quadrangle. A, Large open concentric fold at Nazareth, Pennsylvania. Lithology is Jacksonburg cement limestone facies. B, Recumbent fold in the Leithsville Formation at Glendon, Easton quadrangle, just east of the Nazareth quadrangle border.



A



B

flexed layers. Figure 38B illustrates a large similar fold in a new road cut at Glendon, Easton quadrangle.

Many folds in the Nazareth quadrangle, and elsewhere in the Lehigh Valley, have both flexural-slip and flexural-flow characteristics. These folds, as seen in outcrop, have some relatively more competent layers that deformed by flexural slip. Less competent units in between show thickened crests, indicative of flow, and may have a well-developed cleavage.

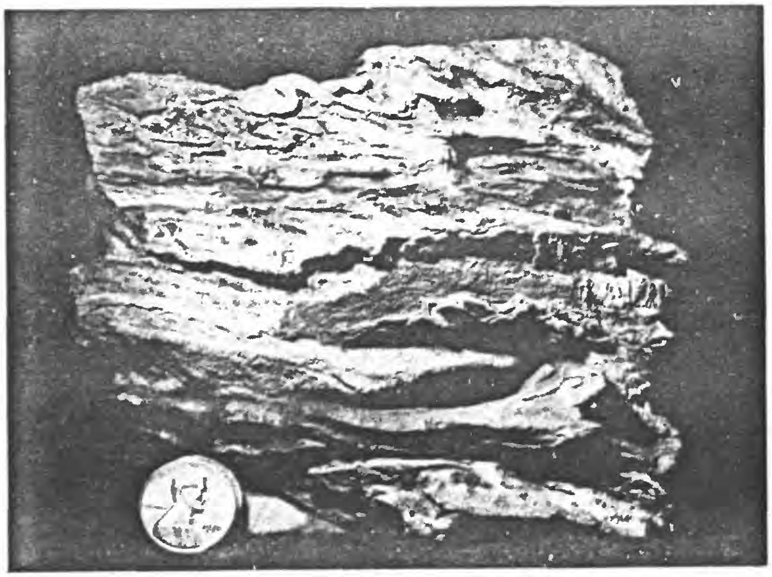
The chief kinematic distinction of the concentric and similar folds described above is that the inherent anisotropy of the rock layering (differences in thickness, crystallinity, composition, etc.) apparently controlled the processes of deformation and the structural style that resulted. Almost all of the rocks are dolomitic, and presumably such other factors as temperature, confining pressure, and structural environment were essentially equal.

The folding that occurs in many Epler limestones, the argillaceous limestones of the Jacksonburg cement rock facies, and the slaty rocks of the Bushkill Member of the Martinsburg Formation is manifestly different from that in the dolomitic rocks. Geometrically these folds range from similar to disharmonic, many are isoclinally recumbent, and hinge areas are markedly thickened and peaked. Flow cleavage is parallel to axial planes of folds and is prominent to pervasive. It cuts indiscriminantly across layers of bedding except where lithologies of strongly contrasting competence ("ductility" in the parlance of Donath and Parker, 1964) are in contact, in which case the cleavage attitude differs slightly from bed to bed.

Figures 31 and 34 show microscopic examples of such folds. The folds described above belong to the passive class of Donath and Parker (1964) because the layering had little or no control over the process of deformation and the structural style that resulted. The predominant folding mechanism was slippage or flowage across layer boundaries, and the layering acted only as a marker of such movements.

Wildly disharmonic folds may occur in rocks that contain strong contrasts in layer ductility, such as silty-banded Epler limestones, and thin, interbedded Epler limestones and dolomites. Contorted flow has resulted where more ductile rocks have been forced to accommodate to the structural movements of the relatively more competent layers. Figure 39 illustrates such disharmonic folding in hand specimens from the Epler Formation. Folding of this nature can be seen on a larger scale in several localities in the Lehigh Valley. One such locality is the western face of the Trumbauer Quarry, 1/2 mile east of the town of Nazareth. There, thickly bedded coarsely crystalline limestone of the Jacksonburg cement limestone facies overlies interbedded medium thick limestones and dolomites of the Epler Formation. The Jacksonburg limestones occupy the northward-dipping limb of the Nazareth syncline. The dip is quite regular and consistent along the outcrop belt, stratigraphic thickness of the limestones is not distorted, and the beds are uncontorted. However, the subjacent interbedded limestones and dolomites, as seen in the quarry, are so contorted and in such disarray that it is exceedingly difficult to follow a single marker bed for more than a few feet. The writer tried repeatedly over a period of four years to photograph

Figure 39. Photographs of hand specimens illustrating disharmonic folding in Lehigh Valley limestones (Epler Formation). Both specimens from the Nazareth quadrangle north of the village of Brodhead.



A

B

or sketch the relations in the face, but all attempts were relatively unsuccessful because the fresh rocks do not exhibit strong contrasts between beds, the face is actively quarried, and the picture changes with each blasting.

Axes of folds observed in the lower Paleozoic rocks of the Nazareth quadrangle mostly plunge northeastward at angles that range from horizontal to 20 degrees. The most prominent axial orientation is 0-10°, N 60 to 65° E. A few very gentle southwestward plunges occur on this trend also. Less prominent are fold axes that plunge at about 10°, N 82 E. The significance of this trend is not clear because the data are not very abundant. The writer knows of no case in which interference of these fold trends has been recognized.

Studies by A. A. Drake, Jr. (1970), summarizing work on several quadrangles in the Delaware Valley, indicate three principal axial directions. These are ± horizontal, N 60° E; 5°, S 44° W; and 12°, N 80° E. Two of these directions closely correspond to the principal plunge directions recorded in the Nazareth quadrangle. The S 44° W direction does not represent a strong trend in the Nazareth quadrangle, however; although a few readings in that sector were recorded, the spread is too broad to define a trend. Drake (1970) tentatively considers the N 60° E folds to be first generation because they are the most abundant. These are followed in order of abundance by the S 44° W folds, which he considers to be second generation, and by N 80° E folds, which he believes may be third generation. However, there is no direct evidence for such a sequence, as Drake points out, and the writer feels that such conclusions must be justified by further studies.

Faults

Thrust faults of minor displacement are common features of most larger outcrops, and road and quarry cuts in the Nazareth quadrangle. Probably they are very numerous throughout the area, but poor exposure limits their recognition and mappability.

The Stockertown Fault (Davis and others, 1967) is the most important fault in the area. It appears to be a major thrust fault, and was mapped in the Nazareth quadrangle approximately 1 1/2 miles west of the center of the town of Nazareth, as shown on Plate 1, where a small body of the Bushkill Member of the Martinsburg occurs within the main outcrop belt of the Jacksonburg Limestone. The relation there is interpreted in Plate 2 as an antiformal window through the Jacksonburg that exposes the Martinsburg beneath the thrust plate. Similar relations involving the same units occur on the regional strike in the Wind Gap quadrangle (J. B. Epstein, personal communication), north of Nazareth, and in the Bangor quadrangle (Davis and others, 1967) northeast of Nazareth.

The Stockertown Fault was first recognized by J. B. Epstein (oral communication, 1964) in the exploratory drilling records of a cement company in the southern part of the Wind Gap quadrangle. Originally the displacement was believed to be on the order of a few hundred to perhaps a few thousand feet (Davis and others, 1967) Subsequent mapping by Epstein in the Wind Gap quadrangle has revealed more extensive and complex imbricate faulting related to the Stockertown Fault, and the total displacement now is believed to be at least several miles (Drake, 1970).

Drake (1970) suggests possible manifestations of the Stockertown Fault in the Bloomsbury quadrangle, New Jersey, and in the Great Valley southwestward into the Lebanon Valley, Pennsylvania. Although the fault is only poorly known and understood, it appears at this time to be a major structural feature of the Great Valley in Pennsylvania and New Jersey. Reevaluation of structural data from studies already completed and future detailed studies specifically directed at the Stockertown Fault and related structures should provide more data and a much better understanding of this feature.

Other faults of less importance are shown in Plate 1. These are but a fraction of the faults that are actually present, but unless formational contacts are displaced, perhaps to a considerable extent, such features are not recognizable because of poor exposure. Faults in the Jacksonburg Limestone and the Martinsburg Formation are exceptions to this statement. In these units, even apparently insignificant faults observable in road-cuts and quarries commonly are filled with veins of columnar quartz or calcite, or both. Such vein material is very distinctive, especially the quartz because it persists in float. Thus, even in areas of poor exposure intraformational faults sometimes can be traced and mapped in float; typically they are no more than a few tens of feet in length.

Cleavage

Rock cleavage is present throughout the Nazareth quadrangle and occurs in all lithologies and in all stratigraphic units. Cleavage varies somewhat in style and type according to the lithology in which it occurs, as is the case with the folds previously described. This is natural because as will be shown, the cleavage is genetically related to the folding.

There are at least two distinct cleavages present and in order to distinguish clearly between them in this discussion, the S-plane notation of Sander (1930) has been adopted. In this notation planar structures are designated as S-planes, and subscript numbers are assigned according to the relative order of development. Thus bedding, the primary structure, is designated as S_1; the first cleavage, which transects the bedding, is S_2; the second cleavage, which transects both bedding and the first cleavage is S_3; and so on. Other planar structures, such as joints, can be incorporated as necessary or desirable.

Flow Cleavage (S_2)

Flow cleavage is a conspicuous and pervasive feature of the Epler limestones, the Jacksonburg cement rock facies, and the Bushkill Member of the Martinsburg. It is characterized by flowage, recrystallization and parallel orientation of calcite and micaceous minerals. The rocks have a very strong tendency to split along directions parallel to the mineral orientation. Mineral orientation is clearly observable in thin-sections (fig. 34 and 37), but grain sizes generally are too small to verify this in hand specimens, except in phyllitic rocks in the Martinsburg.

Flow cleavage typically is the most conspicuous planar structure in the rocks in which it occurs. Traces of bedding commonly are visible on cleavage planes, especially in the Martinsburg slaty rocks. Epler limestones commonly contain thin layers of quartz silt that mark the bedding. In the Jacksonburg cement rock facies, however, bedding is almost totally obliterated, and it is not at all uncommon to walk away from an outcrop without having found bedding after more than one hour of study. This is especially true of fresh exposures.

The flow cleavage is parallel to the axial planes of observable folds and is generally rather flat, reflecting the regional recumbency of folds in the region. Most folds are isoclinal as well, and as a result bedding and flow cleavage commonly are essentially parallel, except in fold hinges, where they intersect at high angles. Equal-area plots of poles to flow cleavage (S_2) show a strong orientation of strikes at about N 60 E, and a concentration of dips between 10° and 30° southeast, as shown in Figure 40.

In the dolomitic rocks and the thick limestones of the Jacksonburg cement limestone facies of the Nazareth quadrangle the S_2 cleavage is a spaced or fracture cleavage (essentially closely spaced jointing) rather than a flow cleavage. This cleavage is approximately parallel to axial planes of folds, but tends to be fanned more than the flow cleavage described above. On a statistical basis the strike of this cleavage is similar to that of the

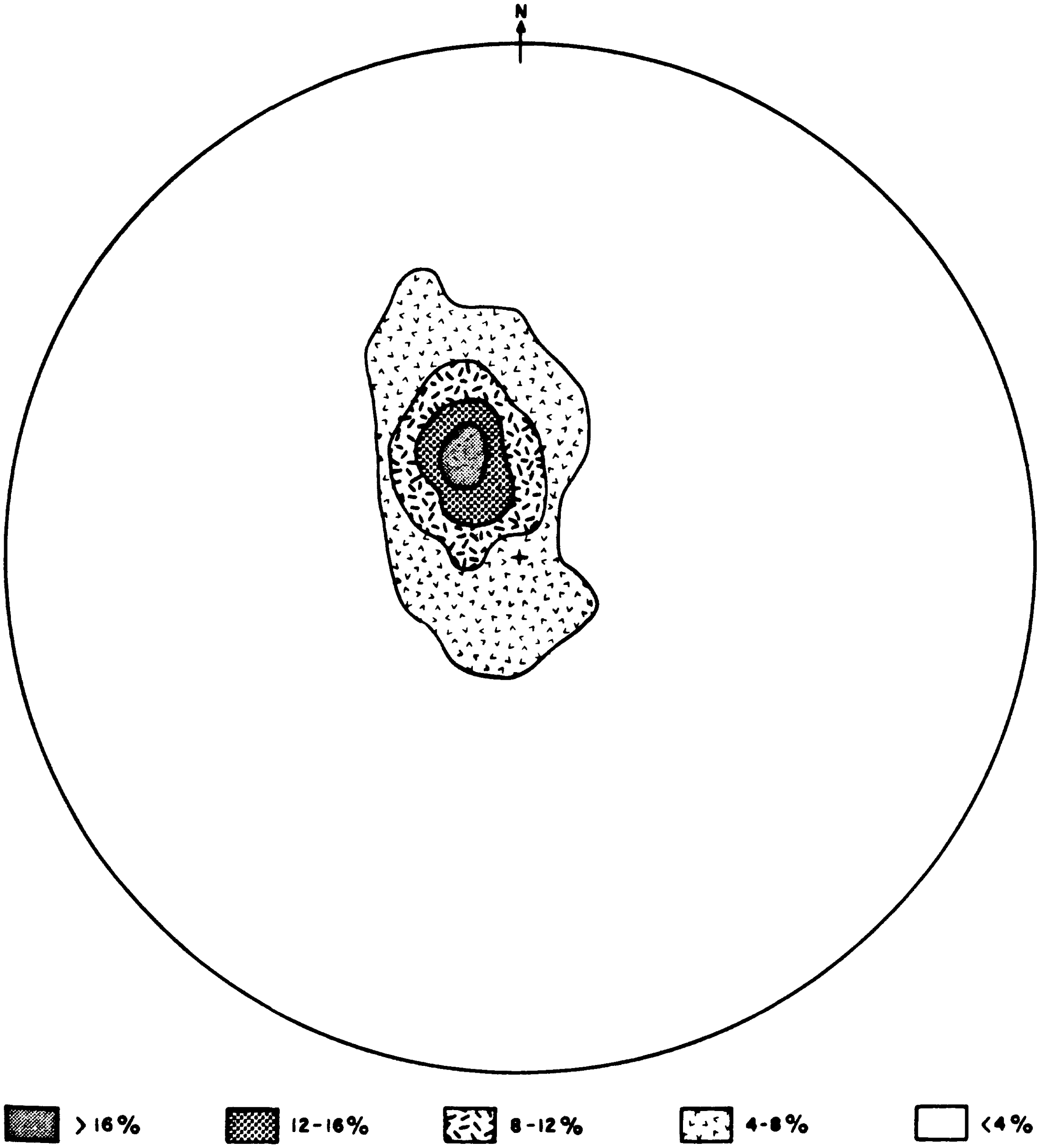


Figure 40. Equal area lower hemisphere projection of 226 poles to flow and fracture cleavage (S_2) in Paleozoic rocks of the Nazareth Quadrangle, Pennsylvania. Contours at 16-12-8-4 percent.

flow cleavage, but the dip is somewhat steeper. The spread of dips plotted on the equal area net (fig. 40) reflects this tendency. The relatively greater competence of the dolomites and thick crystalline limestones is believed to account for the lessened tendency to form flow cleavage. However, in areas of more intense deformation, thin sections show that an incipient flow cleavage is developed. Figure 41 shows the typical development of flow cleavage in rocks of the Nazareth quadrangle.

