Paleotectonic Investigations of the Pennsylvanian System in the United States

Part I. Introduction and Regional Analyses of the Pennsylvanian System
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES
GENERALIZED RECONSTRUCTION OF LANDSCAPE IN THE EASTERN UNITED STATES DURING MIDDLE PENNSYLVANIAN TIME

Prepared with the assistance of Edwin H. Colbert, Robert M. Kosanke, G. Edward Lewis, and James M. Schopf.
Paleotectonic Investigations of the Pennsylvanian System in the United States

Part I. Introduction and Regional Analyses of the Pennsylvanian System

By EDWIN D. McKee and ELEANOR J. CROSBY, Coordinators

GEORGE O. BACHMAN  WILLIAM W. MALLORY
KENNETH G. BELL  WILLIAM J. MAPEL
ELEANOR J. CROSBY  EDWIN K. MAUGHAN
GEORGE H. DIXON  GEORGE E. PRICHARD
SHERWOOD E. FREZON  GERALD L. SHIDELER
ERNEST E. GLICK  GARY F. STEWART
WILLIAM P. IRWIN  HAROLD R. WANLESS
EDWIN D. McKee  RICHARD F. WILSON

Coordinating assistant MARGUERITE W. GLENN

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853

Chapters A through R

UNITED STATES GOVERNMENT PRINTING OFFICE, WASHINGTON: 1975
CONTENTS

[Letters in parentheses designate individual chapters]

(A) Introduction ......................................................... Page 1
(B) New England, by Harold R. Wanless and Kenneth G. Bell ............................................. 9
(C) Appalachian region, by Harold R. Wanless ................................................................. 17
(D) Michigan basin region, by Harold R. Wanless and Gerald L. Shideler ......................... 63
(E) Illinois basin region, by Harold R. Wanless ................................................................. 71
(F) Missouri and Iowa, by Harold R. Wanless ................................................................. 97
(G) Nebraska and adjoining parts of South Dakota and Wyoming, by George E. Prichard .... 115
(H) Kansas, by Gary F. Stewart .......................................................... 127
(I) Arkansas and northern Louisiana, by Ernest E. Glick ................................................. 157
(J) Texas Panhandle and Oklahoma, by Sherwood E. Frezon and George H. Dixon .......... 177
(K) Central and west Texas, by Eleanor J. Crosby and William J. Mapel ...................... 197
(L) New Mexico, by George O. Bachman ............................................................................. 233
(M) Eastern Colorado, by Richard F. Wilson ................................................................. 245
(N) Middle and southern Rocky Mountains, northern Colorado Plateau, and eastern Great Basin region, by William W. Mallory 265.
(O) Montana, North Dakota, northeastern Wyoming, and northern South Dakota, by Edwin K. Maughan ........................................... 279
(P) Arizona, by Edwin D. McKee ...................................................................................... 295
(Q) Nevada and southern California, by Richard F. Wilson .......................................... 311
(R) Pacific Northwest region, by William P. Irwin .......................................................... 329

Index ................................................................................................................................. 333

ILLUSTRATIONS

[All plates are in separate case; plates 15–17 listed in part II. Text figures and tables listed in individual chapters]

FRONTISPICE. Generalized reconstruction of landscape in the Eastern United States during Middle Pennsylvanian time.

PLATE 1. Map showing location of control points for Pennsylvanian System.
2. Geologic map of units directly beneath Pennsylvanian System.
3-9. Maps showing thickness and lithofacies of Pennsylvanian System:
   3. Interval A (3A, isopach; 3B, lithofacies).
   5. Subinterval A: (5A, isopach; 5B, lithofacies).
   8. Interval D (8A, isopach; 8B, lithofacies).
10. Cross sections through rocks of Pennsylvanian age.
11. Map showing total thickness of Pennsylvanian rocks.
12. Map showing geologic units directly above Pennsylvanian System.
13. Generalized chart showing stratigraphic units in major Pennsylvanian divisions.
14. Maps showing paleogeography of Montana and North Dakota during approximately middle Early Pennsylvanian (middle Morrow) time.
Introduction

By EDWIN D. McKEE

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-A
CONTENTS

Division of the system .............................................................. 1
  Units of division ................................................................. 1
  Basis for correlation .......................................................... 1
  Chart of stratigraphic units ................................................. 2
Construction of paleotectonic maps ......................................... 2
  Sources of data ........................................................................ 2
  Structural terms ...................................................................... 3
  Map of control points ............................................................ 3
  Index to localities and sources ............................................... 3
Types of maps and sections ......................................................... 3
  Isopach maps ......................................................................... 4
  Lithofacies maps .................................................................... 4
  Paleogeologic map and map of overlying units ......................... 4
  Cross sections ......................................................................... 5
  Interpretive maps: restored isopach and paleotectonic ............... 5
Acknowledgments ........................................................................ 5
References .................................................................................. 7

ILLUSTRATIONS

[For listing of plates (in separate case) see volume "Contents"]

Figure 1. Map showing areas of responsibility ............................... 2
  2. Diagram showing terms for structural features as used in this paper 3

TABLE

Table 1. Interval assignments for Pennsylvanian ............................ 2
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES,
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

INTRODUCTION

By Edwin D. McKee

The Pennsylvanian is the fourth geologic system to be analyzed and synthesized by geologists of the U.S. Geological Survey in the form of a paleotectonic study covering the conterminous United States. Earlier investigations were of the Jurassic, Triassic, and Permian Systems. Results were published as Miscellaneous Geologic Investigations Maps I-175, I-300, and I-450 and in Professional Paper 515.

The objective of these investigations is to provide in graphic form the factual basis for recognition of tectonic events of each system on a countrywide scale. The maps in this publication depict rock thickness, generalized lithology, ancient geography, and other regional relations of the Pennsylvanian System. Methods of preparation of the maps, the stratigraphic limits of the map units, and various stratigraphic and structural features and their probable tectonic significance are discussed.

Pennsylvanian data were largely compiled between 1961 and 1968 by 16 geologists, including the late Harold R. Wanless,1 who covered the five eastern regions and contributed to several of the special studies. The areas of responsibility of the cooperating geologists are indicated in figure 1. Work in Kansas was done by Gary F. Stewart, of the Kansas Geological Survey.2

Results of this investigation are presented in three units. Part I comprises an introduction and 17 chapters, each describing and discussing one of the regions in which the conterminous United States was divided for purposes of study and mapping. Part II is a synthesis of Pennsylvanian history to accompany interpretive maps of the five divisions of the Pennsylvanian System treated in this publication; it also includes a series of chapters on depositional environments, climatic conditions, and economic products of the system. The final section of part II is devoted to an index of localities and sources used in construction of the principal maps of this publication. Part III consists of the plates on which are presented the major maps and sections.

DIVISION OF THE SYSTEM

Preparation of meaningful lithofacies and thickness maps requires division of the Pennsylvanian System. A lithofacies map for the entire system, undivided, would be largely unintelligible; this system, like others, includes thick sequences of rocks of diverse origin as a result of multiple, sometimes unrelated geologic events.

UNITS OF DIVISION

The system is divided in this paper into five main parts designated, in ascending order, intervals A, B, C, D, and E (pl. 13). Interval A in the eastern part of the country is divided into two parts referred to as subintervals A1 and A2. For each interval and, in some regions, for subintervals, isopach and lithofacies maps (pls. 3-9) show the location and extent of sedimentary basins, the thickness of deposits, and the proportions of the various sedimentary rock types. The intervals can be recognized nearly everywhere that Pennsylvanian rocks occur in the United States; thus, comparison of genetically related events can be made between areas.

Each interval is composed of assemblages of members, formations, and groups that lie between recognizable lithologic contacts which may not and commonly do not coincide with time surfaces that are synchronous in the several regions. Use of the informal term "interval" and the problems of choosing interval boundaries were discussed by McKee and others (1959, p. 5).

BASIS FOR CORRELATION

Assignment of rocks to the Pennsylvanian intervals adopted here is based partly on relations shown in the Pennsylvanian correlation chart 6, compiled by the Committee on Stratigraphy of the National Research Council (Moore, 1944), but numerous revisions based on more recent studies and concepts have necessarily been introduced. Correlations are in part lithologic and in part paleontologic. The major divisions (intervals and

1Deceased June 3, 1970.
2The work was supported by the State Geological Survey of Kansas, University of Kansas, Lawrence, Kansas, William W. Hambleton, Director.
subintervals as used), the essentially equivalent series designations applied in the midcontinent region, and the correlative divisions or units in Pennsylvania and in Europe are shown in table 1.9

CHART OF STRATIGRAPHIC UNITS

A generalized chart (pl. 13) of Pennsylvanian rock-stratigraphic units in the United States, exclusive of Alaska and Hawaii, shows relative stratigraphic positions within columns, each of which represents rock units in a large area. Rock units in this chart are arranged in horizontal rows corresponding to the Pennsylvanian intervals of this publication. Few details of stratigraphic relations are attempted in this chart; overlap, facies change, and intertonguing are barely suggested, and the time span represented by each formation is shown only in a general way.

On plate 13 and throughout this paper, stratigraphic names that have not been adopted by the U.S. Geological Survey and those that have not occasioned any official action are shown in italic. Stratigraphic names adopted by the Geological Survey are not italicized, but where such names are applied to units that are either definitely or probably not the same as those of the type area, the names are enclosed in quotation marks. Because the nomenclature used in this report is from many sources, age assignments of some stratigraphic units may differ from those in use by the U.S. Geological Survey.

CONSTRUCTION OF PALEOTECTONIC MAPS

SOURCES OF DATA

This publication, like the earlier paleotectonic investigations, is an effort to synthesize available information on the system for which it was prepared. The sources of this information are given in the "Index to

---

9This report was prepared before the decision of the U.S. Geological Survey to put “an” and “ian” endings on provincial series terms. Therefore, the Pennsylvanian series names appear as Virgil, Missouri, Des Moines, Atoka, and Morrow in this report.

---

Table 1.—Interval assignments for Pennsylvianian

<table>
<thead>
<tr>
<th>Interval</th>
<th>Midcontinent series</th>
<th>Pennsylvania divisions</th>
<th>European divisions</th>
</tr>
</thead>
<tbody>
<tr>
<td>E</td>
<td>Virgil</td>
<td>Monongahela Group</td>
<td>Stephanian</td>
</tr>
<tr>
<td>D</td>
<td>Missouri</td>
<td>Conemaugh Group</td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>Des Moines</td>
<td>Allegheny Group</td>
<td>Westphalian</td>
</tr>
<tr>
<td>B</td>
<td>Atoka</td>
<td>Pottsville Group</td>
<td></td>
</tr>
<tr>
<td>A</td>
<td><strong>Morrow</strong></td>
<td></td>
<td><strong>Upper Namurian</strong></td>
</tr>
</tbody>
</table>

1In some regions interval A is divided into subintervals A1 and A2. Subinterval A1 includes the Morrow part of the Springer Formation and, in a few places, some rocks of Mississippian age.
Localities and Sources” at the end of part II and in the list of references following each chapter. Both published and unpublished information obtained from geologists in private industry, universities, and government surveys was summarized on data cards, one for each formation or interval at each locality or for the entire geologic system at the locality.

STRUCTURAL TERMS

Some of the structural terms used in describing tectonic features illustrated on the maps are shown in figure 2.

MAP OF CONTROL POINTS

The location of Pennsylvanian stratigraphic sections and of wells that furnished data used in the preparation of maps in this paper is shown on plate 1.

Purposes of the map of control points and an accompanying index (in part II) are to enable the reader to (1) compare data used in preparation of this paper with data from other sources, (2) compare data in this paper with those in other paleotectonic investigations, (3) evaluate the relative significance and reliability of various parts of the maps, and (4) prepare comparable but more detailed maps of local areas by the addition of supplementary data.

Density of control points indicates the relative amount of information available and, when considered in conjunction with an appraisal of the quality of data, makes possible an evaluation of the reliability of isopachs and of lithofacies trends. Control points, in general, are closely spaced in areas where much drilling has been done.

The maximum practical density of control is about one point per township. Where data for additional points are available, even though these points could not be shown, the data were studied and compared with those selected for map representation.

The paucity of control points in some areas is also significant, for it calls attention to places where additional data are needed.

In some areas, points are so widely spaced that interpolation is not warranted. In such places, thickness estimates and rock symbols for isolated points are shown, but isopachs are not used.

Control points in many places represent stratigraphic sections or wells that have been described by more than one source. Some of these alternative descriptions are virtually in agreement; some are not. Where there is disagreement, only one interpretation can be shown on the principal maps, but some others are discussed in the text.

INDEX TO LOCALITIES AND SOURCES

An index of sources to accompany plate 1 is given in “Index to Localities and Sources” at the end of part II. Abbreviations used in this index are explained at the beginning of the list. Entries in the list provide information that will allow the reader to identify the original source of data for each locality and thereby obtain the data, if desired. A complete listing of publications and unpublished theses used in the compilation of data for each region is given in the “References” at the end of each chapter.

TYPES OF MAPS AND SECTIONS

Maps and sections constructed from stratigraphic data gathered for this project constitute the third part of this volume. Most of the maps, and all of those published at a scale of 1:5,000,000, are dominantly factual stratigraphic syntheses. Included are (1) a paleogeologic map showing the distribution of rock units at the surface upon which the sediments of the system were deposited, (2) a map showing the geologic units that now directly overlie the system, (3) a summary isopach map indicating the total thickness of the system, (4) maps indicating thickness for each of the several units, or intervals, into which the system is divided, and (5) maps showing lithofacies for each interval. The interval maps, when compared with each other, convey an impression of the dynamic changes which influenced distribution and nature of the sedimentary rocks.

In order to illustrate, in part, the third dimension of the various intervals represented on the isopach and

![Figure 2.—Terms for structural features as used in this paper.](image-url)
lithofacies maps, a series of cross sections for selected areas has been prepared. These can be used effectively in conjunction with the factual maps and are tied in directly with control points on those maps. They illustrate thicknesses and rock sequences in vertical sections oriented in various directions across the several regions.

Interpretive maps of two types, based largely on data derived from the factual maps, have been developed and serve to give the authors' views of original features of each interval immediately following the time of deposition.

**ISOPACH MAPS**

The isopach maps (pls. 3A-9A, 11) for each interval and for the whole system show thickness of rock units preserved in the stratigraphic record. These maps constitute a rigorous statistical representation of available data but do not indicate for all areas the complete original thickness. Inconsistencies result where erosion has removed part or all of the record or where basal parts of the sections are not exposed or are not penetrated, and caution is necessary in applying the data where information on the entire sedimentary product of a particular time is needed. Thus, these maps in their present form serve as statistical summaries of available data on thickness.

In many regions, thickness data are abundant and are sufficiently complete and reliable to serve as a basis for discussion of paleotectonic implications. For such regions, places of maximum deposition, trends of major depressions, and approximate limits of deposition can be determined. In some other areas, however, paucity of data allows only a few generalizations.

Thickness values for only a few selected points are shown on each map; however, all points shown were used for control in constructing the isopachs. Information provided by these maps may be interpretive to some degree, for geologic bias can be introduced in the assignment of beds to a mapped unit and, in places, the control permits alternative construction of isopachs.

In much of the Basin and Range province of the Western United States, isopachs have been drawn to fit the limited amount of control available. This construction implies that rocks of the mapped unit are known to be present beneath the broad basin areas. In only a few places, however, have wells penetrated pre-Pennsylvanian rocks in these areas, and this construction of isopachs must be regarded as an interpretive reconstruction for the sake of providing as complete a treatment of the United States as possible.

In the Pacific coast region, points are so widely spaced that interpolation between points does not seem justified. Information is sparse and indefinite; only a few tentative thickness values are shown on the total system isopach map, and a few symbols indicating presence of rocks are shown on interval isopach maps.

**LITHOFACIES MAPS**

The lithofacies maps (pl. 3B-9B) for each interval indicate the distribution of average rock types, as preserved today, of the mapped units. Care must be used in interpreting the rock types shown on these maps because they do not necessarily represent all of the originally deposited unit. In some places sections are not complete because the base of the unit is not exposed or has not been penetrated by drill or because the upper part of the unit may have been removed by erosion; in neither case can the original proportions of rock types be computed.

The lithofacies maps show only the relative abundance of various rock types for each area, not how these types are distributed in vertical section or in stratigraphic relation to one another. For this reason, cross sections (pls. 10A-F) have been prepared to show, within the scale limitations, the thickness of rock types and the vertical distribution and stratigraphic relations between rock types.