Crenulation Cleavage (S_3)

Crenulation cleavage, used in the sense of deSitter (1964), is a secondary cleavage that deforms planes of flow cleavage by causing a wrinkling or crenulation of the S_2 cleavage plane. Many geologists refer to this phenomenon as slip cleavage. In many cases, where layering is thin and discernible, flow cleavage and bedding are clearly off-set on planes of crenulation cleavage. Offset on a single plane typically is only a fraction of an inch. In exposures where such displacement occurs, recognition and measurement of the crenulation cleavage is no problem. In many exposures, however, one is acutely aware that the flow cleavage (S_2) is deformed but the crenulation plane (S_3) is all but impossible to locate because the wrinkles and crenulations are of such minor character. Figures 34 and 37 show the microscopic character of crenulation cleavage (S_3). It is also apparent in these photomicrographs that the crenulation or slip cleavage itself may pass into a second flow cleavage that is characterized by recrystallization and parallel orientation of

Figure 41. Photographs of outcrops showing development of flow cleavage (S_2) in rocks of the Nazareth quadrangle. A, Leithsville Formation along Bull Run, south of village of Redington. Bedding dips about 45 degrees to right of picture, cleavage dips to the left. B, Cleavage in the Bushkill Member of the Martinsburg Formation. Marking pencil is on flow cleavage (S_2), lead pencil is on bedding (S_1). Note that cleavage is sub-horizontal, flatter than bedding. Outcrop located on Route 946, 2 miles west of the town of Nazareth.

A

B

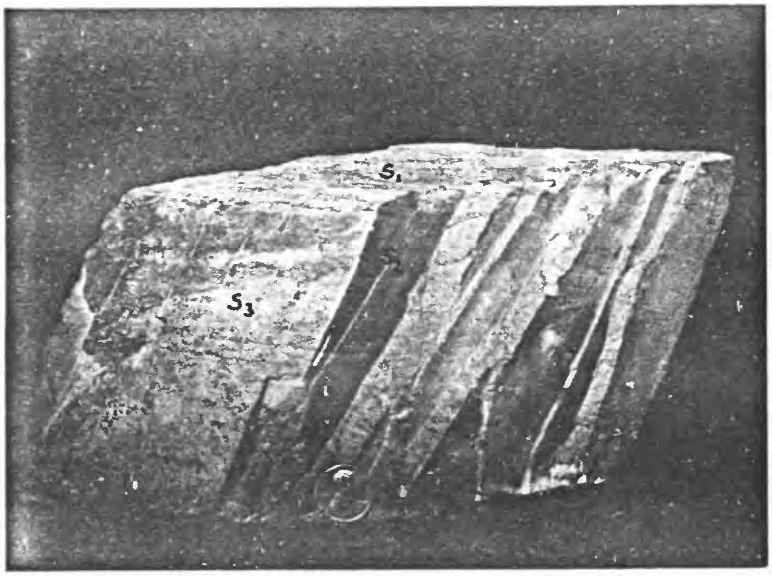
micaceous minerals. Figure 42A pictures a hand specimen that contains two flow cleavages.

In the Nazareth quadrangle crenulation cleavage (S_3) is only weakly developed but has been recognized in Epler limestones, the cement rock facies of the Jacksonburg, and in slates of the Bushkill Member of the Martinsburg, all of which units contain a well developed flow cleavage (S_2). Crenulation cleavage has not been recognized in the dolomitic units, however.

In thin sections of rocks from the Nazareth quadrangle crenulation cleavage can be seen in association with microscopic folds in the flow cleavage, and is essentially parallel to the axial planes of these folds. Macroscopic folds in flow cleavage with related crenulation (slip) cleavage parallel to the axial planes have been observed in the Nazareth quadrangle in the above-mentioned stratigraphic units, but they are not at all common. In other areas of the Lehigh and Delaware Valleys such relations are well developed at the outcrop level. Figure 42B pictures one such outcrop of the Jacksonburg cement rock facies on the grounds of the Saucon Valley Country Club, in the Hellertown quadrangle. The folds are in flow cleavage (the strong planar element) and planes of crenulation (slip) cleavage cut obliquely through the folds, parallel to the axial planes. It seems clear that folds in flow cleavate (S_2) are genetically related to crenulation cleavage (S_3) and are structures of second generation Paleozoic deformation.

Crenulation cleavage (S_3) is oriented essentially parallel in strike to flow cleavage (S_2) but is generally steeper in dip.

Figure 42. Photographs illustrating the nature and development of crenulation cleavage (S_3) in rocks from the Nazareth quadrangle and vicinity. _A_, Hand specimen from the Martinsburg Formation in which both S_2 and S_3 are flow cleavages. Sample collected near the Fish Hatchery, Bloomsbury quadrangle. _B_, Outcrop of the Jacksonburg cement rock facies in which crenulation cleavage (S_3) folds flow cleavage (S_2), and is parallel to the axial planes of the resultant folds in S_2. Outcrop on grounds of the Saucon Valley Country Club, 3 miles southwest of Hellertown, Pennsylvania.



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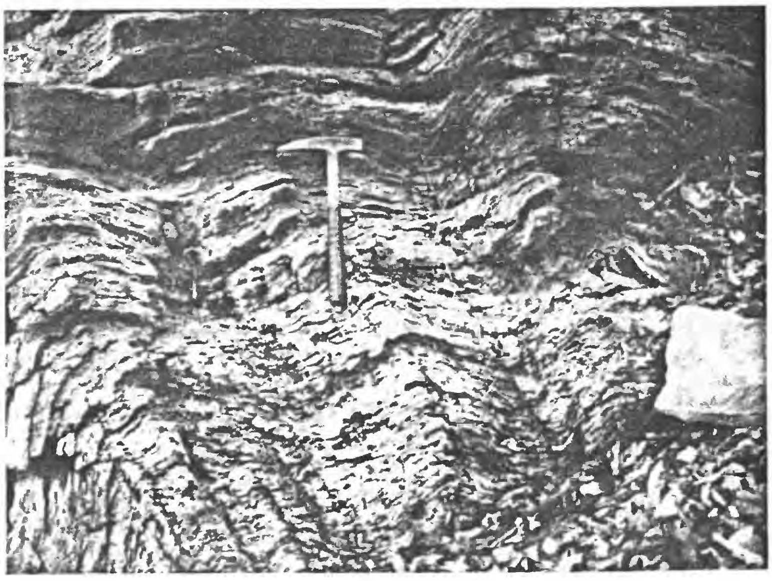


B

Figure 43 is an equal area plot of crenulation cleavage recorded from the Nazareth quadrangle. Sparse data on axes of folds in cleavage (S_2) indicate that these folds plunge gently northeastward or southwestward, and are essentially parallel in orientation to folds in bedding (S_1).

Joints

Joints are ubiquitous in the Lehigh Valley lower Paleozoic rocks. They occur in all units and all lithologies. Joints throughout the area typically are planar and smooth. Locally, in areas of more intense deformation they are slickensided or streaked. Also locally, the joints may be filled with calcite or quartz. Strikes vary widely, but typically dips are steep, mostly between 65° and vertical.

In mapping, questions may arise as to what fractures constitute joints and which are to be treated as fracture cleavage (essentially closely-spaced jointing). All such fractures with spacing of 1 inch or less were treated as cleavage, regardless of the lithology in which they occurred. Spacings from 1 to 6 inches were treated as cleavage or jointing depending on the lithology. In the less competent rocks such spacing was treated as jointing because such rocks in the area typically have a well-developed flow cleavage, and locally a crenulation cleavage. In the more competent thickly-bedded or coarsely crystalline rocks, mostly dolomites, well-developed parallel fractures with spacing of up to 6 inches were treated as fracture cleavage. Admittedly, this approach is somewhat subjective and requires judgment.

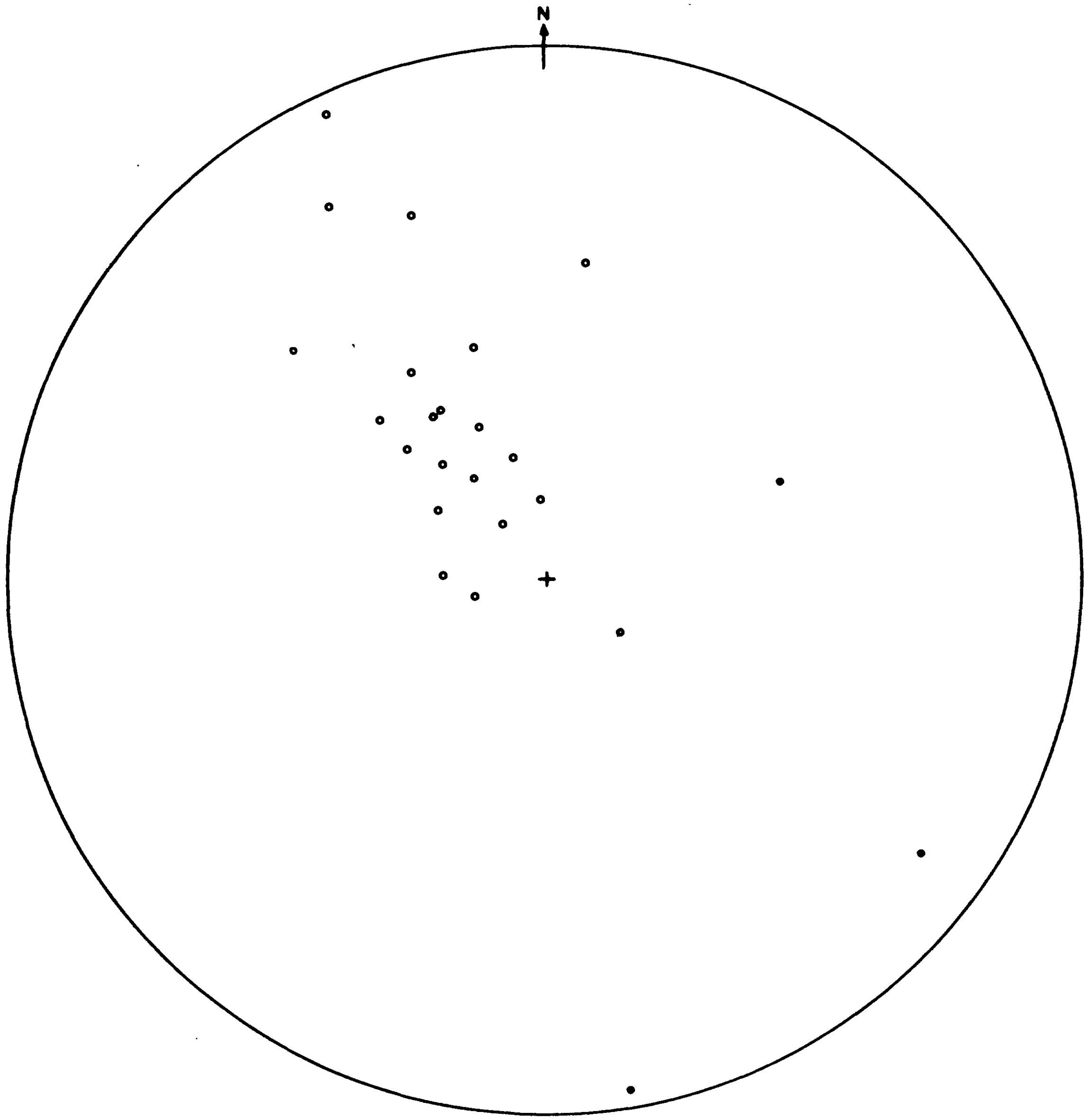


Figure 43. Equal area lower hemisphere projection of 24 poles to crenulation cleavage (S_3) in Paleozoic rocks of the Nazareth quadrangle, Pennsylvania.

However, the writer feels that a more flexible approach is justified in such a sequence of stratigraphically related rocks that obviously responded differently to tectonic stresses.