Lithofacies as shown on the maps were determined by computation of the proportions of rock types in an interval. The colors symbolize lithologic combinations of sandstone, mudstone, carbonate rock, and evaporite. The classification chart that accompanies each lithofacies map is based on these four end-member rock types. Construction and use of the chart were discussed in detail in "Paleotectonic Maps of the Permian System" (McKee and others, 1967, p. 58).

Special facies overprints are used to denote areas in which chert, halite, or volcanic rock in the mapped unit exceeds 10 percent. The determination of these percentages is only approximate, but the environmental implications may be considerable.

Only selected faults have been shown. Faults known to have been active in Pennsylvanian time are shown in black; those active only after the Pennsylvanian are shown in red.

As with isopachs in parts of Western United States, lithofacies patterns have been extended across the broad valleys in a large part of the Basin and Range province. This implies that rocks of the mapped unit are known to be present beneath the valleys; however, lithofacies colors in these places should be recognized as interpretive reconstructions made to provide continuity for as large a part of the United States as possible.

**PALEOGEOLOGIC MAP AND MAP OF OVERLYING UNITS**

The paleogeologic map (pl. 2) and the map of overlying units (pl. 12) provide bracketing information for the Pennsylvanian System. They show the systemic breakdown of the underlying and overlying rocks; in ad-
dition, the Mississippian on the map of underlying rocks and the Permian on the map of overlying rocks are divided into series or series equivalents. Letter symbols indicate selected group names or formation names in the general area of their usage. No attempt has been made to draw boundary lines between groups or formations of comparable ages.

CROSS SECTIONS

In order to show the thickness, vertical distribution, and lateral relations of rock types, 44 cross sections (pls. 10.4-F) are included for use with the thickness and lithofacies maps. Lithofacies colors on the lithofacies maps indicate a composite of rock types in the interval mapped and may represent one of many possible combinations of different rock types. For example, a color indicating carbonate rock mixed with some sandstone may represent a single homogeneous unit of sandy limestone or interbedded layers of sandstone and carbonate rock; the layers may be a few thick strata or they may consist of many thin strata; the sandstone may be calcareous and the carbonate rock may be sandy; and one component may dominate in the lower or upper part of the section or may be eratically distributed. The cross sections permit a more detailed presentation than the lithofacies maps but, like the maps, are limited in detail by the publication scale.

The cross sections are constructed with a horizontal datum at the top. This datum represents the surface of the interval, or the system, at the end of deposition of the unit. The datum is projected across areas where erosion has removed the uppermost deposits, and the position of the top of the remaining rocks below the projected datum was determined by comparison of remaining strata with complete sections and by estimation of the thickness of missing rock. Use of a reconstructed datum results in a cross section that graphically shows the amount of regional sinking during deposition of the sediments, but, to the extent that the amount of erosion is an estimate, the cross sections are interpretive.

The sections show a number of formation names or group names without lateral bounding lines to indicate the geographic limit of terminology usage; many names must be omitted because of space limitation. The sections also indicate, beneath the sections, named structures of Pennsylvanian age and, above the sections, named post-Pennsylvanian structures or geographic features and the age of erosion episodes.

INTERPRETIVE MAPS: RESTORED ISOPACH AND PALEOTECTONIC

Interpretive maps, on a scale of 1:10,000,000, are based on all available information used during preparation of the factual maps and on other data discussed in the text or shown on special maps and in figures. The interpretive maps depict the writers' views regarding the former extent and thickness of the intervals, the places in which subsequent erosion has removed part or all of the record, and the location of various tectonic features. These maps are of two types, referred to as restored isopach maps (pl. 15) and paleotectonic maps (pl. 15). Both types have been prepared for each of the five intervals into which the Pennsylvanian System has been divided.

Restored isopach maps are intermediate in degree of subjectivity between the standard factual isopach maps of this publication and the tectonic maps to be described next. The maps of restored isopachs show at least four features that are important to geological interpretation: (1) the entire area currently occupied by rocks of a given interval (shown in green) and the reconstructed original limits of the interval (indicated by a zero isopach); (2) the inferred original thickness of the rocks of a given interval (shown by isopachs); (3) the probable directions of sediment transport, relative volumes of material, and coarseness of sediment involved (indicated by arrows); and (4) positive and negative structural features, such as basins, platforms, shelves, and geosynclines (indicated by labels).

Paleotectonic maps as prepared for this publication are based on the preceding interpretive isopach maps. They are designed to show in color and by symbols the tectonic features of the region. Colors delineate both positive areas with a large amount of uplift or with little or no uplift and negative areas with small subsidence, large subsidence, or extreme subsidence. Symbols indicate such features as axes of major tectonic structures, zones of overthrusting, areas of slight deformation, areas of severe deformation, and volcanoes and volcanic deposits.

ACKNOWLEDGMENTS

The writers are indebted to many individuals and organizations for basic data and ideas. The contributors of these data are indicated in the section "Index to Localities and Sources." In addition, we acknowledge the especially noteworthy assistance of the following (affiliation at time data were given):

Adams, J. E. Standard Oil Co. of Texas Midland, Tex.


Agniew, A. F. South Dakota State Geologist Vermillion, S. Dak.

Ball, S. M. Pan American Petroleum Corp. Tulsa, Okla.


Branson, C. C. University of Oklahoma Norman, Okla.

Brown, S. L.
Kansas Geological Survey
Lawrence, Kans.

Burchett, R. R.
Nebraska Geological Survey
Lincoln, Nebr.

Cady, W. M.
U.S. Geological Survey
Denver, Colo.

Campbell, H. A.
Atlantic Refining Co.
Oklahoma City, Okla.

Caplan, W. M.
Arkansas Geological Commission
Little Rock, Ark.

Carlson, M. P.
Nebraska Geological Survey
Lincoln, Nebr.

Clark, J. M.
Consulting Geologist
Fayetteville, Ark.

Clements, E. D.
Pan American Petroleum Corp.
Lubbock, Tex.

Culbertson, W. C.
U.S. Geological Survey
Denver, Colo.

Cunningham, B. J.
Texas Panhandle Sample Log Service
Amarillo, Tex.

Danner, W. R.
University of British Columbia
Vancouver, British Columbia

Earle, D. H.
U. S. Geological Survey
Austin, Tex.

Englund, K. J.
U. S. Geological Survey
Washington, D. C.

Flint, N. K.
University of Pittsburgh
Pittsburgh, Pa.

Foster, D. L.
Consulting Geologist
Denver, Colo.

Gilmour, E. H.
University of Montana
Missoula, Mont.

Gries, J. P.
South Dakota School of Mines
Rapid City, S. Dak.

Hadley, J. B.
U.S. Geological Survey
Washington, D.C.

Haley, B. R.
U.S. Geological Survey
Little Rock, Ark.

Hayes, P. T.
U.S. Geological Survey
Denver, Colo.

Hendricks, T. A.
U.S. Geological Survey
Denver, Colo.

Hite, R. J.
U.S. Geological Survey
Denver, Colo.

Hollingsworth, R. V.
Palaeontological Laboratory, Inc.
Midland, Tex.

Jacques, T. E.
Kansas Geological Survey
Lawrence, Kans.

Kinell, C. B., III
Kansas Geological Survey
Lawrence, Kans.

Krieger, M. H.
U. S. Geological Survey
Menlo Park, Calif.

Lantz, R. J.
U.S. Geological Survey
Washington, D.C.

McMannis, W. J.
Montana State University
Bozeman, Mont.

Maher, J. C.
U.S. Geological Survey
Menlo Park, Calif.

Merriam, D. F.
Kansas Geological Survey
Lawrence, Kans.

Meyer, R. F.
U.S. Geological Survey
Washington, D.C.

Mitchell, J. G.
American Stratigraphic Co.
Denver, Colo.

Momper, J. A.
Pan American Petroleum Corp.
Tulsa, Okla.

Moore, R. C.
Kansas Geological Survey
Lawrence, Kans.

Mudge, M. R.
U.S. Geological Survey
Denver, Colo.

Oriel, S. S.
U.S. Geological Survey
Denver, Colo.

Reed, E. C.
Nebraska State Geologist
Lincoln, Nebr.

Rice, E. M.
Trowbridge Sample Service
Jackson, Miss.

Roen, J. B.
U.S. Geological Survey
Washington, D.C.

Schuman, R. L.
Kansas Geological Survey
Lawrence, Kans.

Scott, H. W.
Michigan State University
East Lansing, Mich.

Simon, J. A.
Illinois Geological Survey
Urbana, Ill.

Skees, W. A.
Pan American Petroleum Corp.
Jackson, Miss.

Stewart, J. H.
U.S. Geological Survey
Menlo Park, Calif.

Stone, C. G.
Arkansas Geological Commission
Little Rock, Ark.

Takken, Suzanne
Mobil Oil Company
Oklahoma City, Okla.

Taylor, I. D.
Texas Panhandle Sample Log Service
Amarillo, Tex.

Tyrrell, W. W., Jr.
Pan American Petroleum Corp.
New Orleans, La.

Verville, G. J.
Pan American Petroleum Corp.
Tulsa, Okla.

Viele, G. W., Jr.
University of Missouri
Columbia, Mo.

West, W. W.
Permian Basin Sample Laboratory
Midland, Tex.

Williams, N. F.
Arkansas Geological Commission
Little Rock, Ark.

Winston, D. O.
University of Montana
Missoula, Mont.

Witkind, I. J.
U.S. Geological Survey
Denver, Colo.

Wood, G. H., Jr.
U.S. Geological Survey
Washington, D.C.

Yates, R. G.
U.S. Geological Survey
Menlo Park, Calif.

The following State geological agencies contributed extensively to the work of Dr. H. R. Wanless in this publication:

Geological Survey of Alabama
University, Ala.

Illinois State Geological Survey
Urbana, Ill.

Indiana Department of Natural Resources
Geological Survey
Bloomington, Ind.

Iowa Geological Survey
Iowa City, Iowa

Kansas Geological Survey
Lawrence, Kans.

Kentucky Geological Survey
Lexington, Ky.

The following State geological agencies contributed extensively to the work of Dr. H. R. Wanless in this publication:

Geological Survey of Alabama
University, Ala.

Illinois State Geological Survey
Urbana, Ill.

Indiana Department of Natural Resources
Geological Survey
Bloomington, Ind.

Iowa Geological Survey
Iowa City, Iowa

Kansas Geological Survey
Lawrence, Kans.

Kentucky Geological Survey
Lexington, Ky.
REFERENCES


New England

By HAROLD R. WANLESS and KENNETH G. BELL

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-B
CONTENTS

Abstract .................................................................................................................. 9
Introduction .............................................................................................................. 9
Paleogeology ............................................................................................................ 9
Age assignment of basin rocks ............................................................................... 11
Pennsylvanian rocks of the New England basins ................................................... 11
   Narragansett basin ............................................................................................. 11
   Boston basin ...................................................................................................... 12
   Worcester locality ............................................................................................... 13
Geologic units directly above Pennsylvanian System .............................................. 14
Post-Pennsylvanian history .................................................................................... 14
References ............................................................................................................... 14

ILLUSTRATIONS

[For listing of plates (in separate case) see volume "Contents"]

   Figure 3. Map showing geographic features and localities in New England mentioned in text . . 10
   4. Map showing tectonic elements in New England in Pennsylvanian time ............ 10
INTRODUCTION

Sedimentary rocks in New England assigned to the Pennsylvanian System on the basis of fossil identifications occur at two localities: in the Narragansett basin of Rhode Island and Massachusetts, and in a large slice within a fault zone near Worcester, Mass. (figs. 3, 4). The so-called Norfolk, Woonsocket, and North Scituate basins to the north and west of the Narragansett basin are outlying structural remnants of that basin (fig. 4). Sedimentary rocks occupying these structural remnants are nonfossiliferous, but they are correlated lithologically with units underlying the fossiliferous rocks of the Narragansett basin. The Boston basin of Massachusetts is occupied mainly by nonfossiliferous sedimentary rocks, some of which may be tentatively correlated with rocks of the Narragansett basin on the basis of lithologic similarities. There is no conclusive evidence indicating that the Boston and Narragansett basins have ever been joined, although distances of less than 2 miles now separate the sedimentary rocks within them. Structural features of the area suggest that the basins were separated by an east-or east-northeast-trending mountain range, and movements on several large faults have reduced the original distance between them.

The stratigraphic sections of the Narragansett and Boston basins are not completely known. The central part of the Narragansett basin is partly submerged under Narragansett Bay, and much of the eastern part of the Boston basin is under Boston Bay. Most of the land areas of these basins are covered by glacial deposits, and a few very resistant strata form most of the outcrops. Also, the Bay areas are highly urbanized. Sedimentary rocks of both basins are poor aquifers and they contain no petroleum reservoirs and no mineral deposits; consequently, there has been very little deep drilling. Sm-Na amounts of low-grade meta-anthracite coal have been mined in the Narragansett basin. The search for and the mining of coal have provided most of the information on the fossiliferous part of the sedimentary sequence. The limited detailed knowledge of the stratigraphic sections of both basins has come mostly from small exposures along the bay shores, from excavations, and from a few deep wells and bore holes.

PALEOGEOLoGY

The basement of the Narragansett and Boston basins is a crystalline complex of metamorphic and igneous intrusive rocks. Ages of various units of the basement possibly range from late Precambrian to Middle Devonian. The only known fossiliferous rocks in the basement are some slightly metamorphosed remnants of the Weymouth Formation of Early Cambrian age and the Braintree Argillite of Middle Cambrian age, in the eastern part of the Boston basin, and some remnants of slaty rocks at North Attleboro, Mass., near the present

Deceased June 3, 1970.
east edge of the Narragansett basin, which contain a Lower Cambrian fauna (Shaler, 1888). All these remnants are isolated by enclosing intrusive rocks.

Other stratified rocks, lying in a belt generally to the north and west of the basins, are parts of a thick sequence of quartzite, slate, marble, and hornblende schist or amphibolite. These rocks are included in the Blackstone Series (Woodworth, 1899, p. 104-109) and in the Westboro Quartzite and Marlboro Formation (Emerson and Perry, 1903; Emerson, 1917). Precambrian ages have been assigned to them, but they may be of Cambrian age. Igneous intrusive rocks constitute the major part of the basement. Several episodes of plutonic activity are represented. The earlier rocks are metamorphosed, the later ones are not.

The Boston and Narragansett basins are near the southeast margin of a geosyncline in which tens of thousands of feet of predominantly pyroclastic sediments were deposited during early Paleozoic time. The source of the pyroclastic sediments seems to have been a shelf-type volcano, or volcanoes, situated somewhat to the east of the present basin sites. Only the lower part of the volcanic accumulation, represented by the metavolcanic rocks of the Blackstone Series and equivalent rocks that generally have been included in the Marlboro Formation, crops out near the present borders of the basin. This feature indicates that substantial erosion has occurred and suggests the possibility that the upper part of the volcanic mass may never have been deposited across this area.

The intrusive rocks were mostly, if not entirely, emplaced during and subsequent to the early Paleozoic volcanism. The most widespread unit is the Dedham Granodiorite and its comagmatic phases that form major parts of the basement of both basins and of the uplands surrounding them. These rocks also furnished substantial proportions of the detrital sediments of both basins. The Dedham, the oldest of the plutonic rocks of the area, has been assigned to the Precambrian by Billings (1929), to the early Paleozoic by LaForge (1932), and to the Devonian (?) by Emerson (1917) and by Foye and Lane (1934).

The latest metamorphism of the basement rocks, here as in most other parts of the New England area, occurred during the Acadian orogeny of Early to Middle Devonian age. During this episode, the rocks were folded and subjected to faulting and, probably, some of the younger intrusive rocks were emplaced. A lengthy period of degradation and erosion followed the metamorphism. Intrusive rocks occupying large areas were unroofed and, in many places, they were eroded to below the zone of intrusion breccias. Initially the elastic debris was almost completely swept out of the region, presumably because depositional basins had not developed.

The beginning of basin development coincided with local volcanism. Flows and pyroclastic sediments ejected from vents in the upland areas were spread across the margins of developing basins. These volcanic rocks rest unconformably upon the basement; in some places they rest directly on the crystalline rocks, and in other places they are separated from the crystalline rocks by lithified
soil or arkose ranging from a few inches to a few feet thick. The volcanic rocks consist of the Lynn Volcanic Complex at the north side of the Boston basin, the Mattapan Volcanic Complex at the south side of the Boston basin and north side of the Narragansett basin, and some small unnamed and uncorrelated remnants near the present margin of the Narragansett basin. These volcanic rocks furnished substantial amounts of detritus to the basins.