There is no strong preferred orientation of joints recorded from the Nazareth quadrangle. In a rigid statistical sense there is no significant orientation at all, and in a test of orientation of over 600 joint readings, the weight of the evidence favors isotropicity. This may be a matter of sampling however. In any given outcrop anywhere from one to six different sets of nearly parallel surfaces might occur. Some are well developed, some are not. All joint sets were recorded, and all were part of the raw statistical data. Joints can develop from a variety of causes and at various stages during the process of deformation, and many joints in a given structural-stratigraphic setting could reflect purely random local conditions of stress, individual rock qualities and layer anisotropy. It is possible that the method of sampling the Nazareth joints homogenized the data by including a large number of joints of essentially random distribution, and that the trends of the tectonically important joints are obscured by the overprint of the random sets. Also, it is possible that the apparent randomness is the result of homogenization of sets of different genesis, each of which might have a statistically significant, nonrandom distribution. The problem may be the inability to distinguish the various sets in the field.

When the same data are plotted on an equal area net and contoured in the usual manner, as shown in Figure 44, there does

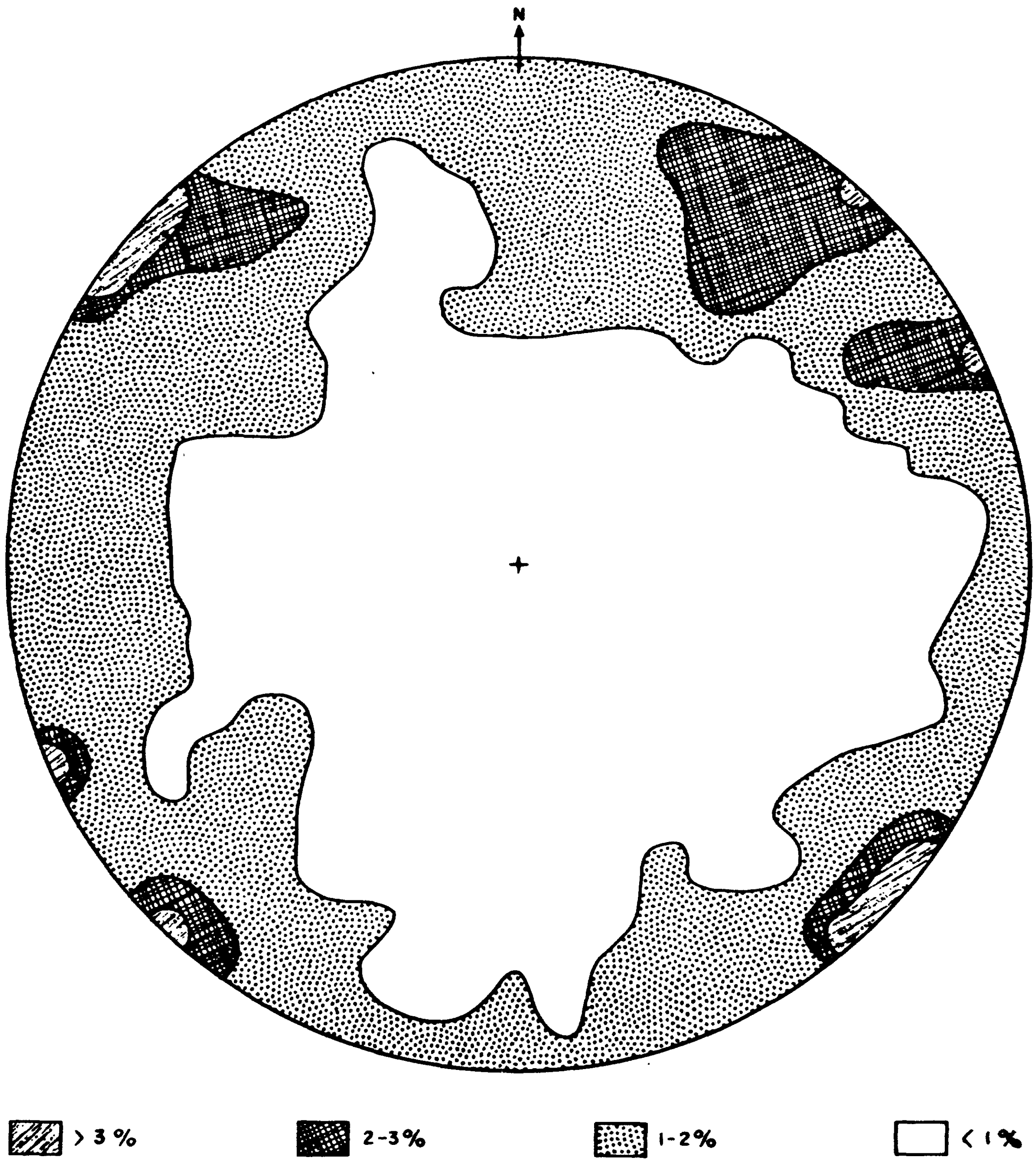


Figure 44. Equal area lower hemisphere projection of poles to 596 joints in Paleozoic rocks of the Nazareth quadrangle, Pennsylvania. Contours at 3-2-1 percent.

appear to be a preferred orientation of two joint sets. One set strikes northeastward, parallel to the regional structure, and dips steeply northwest or southeast. The other set is weakly developed and strikes northwest and dips steeply northeast or southwest. These joint sets have been recognized by almost every worker who systematically recorded joints in the Great Valley. The fact that these sets bear a consistent relationship to the regional structural grain suggests that they are, indeed, primarily related to the regional tectonic framework.

Lineation

Lineation includes all linear rock structures and usually results from movements within a rock body and (or) mineral growth under stress. Lineation observed in the Great Valley rocks of the Nazareth quadrangle comprises fold axes (described in section on folds), streaks and striae on cleavage and bedding planes, elongate oöids, intersections of cleavage (S_2) and bedding (S_1) planes, intersection of cleavage planes (S_2 and S_3), boudinage and mullion structures, pyrite grain elongation, and wrinkles on bedding and cleavage planes (S_2) related to crenulation cleavage (S_3).

Elongate oöids occur in some oölitic dolomites of the Allentown. They have been observed oriented parallel to both bedding and cleavage. Some elongation may result from geostatic rather than tectonic deformation, as was discussed in the section on the Allentown Dolomite. No systematic study was made of these features as they occur only sporadically.

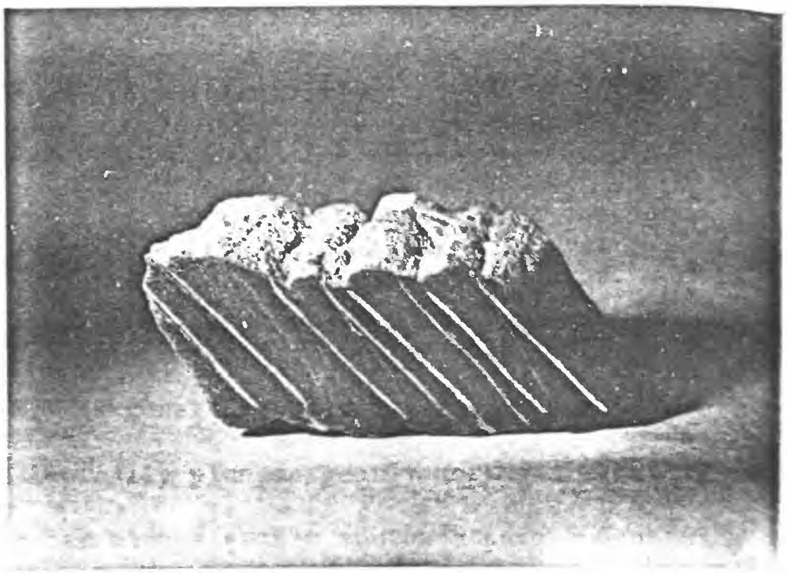
Streaks and striae occur on bedding and cleavage (S_2) planes. They are especially common on bedding planes of the very competent rocks (thick coarsely crystalline limestones and dolomites). Without exception they are oriented parallel to the dip direction of the surface on which they occur, perpendicular to the b tectonic axis, and parallel to the a axis. They attest to the direction of movement (slip) on the planes involved.

Boudinage and mullion structures were observed only in the Jacksonburg Limestone and in some Epler limestones and are pictured in Figure 45. Both structures produce lineations in the bedding planes that are parallel to fold axes, the tectonic b axis.

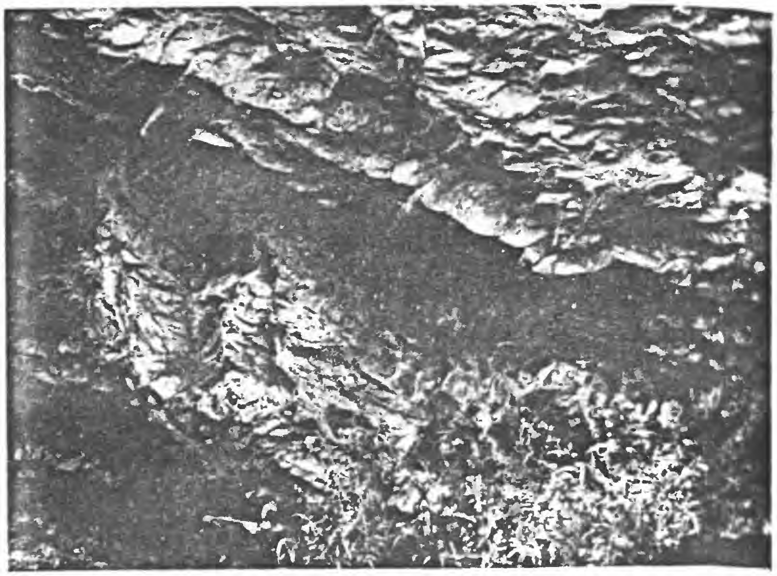
Pyrite grain elongation occurs in the Jacksonburg limestones, particularly in the bentonite horizons. The lineation thus produced is parallel to the axis of the enclosing fold.

The most important and widespread lineation in the Lehigh Valley lower Paleozoic rocks is produced by the intersection of bedding (S_1) and flow or fracture cleavage (S_2). This feature appears on bedding planes as a set of essentially parallel fractures, ruts, or striations that clearly are related to the cleavage. It also is observable on cleavage planes where traces of bedding are visible. It is especially prominent in the Bushkill Member of the Martinsburg Formation because the flow cleavage is pervasive and the bedding may be very thin. Some weathered exposures in the upper part of the Jacksonburg cement rock facies show this very well also, for example the "fence posts" pictured in Figure 34. Where bedding and cleavage (S_2) intersect at high angles, as is the case in

Figure 45. Photographs of some minor structures in rocks of the Nazareth quadrangle. A, Mullion structure in limestone bed, Jacksonburg Limestone. Sample is part of a single bed. Note filling of cleavage by calcite and the conspicuous lineation on the bedding plane. B, Boudinage in the Jacksonburg cement rock facies, forming a strong lineation on underside of bedding. Both photos on rocks from the quarry of the Nazareth Cement Company, Nazareth, Pennsylvania.



A



B

Figure 34, the relations between bedding, cleavage, and the lineation are quite clear in the field. In the Jacksonburg cement rock facies and to a lesser extent in the Bushkill Member of the Martinsburg the rocks typically are isoclinally folded and the intersections are not readily identified. This is especially true in the cement rock facies, in which bedding has been all but obliterated by the development of flow cleavage.

The lineation produced by the intersection of bedding and cleavage (S_2) is generally parallel to folds axes, both major and minor, and typically plunges gently northeastward. Some lineations of the same set plunge gently southwestward. Generally the lineations are quite flat, typically less than 5°, but ranging from 15° to horizontal. This lineation is parallel to the tectonic <u>b</u> axis because of its parallel relationship to fold axes, and is herein designated as L_1. It is a first-generation structure because of its relation to folds in bedding and to flow (or fracture) cleavage (S_2). Figure 46 is an equal area plot of L_1 lineations recorded in the Nazareth quadrangle.