AGE ASSIGNMENT OF BASIN ROCKS

A tentative age range of Early to Late Pennsylvanian for the sedimentary rocks of the New England basins is shown in the correlation chart of the Pennsylvanian System (pl. 13). The coal measures of the Narragansett basin contain a flora dated as Late Carboniferous by Lesquereux (1889), whose publication has been used by all subsequent workers in the area as the basis of age assignment. No fossils have been found in the rocks of the outlying remnants of the Narragansett basin—the so-called Norfolk, Woonsocket, and North Scituate basins.

Fragmentary plant fossils from the Worcester locality were assigned a Carboniferous age by C. D. White (1912, p. 118), who further suspected that additional discoveries would reveal that the beds were of Pennsylvanian, possibly Pottsville, age. Study of additional specimens collected from this locality since 1965 has confirmed the Carboniferous age and indicated the probability of Pennsylvanian age (Grew and others, 1970).

No specimen that can be conclusively identified as a fossil has ever been found in rocks of the Boston basin. Fossil-like material has been described by Burr and Burke (1900) and by Pollard (1965). An age assignment of Pennsylvanian, or even Carboniferous, for the sedimentary rocks of the Boston basin must be based on a few lithologic similarities of the basin complex and comparison of its paleogeological history with those of the Narragansett basin.

Present knowledge of the sedimentary rocks of the Narragansett and Boston basins and of the Worcester locality does not permit division into the mapping intervals (A through E) of this publication. Therefore these rocks are designated as an undivided Pennsylvanian unit and are shown only on plate II.

PENNSYLVANIAN ROCKS OF THE NEW ENGLAND BASINS

Rocks assigned, or tentatively assigned, to the Pennsylvanian System in New England are discussed under Narragansett basin, Boston basin, and Worcester locality. These rocks are mapped on plate II only, and on that plate no isopachs, except zero lines, are shown.

NARRAGANSETT BASIN

Sedimentary rocks of this basin occupy an area about 55 miles long and 25 miles wide (Quinn and Oliver, 1962; Shaler and others, 1899) and occur also in three small outlying remnants, the so-called Norfolk, Woonsocket, and North Scituate basins. The original basin undoubtedly was much larger. On the basis of geologic considerations, Shaler (1899, p. 40) believed that the original basin must have extended far beyond its present limits and that it was reduced by erosion.

Parts of the present boundaries are mapped as faults (Emerson, 1917), and geophysical data indicate that faults determine its present boundaries to a greater extent than had been deduced from previous geological observations (K. G. Bell and others, unpub. data, 1971). The outlying remnants owe their preservation primarily to their positions in relatively downfaulted blocks. Sedimentation features indicate that the original basin was bounded to the north and west by an east- or east-northeast-trending upland or mountain range. It may have been drained by a river along a major northeast-trending axis (Mutch, 1964).

The stratigraphic section of the Narragansett basin is not known in detail. Parts of the section are well exposed along the shoreline, but the outcrops are discontinuous, locally are steeply tilted, and are displaced by faults; therefore, it is not certain that all of the separate parts of the section have been combined in proper sequence. Away from the shoreline, outcrops are too sparsely distributed to furnish much detail. A considerable facies change is encountered from the borders of the basin to the central part, and sections in various places are not likely to match except in some gross features. Only a few published records of deep wells and boreholes exist. These include logs of a hole 705 feet deep in the coal measures of the Rhode Island Formation at Seekonk, Mass., a hole 975 feet deep in the coal measures at Taunton, Mass., and a hole 492 feet deep in the Rhode Island Formation at Portsmouth, R. I. (Shaler and others, 1899). Also several parts of the section preserved in outcrops are described by Shaler, Woodworth, and Foerste (1899).

About 10,500 to 12,000 feet of strata, all detrital nonmarine sediments except for some coal, is estimated to have accumulated in this basin. Isopachs cannot be drawn with accuracy, because there is no detailed knowledge of the sedimentary rocks. Numerous faults throughout the basin and, locally, folds, plus irregularity of surface erosion, cause abrupt thickening and thinning of the sedimentary rocks. Most of the strata probably belong to Pennsylvanian intervals B, C, D, and E, but interval A may be represented in a basal conglomerate. Because the boundaries between the intervals are not established, and because the part of the total thickness assignable to each interval is unknown,
separate interval maps have not been prepared for the New England basins.

The Pennsylvanian section in the Narragansett basin and its outlying remnants is generally divided into the following units, named in presumed ascending order: Pondville Conglomerate, Bellingham Conglomerate (in the Woonsocket and North Scituate remnants only), Wamsutta Formation, Rhode Island Formation (which makes up at least three-fourths of the total thickness of the Pennsylvanian section), Purgatory Conglomerate, and Dighton Conglomerate. Some of these formations intergrade with other formations which are contemporaneous but which represent different facies. This relationship seems to apply especially to the conglomerates, which are best developed along the north and west sides of the basin.

The conglomerate deposits consist of coarse detritus that probably tended to concentrate at or near the heads of alluvial fans that spread out from uplands bounding the basin. Finer grained detritus that was transported toward the interior of the basin probably formed the thick Rhode Island Formation. Thus, this formation, in part at least, is contemporaneous with some of the conglomerate deposits. The coal measures in the upper part of the Rhode Island Formation are indicative of intergradingWARNUTTA Formation. The Pondville at the top localities is a conglomerate containing abundant very large boulders derived from the underlying Blue Hill Granite Porphyry. The Wamsutta consists of interlayered red and gray slates and arkosic sandstones. The combined thickness of the two formations attains a maximum of at least 3,300 feet.

The so-called Norfolk basin is about 25 miles long and 1 to 3 miles wide; it trends northeasterly, and at its southwest end it is joined to the Narragansett basin. It is mostly bounded by faults, but along parts of its northwest side it is bounded by the unconformable contact of sedimentary rocks and the crystalline basement. This structural remnant contains the type exposure represented by the Pondville Conglomerate and the overlying Wamsutta Formation. The Pondville at the type locality is a conglomerate containing abundant very large boulders derived from the underlying Blue Hill Granite Porphyry. The Wamsutta consists of interlayered red and gray slates and arkosic sandstones. The combined thickness of the two formations attains a maximum of at least 3,300 feet.

The so-called Woonsocket basin, about 6 to 8 miles northwest of the present Narragansett basin, is about 10 miles long and 2 miles wide; its strata form the Bellingham Conglomerate and Wamsutta Formation, which have an estimated combined thickness of about 1,000 feet (Quinn and Allen, 1950). The so-called North Scituate basin, lying about 6 miles west of the present Narragansett basin, is about 7 miles long and 3 miles wide; the estimated thickness of its strata is about 1,500 feet (Quinn, 1951).

Environmental and climatic implications of Pennsylvanian rocks of the Narragansett basin seem to indicate changing conditions. Red beds make up a significant portion of the lower part of the section and may indicate a dry climate. Coal deposits occur in the upper part of the section and probably indicate humid climate and swamp environment.

**BOSTON BASIN**

The Boston basin is an arcuate, wedge-shaped structural remnant of a much larger basin. It is about 15 miles wide at the coast and curves westward to southwestward for a length of 25 miles. Much of the eastern part is submerged beneath Boston Bay. The basin boundary on the north and northwest is an overthrust fault zone on which the crystalline basement rocks have overridden the sedimentary rocks. Its south and southeast boundaries are along a zone of vertical and high-angle normal faults which have displacements ranging from a few feet to possibly as much as a few thousand feet. The unconformable contact of sedimentary rocks on the crystalline basement is preserved at a few places along the south boundary. The present Boston basin is formed entirely from the southern part or lowland of the original basin whose size and shape are unknown.

The sedimentary deposits of the Boston basin consist of one almost complete alluvial fan, parts of at least three other fans, and fine-grained detritus carried into the basin lowland. The alluvial fans developed along the north and northwest flank of an upland or mountain range that separated the original Boston and Narragansett basins. Along most of the southern border, detrital sediments were deposited upon an eroded surface of the crystalline basement, but locally the sediments were deposited on intervening flows of the predominantly felsic Mattapan Volcanic Complex. Virtually nothing is known of the basement under the basin lowland. Inasmuch as the region was subjected to erosion for a long period prior to deposition of sediments, the entire basin floor probably consists of a crystalline complex.