Where there is a second cleavage in the group of relatively incompetent rocks, there is also a second lineation (L_2), formed by the intersection of flow cleavage (S_2) and crenulation (slip) cleavage (S_3). This lineation is not common in the Nazareth quadrangle because the crenulation cleavage is weakly developed. Nevertheless, the sparse data, shown in Figure 47, suggest a pattern of gentle northeastward plunge, approximately similar in trend to the trend of the lineation caused by the intersection of bedding

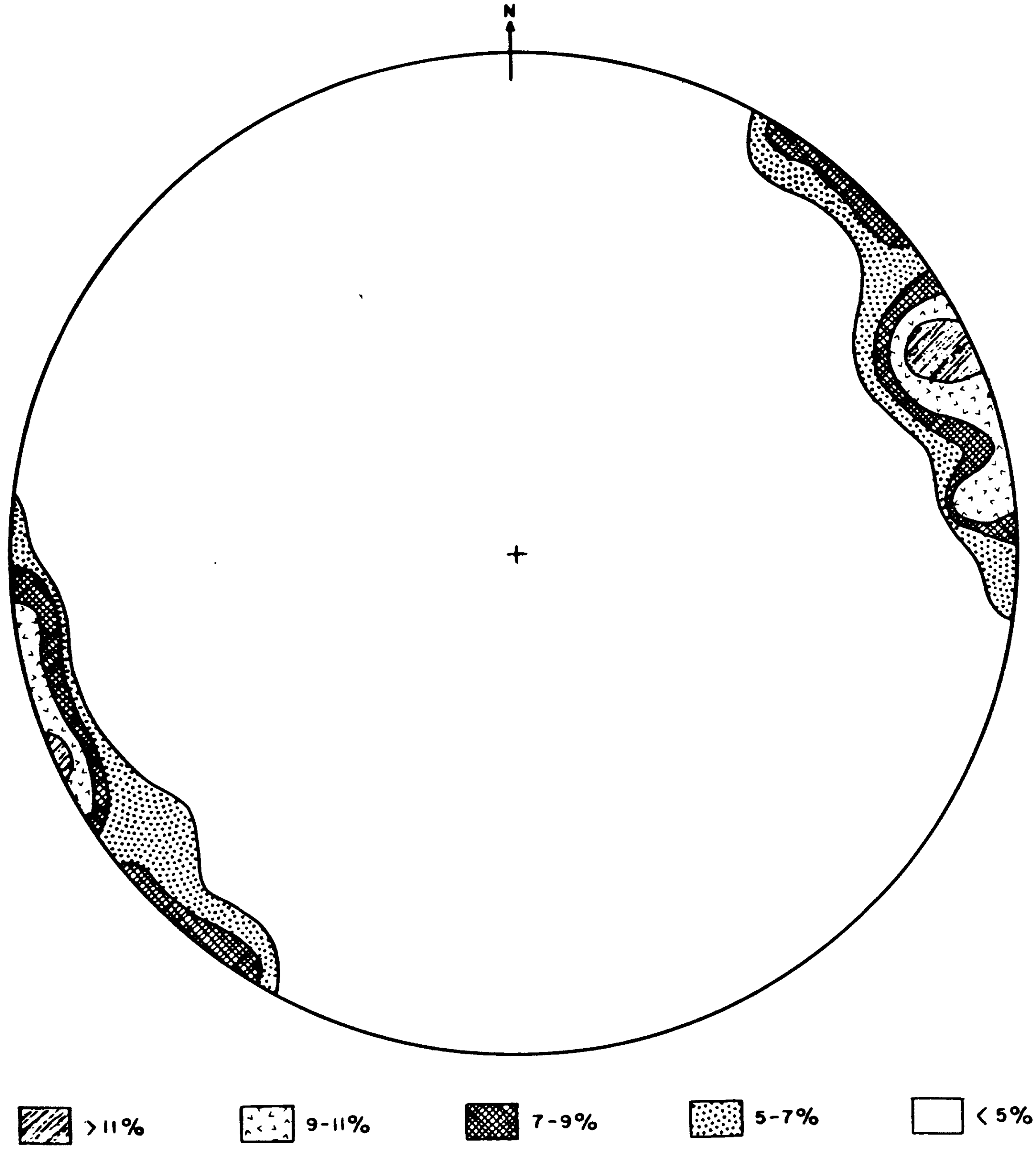


Figure 46. Equal area lower hemisphere projections of 125 lineations (L_1) from cleavage (S_2) - bedding intersections in Paleozoic rocks of the Nazareth quadrangle, Pennsylvania. Contours at 11-9-7-5 percent.

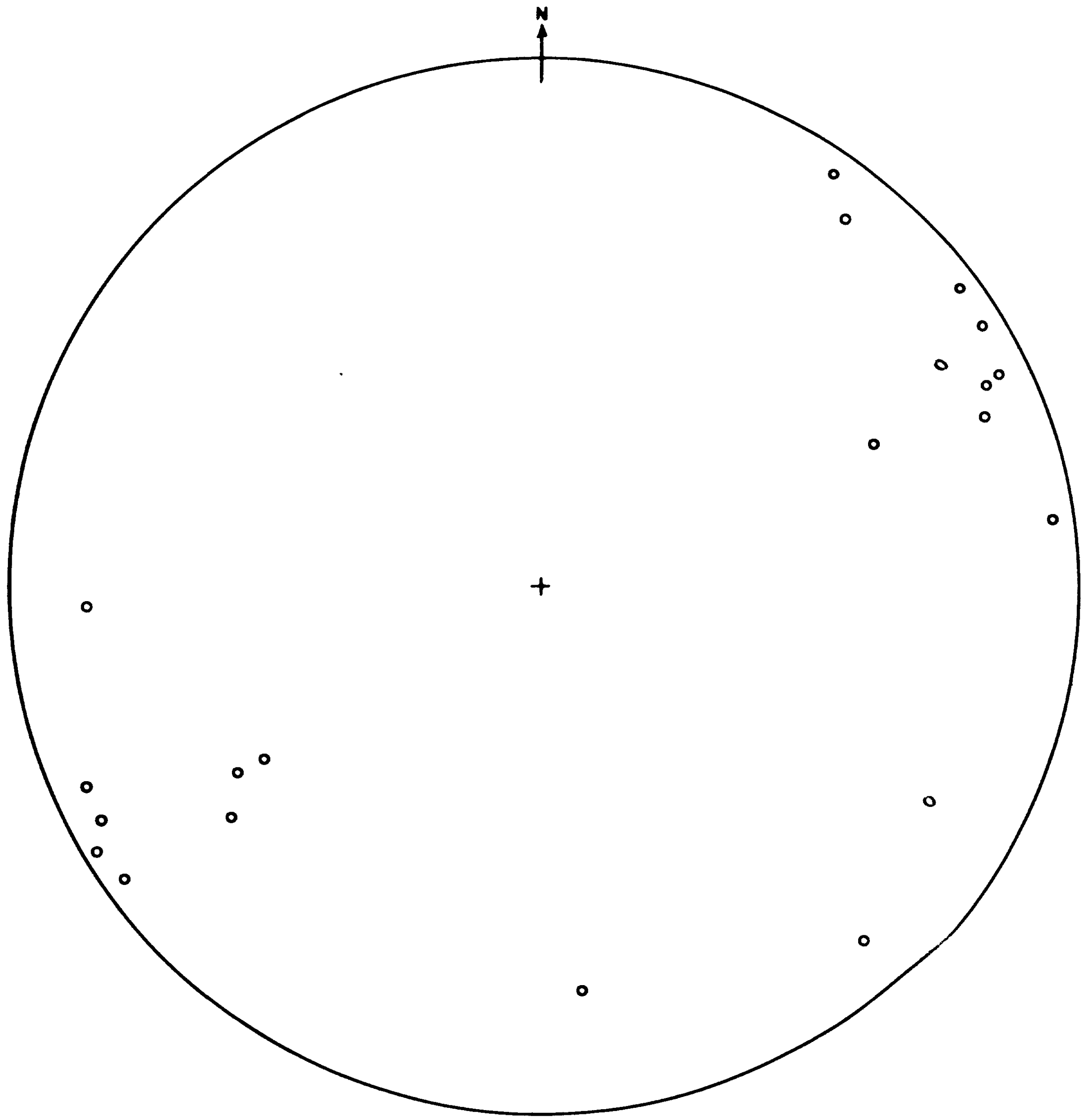


Figure 47. Equal area lower hemisphere projection of lineations (L_2) from intersections of flow cleavage (S_2) and crenulation cleavage (S_3) in Paleozoic rocks of the Nazareth quadrangle, Pennsylvania

and flow cleavage. In areas where this L_2-lineation is well developed it clearly is parallel to axes of folds in cleavage.

The Precambrian-Paleozoic Contact

The Precambrian-lower Paleozoic contact was mapped as a thrust fault throughout Nazareth quadrangle, as shown in Plate 1. The main contact occurs south of the Lehigh River and strikes northeastward. Isolated bodies of Precambrian rock also occur underlying two prominent wooded hills in the Lehigh Valley more than three miles north of the main contact, and at Chestnut Hill, along the eastern border of the quadrangle. Thrust faults were mapped around each of these bodies as well. These Precambrian outliers are probably related because they are approximately on strike with each other, and are parallel to the regional trend of the main Precambrian bodies.

Superficially, there is little specific field evidence for mapping a thrust along the main Precambrian-lower Paleozoic contact in the southeastern part of the Nazareth quadrangle. The contact, as such is not exposed, but it can be rather precisely mapped on float. A number of lines of evidence support the interpretation presented herein, however. In a general way the trace of the contact surface is parallel to the topographic contours and the contact thus appears to be gently to moderately inclined southeastward. In the Easton quadrangle (Drake, 1967a), east of Nazareth, the continuation of the thrust maintains an essentially parallel relation to topography and forms a prominent southeastward reentrant that almost closes upon itself in Morgan Valley. The same thrust fault continues southward from the Nazareth quadrangle into the Hellertown quadrangle where it bends sharply southward around the nose of the Precambrian body and traces a very irregular course into and around Saucon Valley, a large and very prominent reentrant valley floored with Paleozoic carbonate rocks with very complex structures. These relations certainly suggest that the Precambrian-Paleozoic interface dips gently to moderately southeastward beneath the Precambrian rocks and are not consistent with the traditional autochthonous interpretation of this Precambrian body as the core of an anticline.

Similar relations occur along the northern border of Precambrian rocks throughout the region and many have been detailed by Drake (1969). Of particular interest are two localities in New Jersey where examples of Precambrian rocks thrust upon Paleozoic rocks are clearly exposed. One is along a small creek 0.45 mile east of the village of Annandale where a body of amphibolite rests upon a carbonate unit that is believed to be the Allentown Dolomite. The other is 0.5 mile southwest of the village of McPherson, where Precambrian gneiss overlies the Leithsville Formation. There is a small cave here in which dolomite forms the walls and floor, and gneiss forms the roof.

In 1967 the United States Geological Survey, in cooperation with the Pennsylvania Bureau of Topographic and Geologic Survey, drilled a 476 foot hole on Rattlesnake Hill, south of Riegelsville, Pennsylvania. The hole was collared in Precambrian rocks that comprise amphibolite, serpentine, hornblende gneiss and quartzo-feldspathic gneiss. Carbonate rocks of the Leithsville Formation were encountered at a depth of 392 feet (Epstein and others, 1967). This hole was located near the south border of one of the major belts of Precambrian rock in the area. In addition, a growing number of water wells that have been drilled along the edges of Precambrian belts in Pennsylvania and New Jersey have intersected Paleozoic carbonate rocks at shallow depths. These subsurface data considered in conjunction with field relations and map patterns attest to the horizontal displacement of Precambrian rocks over the Paleozoic rocks of the Great Valley.

Further support for this interpretation is provided by geophysical studies of the Reading Prong and Great Valley areas. Aeromagnetic studies by Bromery and others (1959), and Bromery and Griscom (1967), are summarized in Figure 48. Most of the exposed Precambrian rocks are characterized by a "birdseye maple" magnetic pattern, but some bodies of Precambrian rock, for example Camelhump and Pine Top, have no magnetic expression at all. Figure 48 shows that the Precambrian body at Chestnut Hill and the main body that occupies the southeastern part of the Nazareth quadrangle are bounded on the north by strong negative magnetic anomalies. On the south side of these bodies, the aeromagnetic contours grade steeply into the areas underlain by Paleozoic rocks, and ultimately drop off into the negative anomaly that bounds the next Precambrian ridge to the south. These relations suggest that Precambrian bodies do not dip northward under the Palaozoic rocks that bound them on

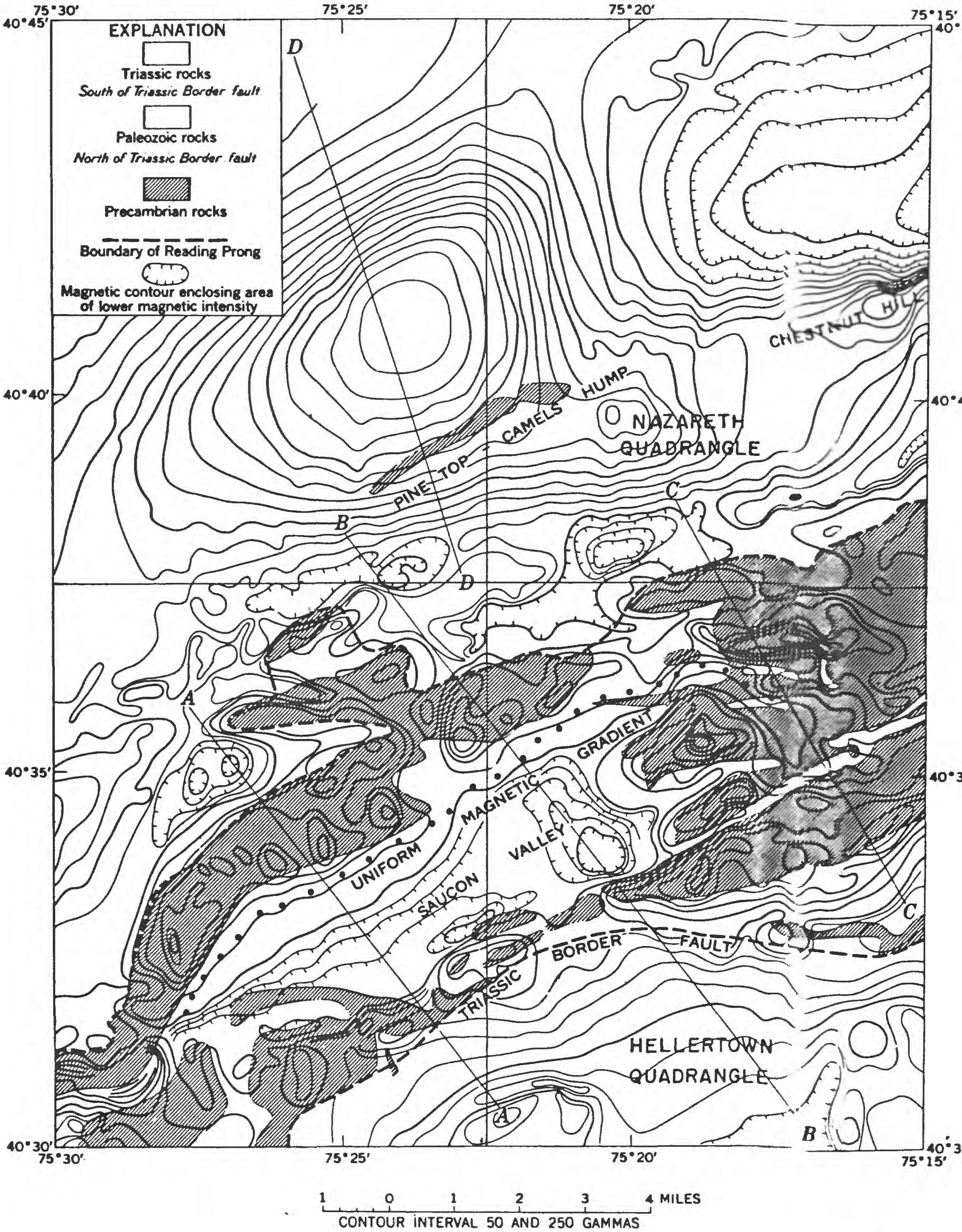


Figure 48. Aeromagnetic and generalized geologic map of the Allentown quadrangle, Northampton, Lehigh, and Bucks Counties, Pennsylvania. Modified from Bromery and others (1959) and Bromery and Griscom (1967).

the north, as they should if they were the cores of anticlines, but that the Precambrian rocks do occur for some distance under the Paleozoic rocks on the south sides of the Precambrian ridges. The Precambrian bodies must terminate at depth on the northern side of the negative anomalies, however. Bromery's magnetic studies (1960; in United States Geological Survey, 1964, 1966) also indicate that the Precambrian bodies are thin.