Strata of the Boston basin are divided into a conglomerate facies (the Roxbury Conglomerate), an argillite facies (the Cambridge Argillite), and interlayered volcanic rocks of the Brighton Melaphyre. Some early investigators believed that the Cambridge Argillite overlay the Roxbury Conglomerate in conformable sequence (Woodworth, 1893; Sayles and LaForge, 1910; Emerson, 1917, p. 56-57). Recent studies made possible by the construction of several tunnels have shown that the conglomerate interfingers with and grades into the argillite. Only the uppermost part of the argillite may have been deposited subsequent to the conglomerate.

The Roxbury Conglomerate was divided by Laurence LaForge (in Emerson, 1917, p. 56-57) into three members, in ascending order, Brookline Conglomerate, Dorchester Slate, and Squantum Tillite. This sequence
of units has been recognized only in the almost completely preserved alluvial fan lying mostly in the south part of Boston. The conglomerate ranges from a boulder facies near the heads of fans to small-pebble facies interlayered with slate and sandstone at the toe. Its constituents are rocks of the crystalline basement complex and felsic volcanic rocks of the kinds comprising the Lynn and Mattapan Volcanic Complexes; locally, it includes a large proportion of basic volcanic rock from the Brighton Melaphyre. The matrix commonly has a greenish color. The conglomerate probably was deposited subaerially, except near and at the toe of fans where subaqueous stratification is predominant.

The Dorchester Slate Member of the Roxbury consists of subaqueously deposited, reddish and purplish “slate” and sandstone interlayered with thin beds of pebble conglomerate. Recent work by Dott (1961) and by others has led to a conclusion that the so-called Squantum Tillite Member is not of glacial origin as proposed by Sayles (1914) but represents mudflows or landslides, or perhaps both. Observations by Bell (coauthor of this paper) indicate that various outcrop areas of the “tillite” are at different stratigraphic levels near the top of the conglomeratic sequence, thus recording more than one depositional event.

The Cambridge Argillite was formed mainly from fine-grained detritus carried into a basin lowland. The sediment was deposited subaqueously in thin beds, and in most places it is varved. It is predominantly various shades of gray; a few beds or layers are somewhat reddish brown. Tuffaceous beds, ranging from a few inches to several feet thick, are interlayered with the argillite.

A few small remnants of green, red, and yellow quartzite have been named the Tufts Quartzite Member of the Cambridge Argillite by Billings (1929, p. 106). Whether this quartzite is a thin unit within the argillite or is the remnant of an overlying unit is not certain. Pebbles, cobbles, and boulders of red mudstone and arkosic sandstone occur in glacial and beach deposits in the northern part of the basin. Because these rocks have not been found in the uplands north of the basin, either as bedrock or in glacial deposits, they are presumed to have been derived locally from areas not covered by marshes.

The interlayered volcanic rocks consist of andesitic and basaltic flows and tuffs. In the eastern part of the basin some of the flows are pillow lavas. The subaqueous volcanic rocks have a greenish cast. The subaerial volcanic rocks of the central and western part of the basin are green, red, or gray. Tuff beds are various shades of red and pink, or they are almost white. Tuff is a major constituent of the argillite and of the matrix of the conglomerate.

The maximum thickness of strata in the Boston basin may be as much as 6,000 feet. The thickest part of the section is presumed to have been in the north-central or northern part of the present basin in what was the main part of the lowland of the original sedimentary basin. Thicknesses of several hundreds of feet for the conglomerate facies can be inferred from outcrops. The lack of a completely exposed section on any of the fans and the presence of faults of unknown displacement preclude any but crude estimates of thickness. No holes have been drilled through the stratified rocks to the basement complex in this basin. Construction projects, such as tunnels, have provided information on only short segments of the section.

A few deep borings have been made in the basin, but drilling has been limited, mainly because there are no coal deposits and the rocks are not good aquifers. A well drilled near Causeway Street, in Boston, was briefly described by Hunt (1875), who reported: “This boring was carried to a depth of 1,750 feet, and, though I have not been able to obtain an exact record of it, it is said to have been almost wholly in argillite or clay slate, though at the bottom a crystalline rock was reached.” If a crystalline rock was penetrated, this rock probably was one of the diabase dikes that are common throughout the basin. Another well, drilled to a depth of 2,503 feet near Providence Street, Boston, was briefly described by Crosby (1884). He reported that the rock which was penetrated to a depth of 2,200 feet was slate, and that the drill cuttings from below that depth indicated a conglomerate.

The absence of coal deposits and plant remains in the sedimentary rocks of the Boston basin may be indicative of a dry climate at the time of deposition. Red beds are not characteristic of these deposits, but they constitute a minor part of the conglomeratic facies believed to have been subaerially deposited; they may have been a large part of the overlying arkosic sandstone unit represented by remnants of the Tufts Quartzite Member. A major part of the basin probably was filled with aqueous sediments, indicating that the lowland was occupied by a body of water. Whether this body was an inland lake, a sea, or part of an ocean has not been determined. Volcanoes were active in the area throughout the time of deposition, and they may have been a factor in suppressing both plant and animal life, thus partly accounting for the absence of fossils.

**Worcester Locality**

Carbonaceous phyllite containing fossil plant remains and coal occurs in a fault zone near the northeastern part of Worcester, Mass. The fossiliferous rock probably is a slice that has been dragged along the hanging-wall side of an overthrust fault. All other rocks in the locality belong to older, nonfossiliferous, more highly metamorphosed formations, or they are of intrusive types. Fossil identifications indicate a possible Pennsylvanian age for the phyllite (White, 1912; Grew and others, 1970). The phyllite may be correlative with the
Narragansett basin coal measures lying about 30 miles to the southeast.

GEologic UNITS DIRECTLY ABOVE PENNSYLVANian SYSTEM

The Narragansett and Boston basins contain no consolidated strata younger than Pennsylvanian age. Glacial deposits of Pleistocene age cover much of the present land area of both basins. As in the northern Appalachians, this region may once have been partly covered by coastal plain sediments of Cretaceous and early Tertiary age, but, if so, these deposits were removed before glaciation of the Pleistocene. Remnants of coastal plain deposits are found near the east coast of Massachusetts, at Marshfield, on Cape Cod, and on Martha's Vineyard (Woodworth and Wigglesworth, 1934; Zeigler and others, 1960).

POST-PENNsylvanian HISTORY

The region within which the Narragansett and Boston basins lie was folded during the Appalachian orogeny and subsequently was disrupted by large-scale faults (K. G. Bell and others, unpub. data, 1972). The episode of regional faulting probably began during the Perman and continued through the Triassic. Many dikes, mostly diabase, were emplaced prior to cessation of faulting. These dikes are generally considered of Triassic age. Locally, coastal plain sediments may have been deposited during the Cretaceous and Tertiary, and the region was glaciated during the Pleistocene.

REFERENCES


Mutch, T. A., 1964, Pennsylvania paleogeography in the Narragansett basin [abs.]: Geol. Soc. America Spec. Paper 76, p. 120.


Woodworth, J. B., and Wiggleworth, Edward, 1934, Geography and geology of the region including Cape Cod, the Elizabeth Islands, Nantucket, Martha's Vineyard, No Mans Land, and Block Island: Harvard Coll. Mus. Comp. Zoology Mem., v. 52, 322 p.
Appalachian Region