The aeromagnetic expression of the noses of the Precambrian rocks outlined in Figure 48 also are instructive. The Precambrian mass in the southeastern part of the Nazareth quadrangle noses out southwestward just south of the quadrangle border, where it trends straight into a negative anomaly. Other digitations of the same Precambrian body, south of the Nazareth quadrangle, are met by a strong negative anomaly in Saucon Valley. If these Precambrian ridge bodies were rooted, the ends of the ridges would be the plunging noses of anticlines that would have aeromagnetic expression along the plunge beneath the Paleozoic rocks.

The relation described above for the Precambrian rocks in the Nazareth quadrangle are not unique. The aeromagnetic studies of Bromery and Griscom (1967) show that such relations at the Precambrian-Paleozoic interface abound throughout the Reading Prong in Pennsylvania. The Precambrian ridges are thus envisioned to be rootless southeastern dipping bodies that are underlain at depth by nonmagnetic material. The interpretation of the Precambrian rocks of the Reading Prong as allochthonous is favored by most geologists currently working in the area, among them Geyer (1963), Drake

(1967a,b; 1969; 1970), Drake and others (1967, 1969), and V. E. Gwinn (personal communication, 1966).

The interpretation of the Precambrian rocks of the Nazareth quadrangle, and, by implication, the Precambrian rocks of the Reading Prong as unrooted allochthonous masses is also consistent with regional gravimetric data. Isachsen (1964) noted that the Green Mountains anticlinorium (fig. 49), a rooted thrust, lies on the axis and the steep western gradient of a regional gravity high that persists through the middle Appalachians. Westward of this high is a strong negative gravity anomaly that coincides with the Middlebury synclinorium, a known basement depression. South of the Green Mountains, the Precambrian bodies, all of which are known to be displaced westward on thrust or fold nappes, lie west of the gravity high. Southward through the Reading Prong, the westward displacement of the Precambrian bodies from the gravity high increases to the extent that the southwestern end of the Prong lies in the middle of a gravity trough, the location of which appears to be independent of the configuration of the Precambrian rocks (Isachsen, 1964). Regional gravity studies by Woolard (1943) suggest that the ridges of Precambrian rock in the Reading Prong are allochthonous, and indicate a thickness of no more than 500 feet for Musconetcong Mountain, a major Precambrian ridge in New Jersey.

Several lines of evidence thus fortify the interpretation that the Precambrian rocks in the Nazareth quadrangle are displaced over the Paleozoic rocks. The amount of displacement appears to be at least three miles because small Precambrian outliers occur within

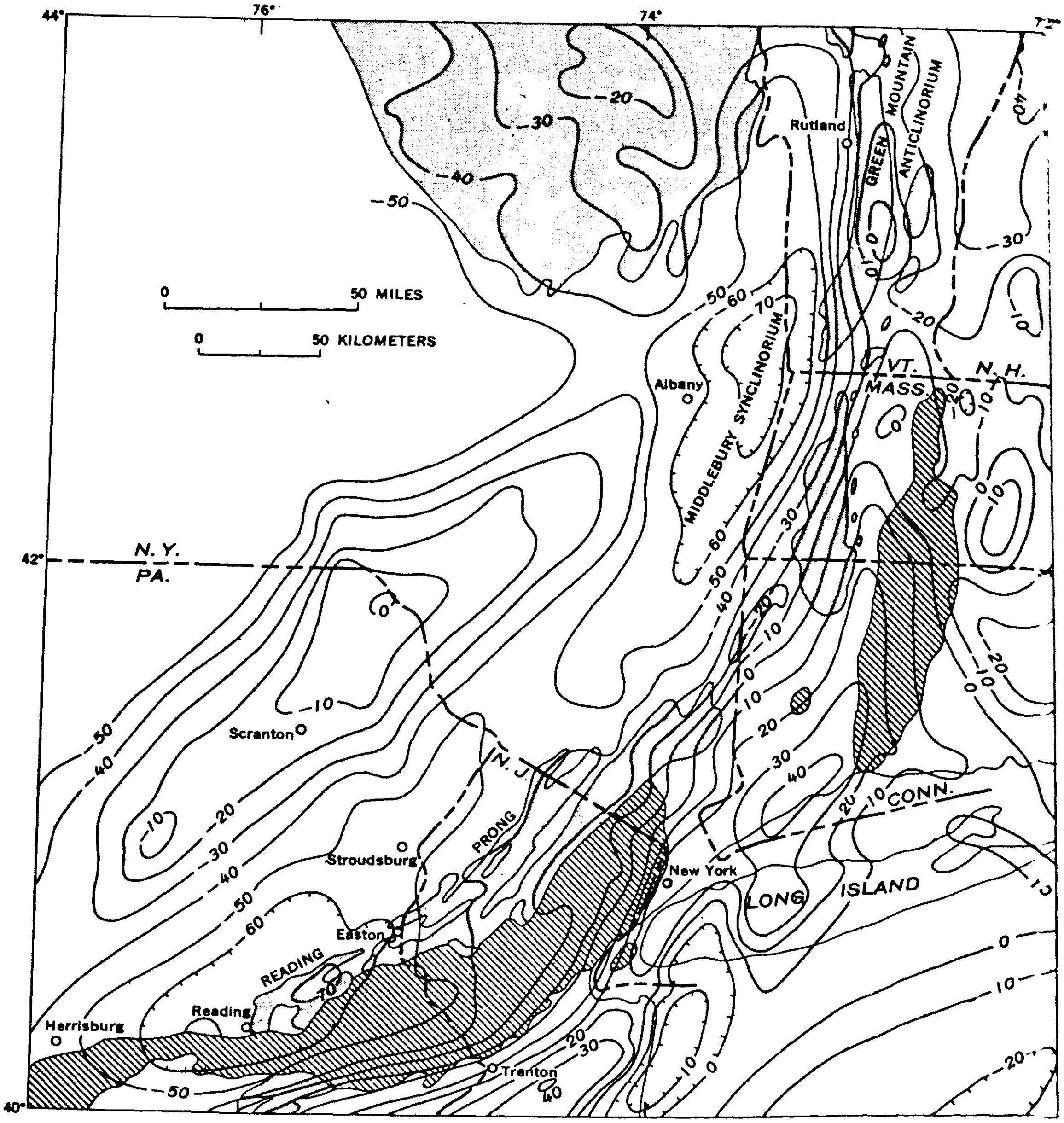


Figure 49. Map showing relation of Precambrian rocks to simple Bouguer gravity nomalies in the northeastern United States. Map modified from the Tectonic map of the United States (U.S. Geol. Survey and Am. Assoc. Petroleum Geologists, 1961, and Bouguer gravity anomaly map of the United States (Am. Geophys. Union, Spec. Comm. Geophys. and Geol. Study Continents, 1964). Precambrian rocks, stippled; Triassic rocks, hatchered; Paleozoic rocks and Coastal-Plain deposits, unpatterned. Contour interval 10 milligals. Figure from Drake, 1970.

the carbonate terrane of the Lehigh Valley about three miles northwest of the main body of Precambrian rock. This interpretation assumes that the smaller bodies were once part of the larger ones. Although the Precambrian-Paleozoic interface is mapped as a thrust fault throughout the area, there is surprisingly little stratigraphic displacement on the thrust (or thrusts), except in cases where Precambrian outliers occur well within the Great Valley. In the Nazareth quadrangle the maximum stratigraphic displacement on the thrust probably is no more than six or seven hundred feet. The Paleozoic stratigraphic section in front (northwest) of the main thrust is complete in terms of the number of units present. The Leithsville Formation is present throughout. It is about 1,000 feet thick; hence, the stratigraphic displacement must be somewhat less than 1,000 feet. Drake (1969) notes that nowhere in the mapped length of this thrust front, a distance of nearly 40 miles, is the stratigraphic displacement much more than 1,000 feet. This point is of great importance in the interpretation of the regional geologic picture to be considered in another section.

Regional Structural Relations

Historical Summary

The structural synthesis of the Reading Prong-Great Valley area has been hotly contested for generations. It has been noted previously that early workers (Rogers, 1858; Prime, *in* Lesley, 1883; Wherry, 1916) considered the ridges underlain by Precambrian gneiss to be the cores of anticline. Miller (1925) and Miller and others (1939, 1941) recognized that such anticlines commonly were overturned and that they may be locally thrust faulted, but strongly believed that the Precambrian was essentially autochthonous and deeply rooted. Intermontane valleys floored with Paleozoic carbonate rocks were considered synclinal or block-faulted structures, and normal faults are shown bounding the south sides of many of the Precambrian ridges.

Stose and Jonas (1935) and Stose and Stose (1940) presented a sweeping departure from the above widely-held views. They considered the Precambrian rocks and the associated Hardyston Quartzite as entirely allochthonous, and viewed them as remnants of a great overthrust sheet that moved over the Paleozoic carbonate on a flat thrust plane.

Under this interpretation the carbonate-floored intermontane valleys would represent windows in the thrust sheet, and small outliers of Precambrian rock in the Great Valley, such as Camelhump and Pine Top of the Nazareth quadrangle, would be klippen. B. L. Miller (1944) specifically refuted the flat-thrust hypothesis by presenting drill data and other evidence that showed beyond doubt that carbonate rocks do not underlie gneissic rocks at the shallow depths demanded by the Stose-Jonas interpretation. Miller's evidence against the flat-thrust hypothesis is decisive and overwhelming, and his interpretation of the structure of Reading Prong was unchallenged until modern regional studies by the United States Geological Survey and the Pennsylvania Geological Survey began in the late 1950's. These studies found that thrust faults and major overturning are very important factors in the distribution of the Precambrian rocks throughout the Reading Prong, and that some hills and ridges underlain by Precambrian rocks and the Hardyston Quartzite are indeed klippen (Gray and others, 1960; Socolow, 1961; Field Conference of Pennsylvania Geologists, 1961; Geyer and others, 1963). Geologic mapping near the Delaware River in Pennsylvania and New Jersey suggested to A. A. Drake, Jr., and J. B. Epstein (in United States Geological Survey, 1962) that the Precambrian ridges there may be detached parts of a single thrust sheet. The revival of the Stose and Jonas (1935) overthrust interpretation was complete when Isachsen (1964) suggested that the entire Reading Prong, including the associated lower Paleozoic rocks, and the Hudson Highlands is an allochthonous mass that moved northwestward a considerable distance as a klippe or detached nappe.

Current Interpretation

The structural and geophysical data from the Nazareth quadrangle and surrounding area, other regional studies by geologists of the United States Geological Survey and the Pennsylvania Geological Survey, and B. L. Miller's (1944) specific evidence against the flat-thrust hypothesis require a regional tectonic picture to account for the following aspects of the geologic structure.

1. Regional overturning of the lower Paleozoic rocks of the Great Valley.

2. Allochthonous, Precambrian bodies that overlie lower Paleozoic rocks, but which also dip southeastward beneath such Paleozoic cover.

The most plausible interpretative tectonic picture that is consistent with the geologic and geophysical data is a regional fold-nappe structure involving both Precambrian and lower Paleozoic rocks. A nappe is an extensive sheet of rock that has moved for a considerable distance (on the order of miles) over rock units beneath and in front of it. It may be the upper plate of a regional low-angle overthrust (thrust-nappe), or a regional recumbent fold (fold-nappe). The typical examples are in the Swiss Alps.

The recognition and documentation of the geometry and regional extent of recumbent folds in the Great Valley of Pennsylvania dates from the work of Carlyle Gray (1950, 1951, 1952, 1954; Gray and others, 1958) in Berks and Lebanon Counties. Gray (1959) formally adopted alpine "nappe" tectonics to characterize certain recumbent folds of regional significance in the Great Valley. It is interesting to note, however, that the similarities of the complex fold structures of the Great Valley to those of the Alps of western Europe was recognized by geologists whose work, nearly a century ago, constituted the Second Geological Survey of Pennsylvania (Lesley, 1883). This report on Lehigh and Northampton Counties is even illustrated with a diagram showing a large, complex recumbent anticline in the Italian Alps. Of course, the full appreciation of the regional significance of such structures in the Alps did not occur until after the work of the Second Survey was published.

Subsequent field studies by geologists of the Pennsylvania Geological Survey (Geyer and others, 1958, 1963; MacLachlan, 1964; Field Conference of Pennsylvania Geologists, 1966) led to the recognition of the Lebanon Valley nappe, which includes Precambrian rocks of the Reading Prong and all overlying Cambrian and Ordovician units (tabs. 4 and 12) up to and including part of the Martinsburg Formation. This nappe, which underlies the entire Great Valley (Lebanon Valley section) east of Harrisburg, Pennsylvania, represents the complexly folded inverted limb of an immense recumbent anticline (Field Conference of Pennsylvania Geologists, 1966).

The fold-nappe interpretation was extended to the Delaware and Lehigh Valley by geologists of the United States Geological Survey (A. A. Drake, Jr., in United States Geological Survey, 1966; J. M. Aaron, in United States Geological Survey, 1967; Davis and others, 1967; Drake, 1967a, 1967b, 1969, 1970; Drake and others, 1969). In that area, which includes the Nazareth quadrangle described herein, at least one grand recumbent fold, the Musconetcong nappe (Drake, 1969), underlies the Great Valley and includes both Precambrian rocks and the lower Paleozoic rocks up to and including part of the Martinsburg Formation. The Musconetcong nappe is illustrated in Figure 50. The fact that this nappe occupies the same general area and includes essentially the same rock units as the Lebanon Valley nappe probably is not fortuitous. It is likely that they are indeed one and the same structure, as noted by Drake (1969). The Lebanon Valley nappe represents the inverted limb of the structure, whereas the Musconetcong nappe, which is northeast

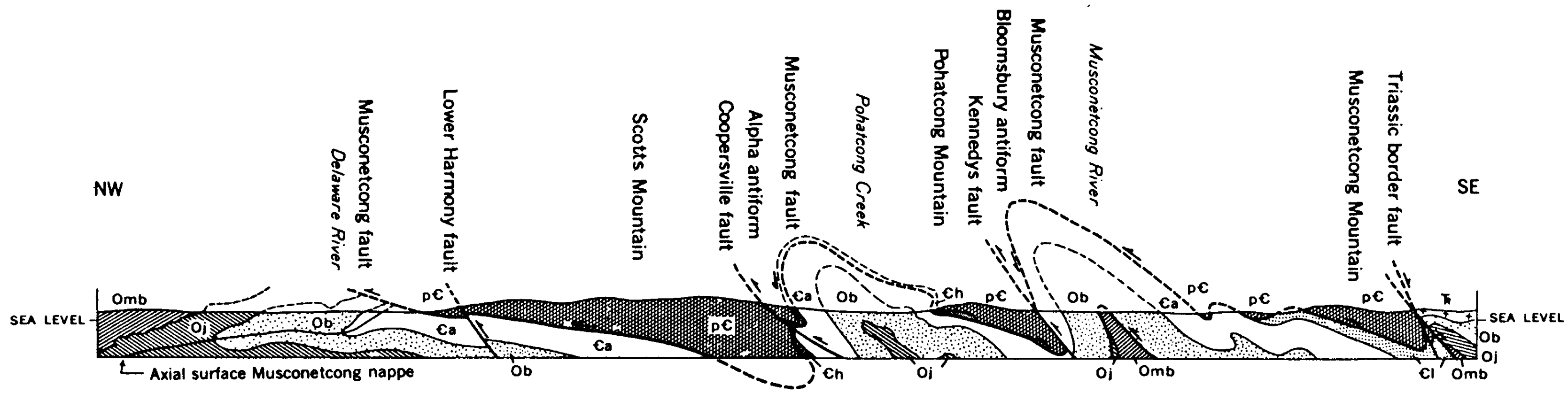


Figure 50. Geologic section across the Reading Prong in the Delaware Valley. Symbols: Tr, Triassic rocks; Omb, Martinsburg Formation; Oj, Jacksonburg Limestone; Ob, Beekmantown Group; Єa, Allentown Dolomite; Єl, Leithsville Formation; Єh, Hardyston Quartzite; pЄ, Precambrian rocks. Modified from Drake, 1970.

of the Lebanon Valley nappe on the regional plunge, exposes the inverted limb in eastern Pennsylvania and the upper (normal) limb in New Jersey, farther northeastward. The brow of the nappe is in the Martinsburg Formation (Davis and others, 1967; Drake and others, 1969), and the Precambrian rocks form the core. Under this interpretation the Precambrian core of this regional fold-nappe lies on top of the Paleozoic rocks of the lower (inverted) limb. The present outcrop patterns that result in linear fold belts in the Great Valley are interpreted as the result of refolding and faulting of this inverted limb, causing stratigraphically younger units to occur in the cores of antiforms, and older units in the troughs of synforms. The Precambrian rocks, likewise, occupy the cores of synformal structures. Such a geometric relationship is consistent with the aeromagnetic data, which indicates that the Precambrian bodies are thin, on the order of 500 feet at Musconetcong Mountain (Woolard, 1943), and that these bodies do not plunge under the Paleozoic as they should if the structure were autochthonous anticlinal.

The fold-nappe interpretation also is favored by the essentially normal stratigraphic relations between the Precambrian and lower Paleozoic rocks. As noted earlier, the contact front has been mapped over a length of nearly 40 miles, and nowhere is the stratigraphic displacement much more than 1,000 feet, and commonly it is less. It seems unlikely that such near-conformity would result if the Precambrian blocks were parts of imbricate thrust sheets.

The nappe is envisioned as underlying most of the Great Valley in eastern Pennsylvania and western New Jersey, from the

Triassic basin on the south to the Martinsburg terrane on the north. The total width is 17 to 20 miles. Clearly, if this interpretation is correct, the Musconetcong-Lebanon Valley nappe is a major structural feature of the middle Appalachians.

The location of the root zone, the area or zone from which the fold-nappe arose and where the axial plane presumably steepens, is not known. The lower Paleozoic rocks south of the Reading Prong Precambrian belt, between the Precambrian rocks and the Triassic basin, are inverted and recumbent, with antiformal and synformal structures similar to those that occur in the Great Valley north of the Precambrian (V. E. Gwinn, cited in Drake, 1970, p. 286). Such relations show that the root zone must lie farther southeast, probably beneath the Triassic cover. Aeromagnetic studies (Bromery and Griscom, 1967) reveal trends of Precambrian rocks beneath the Triassic cover, and Bromery (in United States Geological Survey, 1966) described steep east-trending gravity gradients that occur in the Triassic basin near Daylestown, Birdsboro, and Mount Hope, Pennsylvania. He interpreted these as reflections of profound faulting in the pre-Triassic rocks in the basin floor. Such profound faulting is postulated by Woodward (1964) on the basis of geologic evidence and analysis of the geometry of Appalachian fold patterns. The postulated fault, a right lateral wrench fault, is believed to have a displacement of between 75 and 100 miles. However, Burchfield and Livingston (1967) point out that such steep linear belts need not be wrench faults, and that some may indeed be root zones. They postulated that the Brevard zone of the southern Appalachian

Piedmont, which previously had been considered a major strike-slip fault on the basis of studies by Reed and Bryant (1964), actually has many characteristics of linear zones in the Swiss and Austrian Alps that are well-established as root zones of nappes. It is conceivable then that the major fault beneath the Triassic basin is both a wrench fault and a root zone, as was postulated by Reed and others (1970) for the Brevard zone. Moreover, it is possible, perhaps likely, that the sub-Triassic fault zone was modified by later tectonic events that may have contributed to the apparent strike-slip displacement and otherwise modified its possible root zone characteristics.

The above-cited recumbent relations of lower Paleozoic rocks on the south side of the Reading Prong Precambrian belt, between the Prong and the Triassic basin, in addition to showing that the root zone must be located farther southeast, provide another piece of evidence for the nappe interpretation. The fact that such regionally inverted structures occur both northwest and southeast of the Precambrian belt, in conjunction with the aeromagnetic data, that show that the Precambrian is not rooted, strongly indicates that the inverted limb of the nappe occurs both in front of and behind the Precambrian mass, and that the Precambrian mass thus is superjacent.

Chronology of Deformation

Structural relations in the Nazareth quadrangle and throughout the region, clearly indicate that the lower Paleozoic rocks have been deformed at least twice. The evidence for this includes a generation of largely recumbent folds that are related to flow cleavage (S_2), another generation of folds that are related to crenulation cleavage (S_3), and clear evidence in thin-sections, hand specimens, and outcrops that flow cleavage (S_2) is transected by crenulation cleavage (S_3), indicating that the crenulation cleavage must be younger. In addition, the nappe structure, the inverted lower limb of which underlies most of the Nazareth quadrangle and the Lehigh Valley, has been refolded into antiforms and synforms. A possible example of such a second generation fold in the upper limb of the nappe is the prominent syncline at the town of Nazareth. Faults cut all folded structures and hence are later, as well.

In the Nazareth quadrangle and throughout other mapped areas of the Lehigh and Delaware Valleys, the nappe structure involves only rocks up to and including the Martinsburg Formation (Middle and Upper Ordovician). The contact between the Martinsburg and the underlying Shawangunk Conglomerate (Silurian) is an angular unconformity. Thus it is reasonable to ascribe the nappe structure to the Taconic orogeny (Middle to Late Ordovician). The Lebanon Valley nappe, believed to be contiguous with the Musconetcong nappe of the Lehigh-Delaware Valleys, also is believed to have originated in the Taconic orogeny (Field Conference of Pennsylvania Geologists, 1966). It is thought by the writer and many others who have worked only in the Martinsburg and older Paleozoic rocks of the area that the flow cleavage (S_2) which is so pervasively developed in the argillaceous rocks and limestones also dates from the Taconic, and thus is related to nappe emplacement. Behre (1927) associated intense recumbent interpretation would demand that the nappe be emplaced during the Taconic orogeny without the development of a tectonitic fabric, or that the evidence for such a fabric has been destroyed by later deformation. It is difficult to conceive that such complex recumbent structures as are described from much of the Great Valley in Pennsylvania could have been emplaced without developing a tectonitic fabric of some kind. It is equally difficult to believe that if such rocks did develop such a fabric during nappe emplacement in the Taconic orogeny, that all traces of this fabric could have been obliterated by later (Acadian, Appalachian, or younger) deformation. In this regard, however, Epstein (1969) cites studies in progress (1968) by I. B. Alterman, Columbia University, in the Martinsburg outcrop belt from southeastern New York to eastern Pennsylvania. Alterman's work indicates that in several localities an earlier cleavage, previously unreported, which has been nearly obliterated by later deformation, can be recognized. Alterman believes that this less distinct, earlier cleavage developed as an axial plane cleavage in the nappe. Another possibility is that the cleavage in the Martinsburg and older rocks and that in the Silurian and younger rocks are of different tectonic generations, although statistically their attitudes are essentially similar. Such an occurrence would seem to require folding about a similar axis (or axes) in two or more orogenies (Taconic and Appalachian) widely separated in time. This does not seem unreasonable, however, in view of the probability that the earlier deformation imposes a structural grain (folds, faults, cleavage, etc.) that could strongly influence, perhaps folding and development of slaty cleavage with the Taconic orogeny, as did Drake and others (1960), and in both papers such features are taken to show that in this area the Taconic orogeny was considerably stronger than the Appalachian. Maxwell (1962) also believes that the cleavage is Taconic, but maintains that it is the result of high pore water pressures and very limited stress. He associates the cleavage with diagenesis rather than orogenesis.

Recent work by J. B. Epstein (1969) on the Martinsburg and younger rocks raises questions on the interpretation of the S_2 cleavage as a Taconic feature. Epstein presents several lines of evidence to show that the slaty cleavage in the Martinsburg and, by inference, the older rocks in which it occurs as well, developed late in the Paleozoic during the Appalachian orogeny. Among his evidence are the observations that the attitude of slaty cleavage in the Martinsburg is similar to that of slaty cleavage in several overlying units (above the angular unconformity), that a second cleavage (S_3 = slip = crenulation) cuts the slaty cleavage in the Martinsburg, and also appears in overlying formations, that in a few localities fold axes pass from the Shawangunk Conglomerate into the Martinsburg Formation without deflection, and that the cleavage in the Martinsburg is parallel to the axial planes of these folds or is slightly fanned.

The above lines of evidence, among others, indicate to Epstein that the flow cleavage in the Martinsburg and older units is post-Taconic in age, and that it must therefore be related to refolding of the nappe, rather than nappe emplacement. This control the course of later deformation within the same rock sequence. Thus, it is conceivable that a fabric developed during the Taconic orogeny could have been rejuvenated during the Appalachian orogeny and transferred its strain pattern to the younger rocks. Or, it is possible that the Appalachian orogeny was not influenced by the Taconic grain, but that Appalachian stress field was oriented similar to that of the Taconic, and the resultant fabrics thus are similar. Such a configuration of the Appalachian orogeny might rejuvenate or accentuate Taconic fabrics, but it is not necessary that these fabrics be transferred to the younger rocks.

The Taconic orogeny has not as yet been independently dated in the Reading Prong of Pennsylvania or New Jersey by radiometric means. In Duchess County, New York, the southern part of the Taconic region, the Hudson River Pelite was metamorphosed a minimum of 430 m.y. B.P., very close to the Ordovician-Silurian boundary, as determined by Long (1962) by Rb-Sr analysis of muscovites. This pelite is at least a partial equivalent of the Martinsburg.

The crenulation (slip) cleavage that occurs sparsely in pelitic rocks and in some limestones of the Nazareth quadrangle post-dates the flow cleavage (S_2). This means that the S_3 cleavage is no older than late Ordovician, but it may be as young as late Paleozoic, perhaps younger. Much of the problem of dating the S_2 cleavage also applies to the S_3, except that S_3 must be the younger.