By HAROLD R. WANLESS

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-C
## CONTENTS

<table>
<thead>
<tr>
<th>Abstract</th>
<th>17</th>
<th>Interval B—Continued</th>
<th>36</th>
</tr>
</thead>
<tbody>
<tr>
<td>Region defined</td>
<td>18</td>
<td>Stratigraphic relations</td>
<td>36</td>
</tr>
<tr>
<td>Paleogeology</td>
<td>19</td>
<td>Upper boundary of interval B</td>
<td>36</td>
</tr>
<tr>
<td>Units underlying Pennsylvanian</td>
<td>19</td>
<td>Thickness trends</td>
<td>37</td>
</tr>
<tr>
<td>Northern Appalachian area</td>
<td>20</td>
<td>Northern Appalachian area</td>
<td>37</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>22</td>
<td>Central Appalachian area</td>
<td>38</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>22</td>
<td>Southern Appalachian area</td>
<td>38</td>
</tr>
<tr>
<td>Lower boundary of Pennsylvanian</td>
<td>22</td>
<td>Lithofacies trends</td>
<td>38</td>
</tr>
<tr>
<td>Interval A</td>
<td>24</td>
<td>Sources and environments of deposition</td>
<td>39</td>
</tr>
<tr>
<td>Formations included</td>
<td>24</td>
<td>Sources</td>
<td>39</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>25</td>
<td>Environments of deposition</td>
<td>40</td>
</tr>
<tr>
<td>Upper boundary of interval A</td>
<td>25</td>
<td>Paleotectonic implications</td>
<td>41</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>25</td>
<td>Interval C</td>
<td>42</td>
</tr>
<tr>
<td>Northern Appalachian area</td>
<td>25</td>
<td>Formations included</td>
<td>42</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>27</td>
<td>Stratigraphic relations</td>
<td>42</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>27</td>
<td>Upper boundary of interval C</td>
<td>44</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>27</td>
<td>Thickness trends</td>
<td>44</td>
</tr>
<tr>
<td>Northern Appalachian area</td>
<td>27</td>
<td>Lithofacies trends</td>
<td>44</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>28</td>
<td>Sources and environments of deposition</td>
<td>45</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>29</td>
<td>Sources</td>
<td>45</td>
</tr>
<tr>
<td>Sources, environments of deposition, and paleotectonic implications</td>
<td>29</td>
<td>Environments of deposition</td>
<td>45</td>
</tr>
<tr>
<td>Subinterval A</td>
<td>29</td>
<td>Paleotectonic implications</td>
<td>46</td>
</tr>
<tr>
<td>Formations included</td>
<td>29</td>
<td>Interval D</td>
<td>46</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>30</td>
<td>Formations included</td>
<td>46</td>
</tr>
<tr>
<td>Anthracite fields area</td>
<td>30</td>
<td>Stratigraphic relations</td>
<td>46</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>30</td>
<td>Upper boundary of interval D</td>
<td>47</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>30</td>
<td>Thickness trends</td>
<td>47</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>31</td>
<td>Lithofacies trends</td>
<td>48</td>
</tr>
<tr>
<td>Anthracite fields area</td>
<td>31</td>
<td>Sources and environments of deposition</td>
<td>49</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>31</td>
<td>Sources</td>
<td>49</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>31</td>
<td>Environments of deposition</td>
<td>49</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
<td>31</td>
<td>Paleotectonic implications</td>
<td>50</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>31</td>
<td>Interval E</td>
<td>50</td>
</tr>
<tr>
<td>Subinterval A</td>
<td>32</td>
<td>Formations included</td>
<td>50</td>
</tr>
<tr>
<td>Formations included</td>
<td>32</td>
<td>Stratigraphic relations</td>
<td>50</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>32</td>
<td>Upper boundary of interval E</td>
<td>50</td>
</tr>
<tr>
<td>Upper boundary of subinterval A</td>
<td>32</td>
<td>Thickness trends</td>
<td>51</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>33</td>
<td>Lithofacies trends</td>
<td>51</td>
</tr>
<tr>
<td>Anthracite fields area</td>
<td>33</td>
<td>Sources and environments of deposition</td>
<td>52</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>33</td>
<td>Sources</td>
<td>52</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>33</td>
<td>Environments of deposition</td>
<td>52</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>33</td>
<td>Paleotectonic implications</td>
<td>53</td>
</tr>
<tr>
<td>Anthracite fields area</td>
<td>33</td>
<td>Total thickness of Pennsylvanian rocks</td>
<td>53</td>
</tr>
<tr>
<td>Central Appalachian area</td>
<td>33</td>
<td>Thickness trends</td>
<td>53</td>
</tr>
<tr>
<td>Southern Appalachian area</td>
<td>33</td>
<td>Northern Appalachian area</td>
<td>53</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
<td>34</td>
<td>Central Appalachian area</td>
<td>54</td>
</tr>
<tr>
<td>Sources</td>
<td>34</td>
<td>Southern Appalachian area</td>
<td>55</td>
</tr>
<tr>
<td>Environments of deposition</td>
<td>34</td>
<td>Pennsylvania components of crystalline belts, Eastern United States</td>
<td>55</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>34</td>
<td>Geologic units directly above Pennsylvanian System</td>
<td>56</td>
</tr>
<tr>
<td>Interval B</td>
<td>35</td>
<td>Units overlying Pennsylvanian</td>
<td>56</td>
</tr>
<tr>
<td>Formations included</td>
<td>35</td>
<td>Paleotectonic implications</td>
<td>56</td>
</tr>
<tr>
<td>References</td>
<td>57</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
CONTENTS

ILLUSTRATIONS

[For listing of plates (in separate case) see volume “Contents”]

Figures 5-9. Maps showing.
5. Geographic features and localities in the northern Appalachian region .............................................. 20
6. Geographic features and localities in the central and southern Appalachian region .................................. 21
7. Tectonic elements in the central and southern Appalachian region in Pennsylvanian and post-Pennsylvanian time .................................................................................................................. 23
8. Tectonic elements as postulated by the author for the northern Appalachian region in Pennsylvanian and post-Pennsylvanian time .................................................................................................................. 24
9. Pennsylvanian coal regions and fields in the eastern United States ................................................................. 26
10. Diagrammatic section showing relative thicknesses of Lower and Middle Pennsylvanian rocks (intervals A, B, and C) in western and eastern parts of the northern Appalachian region ....................................................................................................... 53

TABLES

Table 2. Stratigraphic units mentioned in text for interval A but not shown on plate 13: Appalachian region .......... 28
3. Number of repetitions, by area, of lithologic types in subinterval A: Appalachian region .................................. 35
4. Stratigraphic units mentioned in text for interval B but not shown on plate 13: Appalachian region .................. 37
5. Number of repetitions, by area, of lithologic types in interval B: Appalachian region ....................................... 39
6. Stratigraphic units mentioned in text for interval C but not shown on plate 13: Appalachian region ................. 43
7. Stratigraphic units mentioned in text for interval D but not shown on plate 13: Appalachian region ............... 47
8. Differing assignments of Upper Pennsylvanian sequences to intervals D and E in the Anthracite fields of Pennsylvania ...................................................................................................................... 47
9. Stratigraphic units mentioned in text for interval E but not shown on plate 13: Appalachian region ............... 50
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES.
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

APPALACHIAN REGION

By Harold R. Wanless

ABSTRACT

Rocks of Pennsylvanian age in the Appalachian region extend from northeastern Pennsylvania and southwestern New York, southward to Alabama and Mississippi. The region has a length of 1,050 miles and a maximum northwest-southeast width, in Ohio and West Virginia, of 250 miles. It consists of parts of New York, Pennsylvania, Ohio, Maryland, West Virginia, Virginia, Kentucky, Tennessee, Georgia, Alabama, and Mississippi.

On the basis of differences in its history during Pennsylvanian time, this region may be divided into three areas: (1) The area of the Anthracite fields of northeastern Pennsylvania and the coal fields of the northern part of the Appalachian bituminous basin in Pennsylvania, Ohio, Maryland, New York, and West Virginia north of lat 38°; (2) the central part of the Appalachian bituminous basin, consisting of West Virginia south of lat 38°, Virginia, eastern Kentucky, and Tennessee; and (3) the southern part of the Appalachian bituminous basin including the Black Warrior basin and many smaller basins separated by anticlines or thrust faults in Georgia, Alabama, and Mississippi.

Massive limestones of the Meramec and the lower part of the Chester Series were more resistant than other strata beneath the Pennsylvanian rocks, and they formed a low cuesta extending from central Ohio northeast to Pittsburgh, Pa. East of Pittsburgh these limestones were more argillaceous and did not form prominent uplands. The cuesta of Mississippian strata separated a northern Appalachian depositional basin from a southern one during interval A of the Pennsylvanian and in the early part of interval B time, but it was buried by sediment before the close of interval B time, when the basins were combined.

Strata of intervals A (subintervals A1 and A2) and B are the only rocks of Pennsylvanian age present in Tennessee, Georgia, Alabama, and Mississippi. The rocks of these intervals attain a thickness of more than 12,000 feet in Winston County, Miss., as opposed to about 100 feet in northern Pennsylvania and Ohio.