In sum, there is direct evidence for only two stages of folding in the Lehigh-Delaware Valleys. The first stage was during the Taconic orogeny and involved the formation of a regional fold-nappe system, the inverted lower limb of which underlies most of the Nazareth quadrangle and the Lehigh Valley. The second stage was during the Appalachian orogeny and was characterized by refolding of the nappe. The question of cleavage development is an open one at this time. As far as the Martinsburg and older rocks in the Nazareth quadrangle are concerned, it is entirely reasonable to associate the S_2-flow cleavage with nappe development in the Taconic orogeny and the S_3-crenulation cleavage with refolding of the nappe at a later time, probably in the Appalachian orogeny. However the presence in other areas of similar cleavages with similar orientations in rocks that clearly post-date the nappe has led some to believe that all the cleavage in all of the Paleozoic rocks of eastern Pennsylvania dates from the Appalachian orogeny. Solutions to this conflict are largely matters of opinion. One possible solution is that the Appalachian folding was strongly influenced by the structural grain, including cleavage, established in Taconic deformation, thus fabric orientations typical of the Taconic may have been transmitted to younger rocks during the Appalachian orogeny. The flow cleavage, in this case, could therefore be both Taconic and Appalachian in origin. The crenulation cleavage, in any case, is a younger structure.

Geologists of the Pennsylvania Geological Survey working west of the Lehigh Valley (Field Conference of Pennsylvania Geologists, 1966) believe that the Lebanon Valley nappe was emplaced during the Taconic orogeny, mostly unaccompanied by development of a tectonitic fabric, and that major thrusting could have occurred in the Acadian

(Devonian) orogeny. There is no structural evidence to assign such an age to deformation in the Lehigh Valley, nor is there direct radiometric evidence that the Acadian orogeny affected these rocks. Numerous radiometric dates on micas from the Manhattan Prong, New York, indicate a major (and probably the last) thermal event and metamorphism occurred there 360 m.y. B.P. (Long and Kulp, 1962), that is, during the Devonian period. Several isotopic ages (Long and Kulp, 1962) suggest that the eastern side of the New Jersey Highlands was partially reheated and recrystallized during the metamorphism in the Manhattan Prong 360 m.y. ago. There is no evidence for this event farther southwest, in the Reading Prong, however.

Origin of the Musconetcong-Lebanon Valley Nappe

Structures of the sort believed to underlie the Nazareth quadrangle and much of the Great Valley in eastern and east-central Pennsylvania are of great interest and importance because of their bearing on the mechanisms of rock deformation and mountain-building. Of particular interest is the amount of crustal shortening implied by such structures, in relation to the kind and magnitude of the forces that acted to form the structures. Unfortunately, the geometry of the nappe structure in the Great Valley is poorly known, as is its configuration and extent at depth, and its relation to rocks southward beneath the Triassic basin. This does not allow much detailed analysis of its origin. Such analysis must await further geologic and geophysical studies. The discussion that follows is speculative, therefore, and is based upon what little is known about the nappe as it is interpreted herein.

It has been pointed out previously that, beginning with the Jacksonburg cement rock facies, fine epiclastic sediment was shed into the depositional area represented by the eastern Pennsylvania lower Paleozoic rocks, and that this sedimentation pattern continued and became more pronounced with the deposition of the Martinsburg sediments. Further, it was noted that these sediments, the Martinsburg in particular, probably are derived from a rising eastward or southeastward sourceland that persisted throughout the remainder of the Paleozoic. This rising sourceland also provided the submarine slopes down which flowed Martinsburg turbidity currents, and the currents themselves could have been generated in part by tectonic activity (earthquakes and the like) related to the evolution of the sourceland. Thus the Martinsburg is a synorogenic deposit similar to flysch in the European Alps.

Doubtless this landmass, a tectonic welt, is one manifestation of the Taconic orogeny in the middle Appalachians. Its influence on sedimentation seems to be clearly recorded. It may, however, have had a great influence on the development of geologic structures as well. Consider the following points:

1. The Great Valley nappe system involves only Martinsburg and older rocks.
2. The Martinsburg (Middle and Upper Ordovician) is unconformably overlain by the Lower and Middle Silurian

Shawangunk Conglomerate. Thus it seems unlikely that the Martinsburg was buried deeply by superjacent sediments at the time of nappe formation, and in fact, it is conceivable that the upper Martinsburg itself may post-date the nappe.

3. The nappe involves both Precambrian and superficial rocks.
4. If the present interpretation is correct, the inverted limb of the structure is exposed over a wide area of the Great Valley in central and eastern Pennsylvania.
5. The magnitude and complexity of the nappe implies a considerable amount of crustal shortening.

These points suggest to the writer that gravity sliding on the flank of an eastward or southeastward tectonic welt may have played an important part in the development of the nappe structure. The considerable lateral displacement involved, the probable relatively plastic condition of much of the sedimentary pile, especially the Martinsburg and upper Jacksonburg, and the preserved inverted limb of the nappe, suggest that lateral compression acting alone was not the immediate driving force of the structure. It is difficult to imagine such stress being transmitted through such plastic rocks as the Martinsburg and upper Jacksonburg, especially since these rocks were not deeply buried at the time. On the other hand, some lateral compression would seem to be required to involve basement rocks in the core of the nappe structure. The nappe could have been initiated by structural events such as folding or thrusting under lateral compression on the flank of the growing southeastward tectonic welt. The structure continued to grow until it collapsed under its own weight, at which time it continued to move down slope under the influence of gravity as an essentially superficial feature. A slope suitable for gravity gliding is implicit in the growth of a major landmass to the east, and deSitter (1964) cites studies that show that such slopes may be very gentle indeed, perhaps as little as 3°. Further, deSitter (1964) points out that marginal troughs of orogenic belts seem to be the most favorable recipients of gliding nappes, and that an inverted position of a large mass, in particular when it is not laminated or reduced in thickness, is strong evidence of the gliding nature of the transport mechanism. The regional geologic relations of the Musconetcong-Lebanon Valley nappe system, as it is presently understood, are consistent with these two points. Moreover, the Martinsburg argillaceous sediments, which formed the outer envelope of the nappe, should have been a particularly suitable shearing horizon on which such down slope movements could occur. Probably other relatively incompetent rocks or horizons contributed in this respect also. With this background in mind, the Stockertown fault may be one of the fundamental shear horizons, perhaps the most important single horizon along which nappe emplacement occurred. The displacement on this fault, as has been mentioned previously, is believed to be on the order of at least several miles. In the Bloomsbury quadrangle Drake (1967b) has mapped a body of Martinsburg in contact with the Epler Formation in the core of an antiform. The contact between these units is a folded thrust (Fish Hatchery

Fault). This body is nine miles southeast of the main Martinsburg belt and is only slightly more than a mile north of Musconetcong Mountain, a major Precambrian body that is strongly believed to be rootless and quite thin, probably no more than 500 feet thick (Woolard, 1943). Drake (1969) believes that the Fish Hatchery fault may be co-extensive with the Stockertown Fault. If this interpretation is correct, the Stockertown-Fish Hatchery Fault system very probably is the principal shear horizon along which nappe emplacement occurred.

Some shearing of the basement rocks through the sedimentary covers is evident from the fact that Pine Top and Camelhump, the two klippen of Precambrian rocks that occur within the Lehigh Valley, are displaced three miles northward, ahead of the main Precambrian belt, and are in contact with Beekmantown rocks. Rocks in both klippen are extensively sheared. Also Precambrian rocks immediately adjacent to the contact with Paleozoic rocks typically show some degree of shearing and mylonitization, as do adjacent Paleozoic rocks.

Chapter V

SUMMARY AND CONCLUSIONS

Stratigraphic Geology

The Nazareth quadrangle comprises parts of the Reading Prong and Great Valley sections of the middle Appalachians. The Reading Prong is underlain by a variety of Precambrian rocks; the Great Valley is underlain mostly by carbonate rocks. This study focuses mainly on two aspects of the geology of the Nazareth quadrangle and surrounding area: 1) the origin, depositional history, and stratigraphic significance of the Cambrian and Ordovician sedimentary rocks that underlie the Lehigh Valley section of the Great Valley, and 2) the geologic structure of these rocks and their structural relationship to the Precambrian rocks of the Reading Prong.

The Precambrian rocks can be divided into two stratigraphic sequences according to metamorphic grade and apparent age. The older sequence is composed of metasedimentary and meta-igneous gneissic rocks that belong to the granulite facies of regional metamorphism. Folds in the foliation of these rocks are mostly asymmetric or overturned to the northwest, and most plunge gently northeastward. The younger sequence is composed of ferruginous quartzite, arkose, conglomerate, altered dolomitic marble and, outside the Nazareth quadrangle, purple slate and metarhyolite, all of which are in the greenschist facies of regional metamorphism.

Structural relations within the Precambrian rocks suggest that only one episode of plastic deformation affected these rocks. Radiometric evidence, mostly from the New York City area, indicates two periods of metamorphism, at about 1150 m.y. B.P. and about 835 m.y. B.P. (Long, 1962; Long and others, 1959; Long and Kulp, 1962). These dates are based on K-Ar and Rb-Sr determinations on biotites from gneissic rocks at several localities in the Jersey Highlands, and on concordant U-Pb apparent ages of zircon (for the 1150 m.y. event) in the Canada Hill Gneiss at Bear Mountain, New York.

The Precambrian rocks are overlain unconformably by the Hardyston Quartzite (Lower Cambrian), a 100-foot thick unit of arkosic and quartzose rocks that form the base of the Paleozoic sequence in eastern Pennsylvania. Textures and mineral composition of the Hardyston suggest that it is mostly locally derived from a high grade metamorphic and igneous terrane that was essentially similar to the Reading Prong as it stands today, and that it probably is of both marine and alluvial origin. The lower arkosic, poorly sorted rocks are believed to be alluvial deposits that developed as a sedimentary veneer on the flank of a shield of Precambrian rocks. These deposits were inundated by a transgressing sea moving westward onto the North American Shield. The advancing sea reworked the upper part of the alluvial deposits and added deposits that were more quartzose, cleaner, finer, and better sorted.

The sedimentation pattern established with the deposition of the Hardyston sediments continued throughout the Cambrian and much of the Ordovician, except that carbonates predominate until the Middle Ordovician. By this is meant that the shield area to the west and northwest remained the principal source of clastic terrigenous sediment, and open sea lay to the east. Environments and subenvironments within this general paleogeographic picture were the locus of Cambrian and Ordovician carbonate sedimentation.

The Hardyston is overlain, in succession, by the Leithsville Formation (Middle(?) Cambrian), the Allentown Dolomite (Upper Cambrian), the Rickenbach Dolomite (Lower Ordovician), and the Epler Formation (Lower and Middle Ordovician), all largely dolomite units. All contacts are transitional. The Leithsville is composed of dolomite and quartz sand and silt. Along with the Hardyston, it records the submergence of the North American Precambrian shield and represents the transition from dominantly epiclastic sedimentation to carbonate sedimentation. The Allentown Dolomite is an assemblage of aphanocrystalline to coarsely crystalline dolomites that are distinctively cyclic, and which contain a variety of sedimentary and organic structures. The ideal Allentown cycle consists of seven dolomite lithologies that can be related individually to the subtidal, intertidal, and supratidal environments, on the basis of the textures and structures they contain. The ideal cycle begins at the base with a flat-pebble conglomerate that marks the destruction of the top of the preceding cycle. The conglomerate is interpreted as a subtidal or intertidal deposit. Succeeding lithologies are, in order, dololutite (subtidal), dolarenite (intertidal), oölitic dolarenite (intertidal), dolorudite (intertidal), algal stromatolite (intertidal), and dolomicrite (supratidal). The Allentown Dolomite thus is the result of carbonate sedimentation within a sea-marginal tidal complex.

The Rickenbach Dolomite and the Epler Formation compose the Beekmantown Group in easternmost Pennsylvania. The Rickenbach is characterized by thickly bedded coarsely crystalline dolomites that locally are very ruditic or "pseudoruditic"; the Epler contains fine-grained dolomites and numerous interbedded limestones. The Epler represents the transition from the largely dolomitic rocks of the Leithsville-Allentown-Rickenbach depositional epoch to the calcareous rocks of the Jacksonburg Limestone. The Beekmantown rocks contain few sedimentary structures that are indicative of a particular depositional environment. Moreover the lack of such features probably rules out a near-shore tidal complex as a possible depositional environment. The petrographic character of these rocks suggests that the Beekmantown may have originated in relatively quiet marine waters, perhaps behind a large barrier reef or other topographic obstruction.

Most of the dolomite in the sequence of rocks described above is of replacement origin. This is amply demonstrated by the fact that numerous textures and structures that are typical of calcium carbonate sediments now exist only as palimpsest relics (ghosts) in the dolomite mosaic. These features include fossils, oöids, pellets, algal stromatolites, in addition to dolomite crystals that transect allochem boundaries. Most of these dolomites that are clearly of replacement origin are medium to coarsely crystalline. In the partially dolomitized limestones of the Epler Formation, it is apparent that dolomitization is accompanied by increased crystal size. Such replacement dolomite is the result of diagenesis, and is "secondary" in the petrographic sense. There are dolomites in these rocks, however, that contain none of these distinctive calcium carbonate textures or structures, are very finely crystalline to aphanocrystalline (micritic), and are either structureless or are finely laminated. Dolomite in such rocks is believed to be the result of either direct inorganic precipitation of dolomite, or syngenetic replacement of calcite or aragonite before a calcium carbonate fabric developed, and probably before or immediately after the calcium carbonate crystal reached the sea floor. Additional evidence for this mode of origin is found in certain peculiar (eccentric) oöids that occur very sparsely in the Allentown. These structures, unlike all other oöids observed in the Allentown show no evidence of replacement except in the cores. Otherwise, they appear to be accretions of concentric shells of dolomite (presumably primary) and another mineral that now is missing, but which presumably was gypsum or anhydrite (also a primary precipitate). Such primary or penecontemporaneously replaced dolomite is termed syngenetic and at present no distinction can be made between the two modes of origin in rocks such as the lower Paleozoic of the Lehigh Valley.

The syngenetic dolomites of the Lehigh Valley are believed to be analogous to Recent syngenetic dolomote that is forming in such places as the Bahamas (Shinn and others, 1965), the Persian Gulf (Illing and others, 1965), and in several localities in south Florida (Shinn and Ginsburg, 1964). In each of these areas micritic dolomite with features similar to those described in the Leithsville, Allentown, Rickenbach, and Epler is associated with evaporitic processes in hypersaline marine waters. It is likely that the replacement dolomites result from contact with such hypersaline brines. Such brines could form by evaporation in the supratidal environment, or by evaporation of protected shallow waters with restricted circulation and little influx of water of normal salinity, such as shoreward of a topographic barrier of some kind. Such brines are denser than water of normal salinity and could percolate downward through intergranular openings to replace earlier deposited calcium carbonate sediments, and could spread laterally along bedding planes and impervious layers to dolomitize calcareous sediments well outside the area of brine formation. As salinity increases beyond the point at which dolomite begins to precipitate more soluble members of the evaporite sequence, such as anhydrite, gypsum, and halite, may form. That such hypersaline conditions could have prevailed in the environments represented by the Cambrian and Ordovician dolomites of the Lehigh Valley is suggested by B. L. Miller's (1937) report of casts of halite crystals in the Beekmantown of Lehigh County.

The dolomitic sequence described above is overlain by the Jacksonburg Limestone (Middle Ordovician) which supports the principle mineral industry of the Lehigh Valley--the manufacture of Portland cement. The Jacksonburg consists of two superposed units, the cement limestone facies below, and the cement rock facies above. The cement limestone facies is composed of thickly bedded, coarsely crystalline bioclastic limestones; the cement rock facies is fine grained argillaceous limestone. The whole unit is about 1000 feet thick. The Jacksonburg Limestone probably originated in a shallow marine to marginal marine environment. The bulk of the fossil material (bryozoans, crinoids, brachiopods, corals) are bottom-dwellers that typically inhabit warm, clear, shallow water that is protected from strong waves or currents. However, most of this material is fragmental and well sorted, and suggests that considerable turbulence and reworking must have affected these sediments as well, thus the immediate environment of deposition probably was a shallow bank or marine shelf on which transported bioclastic debris was reworked. The cement rock facies appears to have been deposited seaward (southeast) of the bank that was the locus of cement limestone deposition. The argillaceous material, which is much more abundant toward the top of the unit and becomes the principal constituent of the Martinsburg Formation, heralds the onset of the Taconic orogeny.

The Martinsburg Formation (Middle and Upper Ordovician) is represented in the Nazareth quadrangle only by the Bushkill Member, mostly a dark gray, thin bedded claystone slate. The contact with the subjacent Jacksonburg Limestone is transitional. Primary structures in the Martinsburg (McBride, 1962) indicate that paleocurrents originated on the shelf edge of a landmass located southeast of the present outcrop belt. The source area of the sediments appears to have been a terrane of varied lithology, including carbonates, low-grade metamorphic rocks and acid plutonic rocks. The Martinsburg represents the beginning of Appalachian Paleozoic sedimentation from a southeastward source. Several studies (Pelletier, 1958; Yeakel, 1962; Meckel, 1967) have shown that this eastern source area persisted throughout the remainder of the Paleozoic. The growth and rise of this landmass appears to be one manifestation of the Taconic orogeny. Structural relations in the Great Valley suggest that this landmass probably played an important part in the development of the geologic structures there as well.

The Paleozoic rocks of the Nazareth quadrangle are overlain by numerous patches of glacial drift. These are poorly consolidated, poorly sorted, and lack stratification. No distinctive constructional topographic forms that are typically associated with glacial deposits have been observed in this drift. These glacial materials are pre-Wisconsin in age. Traditionally such older glacial deposits in the area have been assigned to the Illinoian advance, but correlation with the type Illinoian of the Mississippi Valley is very uncertain, and use of that name for the eastern Pennsylvania deposits does not seem justified by the data available at this time.

Structural Geology

The lower Paleozoic rocks of the Nazareth quadrangle are complexly folded into a variety of fold types that range in magnitude from microscopic crinkles to major folds up to 1,000 feet or more in wave-length. Geometrically, most folds observed are hybrid types that are neither purely concentric, purely similar, nor purely disharmonic. Similarly, the mechanical processes by which these folds formed probably were combinations of bedding slip and flow. The most important aspect of folding the Nazareth quadrangle and vicinity is the role that lithology apparently has played in determining fold geometry and, by inference, in governing the balance of processes that formed the folds. Concentric folds occur in the thick, very competent dolomitic rocks and in the Jacksonburg cement limestone facies. Similar folds, the most common type in the Lehigh Valley, occur in thin to medium bedded dolomitic rocks and in some interbedded limestones and dolomites of the Epler Formation. Wildly disharmonic folds occur in rock sequences that contain strong contrasts in relative rock competence. Contorted flow has resulted where more ductile rocks have been forced to accommodate to the movements of the relatively more competent layers. Most folding of this type occurs in silty banded Epler limestones and thin interbedded Epler limestones and dolomites. Most observable folds in the area are overturned to recumbent, and many are isoclinal. Overturning is toward the northwest.

Fold axes observed in the lower Paleozoic rocks mostly plunge very gently northeastward at angles that range from horizontal to 10°. The most common axial orientation is N 60° to 65° E. Less prominent are folds that plunge about 10°, N 82° E.

Faults are common features of most of the larger outcrops in the area, but few are mappable, except in areas of good exposure. The Stockertown Fault is the most important fault in the area. It is a major thrust fault that appears to have a displacement of at least several miles (Drake, 1970).

Cleavage is present throughout the quadrangle in all stratigraphic units and in all lithologies. It varies in style and type according to the lithology in which it occurs. There are two cleavages present within the area, but they occur together only in the less competent rocks. The first generation cleavage, designated as S_2, is a pervasive flow cleavage in the Epler limestones, the Jacksonburg cement rock facies and the Bushkill Member of the Martinsburg Formation. Flow cleavage is parallel to the axial planes of observable folds and generally is quite flat, reflecting the regional recumbency of folds in the region. Many folds are isoclinal as well, hence bedding and flow cleavage commonly are essentially parallel. The regional strike of flow cleavage (S_2) is about N 60° E, and dips range from horizontal to about 30° southeast. In the more competent dolomitic rocks and the Jacksonburg cement limestone facies, the S_2 cleavage is a fracture cleavage that is approximately parallel to axial planes of observed folds, but tends to be somewhat steeper and more fanned.

The second generation cleavage is a crenulation cleavage that deforms planes of flow cleavage by causing a wrinkling, crenulation, or folding of the plane. This cleavage occurs only in the less competent rocks. Crenulation cleavage is oriented essentially parallel in strike to flow cleavage but is slightly steeper in dip. Axes of these crinkles and folds plunge gently northeastward or southwestward, and are generally parallel to folds in bedding. It seems clear that crenulation cleavage and cleavage folding are genetically related.

The most important lineation in the Nazareth quadrangle Paleozoic rocks is the intersection of bedding and S_2-cleavage. This lineation is parallel to axes of folds in bedding and typically plunges gently northeastward. A second generation lineation is formed by the intersection of flow cleavage (S_2) and crenulation cleavage (S_3), but it occurs only in the less competent rocks. Sparse data indicate that this lineation plunges gently northeastward or southwestward, parallel in strike with the L_1-lineation.

The Precambrian bodies mapped in the Nazareth quadrangle are considered to be in thrust contact with the lower Paleozoic rocks, and they are not rooted. This interpretation is supported by exploratory drilling on Rattlesnake Hill, a Precambrian ridge south of Riegelsville, Pennsylvania, by several water-wells drilled through Precambrian rocks into Paleozoic carbonate rocks, and by aeromagnetic studies. The aeromagnetic studies, in particular, show that the major Precambrian bodies are bounded on the north by strong negative magnetic anomalies, indicating that Precambrian rocks do not occur at depth beneath the Paleozoic cover. On the south sides of these bodies the aeromagnetic contours grade steeply into areas underlain by Paleozoic rocks and ultimately drop off into the negative anomaly basin that bounds the next Precambrian ridge to the south. Such relations suggest that the Precambrian bodies dip beneath the Paleozoic rocks that bound them on the south, but that they terminate at depth. Woolard (1943) calculated that the Precambrian rocks at Musconetcong mountain are only about 500 feet thick. Such relations as have been described in Nazareth quadrangle occur throughout the Reading Prong in eastern Pennsylvania and western New Jersey. The interpretation of the Reading Prong Precambrian rocks as allochthonous is also consistent with regional gravity data.

The amount of displacement of Precambrian rocks over the lower Paleozoic rocks is at least three miles. Several small outliers of Precambrian rock occur within the carbonate terrane of the Lehigh and Delaware Valleys, and are about three miles northwest of major Precambrian bodies. There is, however, very limited stratigraphic displacement associated with the major Precambrian masses. The Paleozoic stratigraphic sequence in front (northwest) of the Precambrian bodies commonly is complete, or nearly so. In Nazareth quadrangle, the displacement probably is no more than 600 to 700 feet. Drake (1969, 1970) notes that nowhere in the mapped length of the thrust front that bounds the Lehigh-Delaware Valley on the southeast is the displacement much more than 1,000 feet.

The Precambrian rocks of the Reading Prong traditionally have been considered as the core of a regional anticlinorium. Stose and Jonas (1935) challenged this interpretation by asserting that the Precambrian bodies and the associated Hardyston Quartzite are entirely allochthonous, and are remnants of a great thrust sheet that moved over the Paleozoic rocks on a flat thrust plane. B. L. Miller (1944) refuted this interpretation with drill data that showed that the thrust contact could not be flat. Numerous modern studies in eastern Pennsylvania (Bromery and others, 1959; Gray and others, 1960; Socolow, 1961; Field Conference of Pennsylvania Geologists, 1961; Geyer and others, 1963; Drake and Epstein, _in_ United States Geological Survey, 1962) have shown that the Precambrian very probably is allochthonous, however.

The regional tectonic picture currently favored by most workers in the area involves a regional fold-nappe structure that is cored by Precambrian rocks (Geyer and others, 1958, 1963; MacLachlan, 1964; Field Conference of Pennsylvania Geologists, 1966; A. A. Drake, Jr., _in_ United States Geological Survey, 1966; J. M. Aaron, _in_ United States Geological Survey, 1967; Davis and others, 1967; Drake, 1967a, 1967b, 1969, 1970; Drake and others, 1969). The lower inverted limb of such a structure is believed to underlie much of the Lebanon-Lehigh-Delaware Valley area of Pennsylvania. The nappe plunges northeastward, thus the upper limb is exposed in easternmost Pennsylvania and western New Jersey. Precambrian rocks rest on the inverted limb. Present outcrop patterns are the result of refolding of the nappe. The location of the root zone, the zone from which the nappe arose and where the axial plane should steepen, is not known, but it probably is buried beneath Triassic rocks south of the Reading Prong. A zone of profound wrench faulting in Precambrian rocks beneath the Triassic basin is indicated by aeromagnetic studies (Bromery, in United States Geological Survey, 1966) and is suggested by geologic evidence as well (Woodward, 1964). This zone may be related to the root zone of the nappe system.

The nappe must have formed in the Late Ordovician Taconic orogeny because younger rocks are not involved. It was later refolded, probably in the late Paleozoic Appalachian orogeny. Two generations of deformation are also indicated by minor structures. A flow cleavage related to a generation of largely recumbent folds is itself folded and is transected by a crenulation cleavage. There is some question whether the flow cleavage is related to the generation of nappe emplacement, hence a Taconic feature, or to refolding of the nappe, hence a post-Taconic, probably Appalachian feature. In the Late Ordovician and older rocks it is entirely reasonable to relate flow cleavage to nappe development and crenulation cleavage to refolding. However, similar cleavages with similar orientations occur in younger, post-Taconic, rocks as well, and this could suggest that all the cleavage is post-Taconic, and that the nappe developed without formation of cleavage.

Presently, there are no real solutions to this dilemma, and, clearly, further work is needed. If all the observed cleavage is post-Taconic, it is possible, nevertheless, that a cleavage was associated with nappe development but evidence for it was destroyed by later deformation. It is also possible that such evidence is present but has not yet been recognized as such. On the other hand it is possible that the flow cleavage in Late Ordovician and older rocks is Taconic, but was rejuvenated in the later deformation and transferred to younger rocks. Or, perhaps the cleavage in the older rocks is Taconic and that in the younger rocks is Appalachian, and that the two orogenies were simply homoaxial.

The Musconetcong-Lebanon Valley nappe system that underlies the Great Valley in eastern Pennsylvania is believed to have originated as a thrust or a fold that moved Precambrian rocks into superficial rocks on the flank of a landmass rising to the southeast, presumably a manifestation of the Taconic orogeny. The structure grew under directed stress until it collapsed under its own weight, at which time it continued to move northwest down the regional slope under the influence of gravity. The Martinsburg argillaceous sediments that formed the outer shell of the nappe should have been a very suitable shear horizon on which such movement could have occurred, but other incompetent rocks could have been involved also. The Stockertown fault, on which displacement is believed to be on the order of at least several miles (Drake, 1970) appears to be a major shear horizon along which nappe emplacement occurred. Shearing of Precambrian rocks through the sedimentary cover is evident from the fact that Pine Top and Camelhump, klippen of Precambrian rocks that occur within the Lehigh Valley, are displaced three miles northwestward, ahead of the main Precambrian belt, and are in contact with Beekmantown rocks.

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