Interval C extends from the northern Appalachian region southeastward to Virginia and southeastern Kentucky, although paleontologic evidence for correlating rocks included in the interval is not entirely conclusive. Maximum thickness in Virginia is estimated as 1,100 feet, indicating substantial continued downwarp of the central part of the Appalachian region.

Interval D extends from the northern Appalachian region southeastward to northeastern Kentucky; its thickness is as much as 200 feet in Ohio, 500 feet in the bituminous coal field of Pennsylvania, and more than 1,000 feet in the Anthracite field. Interval E extends southward into West Virginia, terminating near the northeast border of Kentucky. It is about 400 feet along its northwestern margin and thickens to about 600 feet in northern West Virginia.

The major source of sediment deposited in the Appalachian geosyncline was a discontinuous series of uplands to the southeast, extending from the vicinity of Philadelphia, Pa., southwestward to eastern Alabama. Additional sources furnished detritus to the northern part of the bituminous field of Pennsylvania, including elevated areas in New York State, Ontario, and Quebec. The most elevated of these areas may have been the Laurentian highlands and the area of the present Thousand Islands along the St. Lawrence River. Chert pebbles in the earliest Pennsylvanian sandstones of northern Pennsylvania and Ohio were derived from Devonian rocks exposed in New York; a northern source for these pebbles is indicated by the crossbedding directional trends.

Coarse conglomerates in the Southern Anthracite field and in the southern part of the Cahaba coal field of Alabama indicate source areas 50 miles or less distant to the southeast during the times of interval A and part of interval B. These sources probably were uplands flanked by piedmont alluvial aprons of sand and gravel, beyond which the detritus was transported down river valleys that have been recognized in northern Ohio and Pennsylvania and in eastern Kentucky and Tennessee. Finer sediment was transported over greater distances to areas that were more distant, or it may have been derived from less elevated areas.

During the early part of Pennsylvanian time (subintervals A1 and A2 and interval B), the sea transgressed into the Appalachian basin along a basinal axis extending across Mississippi and Alabama. Periodically, the sea extended across Tennessee, Kentucky, and West Virginia northward to southern and central Ohio and Maryland. At times during the deposition of subinterval A and interval B, the seas spread westward from eastern Kentucky across the saddle between the Cincinnati arch and the Nashville dome into southern Indiana, western Kentucky, and southeastern Illinois. In the latter part of interval B time and during transgressions of intervals C and D, marine waters spread eastward across northern Indiana and Ohio, and locally extended into Pennsylvania. During one transgression, late in interval D time, the sea reached the Northern Anthracite field near Wilkes-Barre, Pa., but it did not reach eastward to the Appalachian field during interval E.

As pre-Pennsylvanian topography was gradually buried by sediment, some lithologic units, such as coals and underclays, were deposited widely on the nearly level basin floor. This widespread distribution of coal beds is evident in subinterval A1 in Alabama and Tennessee, in interval B in eastern Kentucky and southern West Virginia, and in interval C in Ohio and the western part of the bituminous coal field of Pennsylvania.
The lithologic successions of the Pennsylvanian System are very complex, and abrupt facies changes are normal. The vertical repetition of lithologic sequences in a given area is commonly described as cyclic sedimentation. An influx of coarse to medium detrital sediments from nearby elevated areas formed piedmont fans, stream-channel deposits, flood-plain deposits, deltas, or barrier islands. As the supply of detritus apparently dwindled, sedimentation slowed, and soils supporting land vegetation developed. Vegetable matter, later converted to coal, commonly accumulated in marshy or swampy areas, such as existed on delta platforms; the vegetation then was covered by marine mud or deposits of calcium carbonate in a transgressing sea. During a subsequent marine regression the earliest sediment commonly consisted of beds of mud in a prodelta environment, followed by lenticular deposits of delta distributary or river-channel sand. This sequence of events was repeated, with local variations, many times.

Explanations for the rhythms typical of Pennsylvanian deposits include (1) periodic rejuvenation of bordering highlands; (2) climatic change, which may have reduced erosion by covering the uplands with vegetation; (3) eustatic changes in sea level. Local variations in sedimentation doubtless resulted from abandonment of subdelta and shifts of river drainage to new outlets. All these factors probably played parts in the development of the complex successions in the Appalachian coal fields.

At the beginning of the Pennsylvanian Period, sediment was carried northward across the northern Appalachian geosyncline to deltas near the present northwest border. In the latter part of interval C time, the central part of the northern Appalachian geosyncline subsided irregularly, the maximum subsidence being near the southwest corner of Pennsylvania. A centripetal drainage pattern then developed in this area of subsidence, and, during later Pennsylvanian and Early Permian time, lacustrine limestones repeatedly formed in the resulting basin.

The central and southern parts of the Appalachian coal basin subsided more extensively than did the northern part. Strata of interval B are 50 feet thick in northern Ohio and 3,500 feet thick in western Virginia, a seventyfold thickening. Sedimentation seems to have kept pace with subsidence. In Mississippi and western Alabama, downwarping was accompanied by normal faulting of large displacement, with the downthrown side to the southwest.

Pennsylvania strata on the Chestnut Ridge, Laurel Hill, and Negro Mountain anticlines of Pennsylvania, Maryland, and West Virginia seem to be about one-half as thick as those in adjacent synclines, suggesting differential vertical movement along these folds during Pennsylvanian time.

Near the end of Pennsylvania time and in Permian time, parts of the Appalachian region were deformed into anticlines and synclines and were cut by thrust faults. These structural movements may have induced mass gravitational sliding. Everywhere in the Appalachian region—except in an area of southwestern Pennsylvania, northwestern West Virginia, and southeastern Ohio—upper beds of the Pennsylvanian System are absent, probably because of erosion but possibly because Upper Pennsylvanian rocks were never deposited. Where present, uppermost Pennsylvanian beds are conformably succeeded by Permian strata.

REGION DEFINED

The Appalachian region extends from southern New York and northeastern Pennsylvania southwest to Alabama. Strata of Pennsylvanian age have been traced under a cover of Mesozoic rocks from Alabama west into Mississippi and probably continue west beneath younger rocks of the Mississippi embayment to join the Mississippian rocks exposed in Arkansas. Although all intervals of Pennsylvanian rocks are represented in one or another part of the region, their thickness and lithologic sequence differ so much from place to place that the region is treated in three parts, as follows:

1. Northern Appalachian area, including a small area of southern New York, much of Pennsylvania, Ohio, Maryland, and West Virginia as far south as lat 38° N.

2. Central Appalachian area, including West Virginia south of lat 38° N., eastern Kentucky, Virginia, and Tennessee.

3. Southern Appalachian area, including Georgia, Alabama, and the area of unexposed Pennsylvanian rocks in Mississippi. A large part of this area is known as the Black Warrior basin (Mellen, 1947).

The northern Appalachian area has several outliers of Pennsylvanian rocks, the most important of which forms the Anthracite fields of northeastern Pennsylvania (fig. 9). Each of these is a remnant of a once continuous coal field, downfolded in a deep syncline in the folded Appalachians. The Broad Top coal field of south-central Pennsylvania is another outlier, about 25 miles east of the border of the main northern Appalachian area of Pennsylvanian rocks. Many smaller outliers occur in northern Pennsylvania and northern Ohio. Some of these are strings of outliers along synclinal axes and are separated from each other by anticlines from which Pennsylvanian strata have been eroded. The Georges Creek field is in the Georges Creek syncline of southern Pennsylvania, Maryland, and northwestern West Virginia, a syncline about 120 miles long that divides into two synclines and an intervening anticline in West Virginia. It is 10-20 miles from the border of the main coal field. The eastern part of the main coal field in Pennsylvania and Maryland encloses three anticlines—the Chestnut Ridge, Laurel Hill, and Negro Mountain (fig. 8)—from which rocks of Pennsylvanian age have been eroded.

Several structural features interrupt the continuity of outcrops of the Pennsylvanian rocks in the central Appalachian area. The Cumberland overthrust block (fig. 7), the forward edge of which forms Pine Mountain, extends for about 125 miles from near Jacksonsboro, Tenn., to Russell Fork of Big Sandy River, Va. The block moved about 10 miles to the northwest during faulting. A narrow band of Mississippian strata just above the fault extends the length of Pine Mountain, and borings that penetrated below the thrust plane showed no Pennsylvanian strata beneath. The Paint Creek-Warfield fault zone crosses eastern Kentucky and either bends northeastward into West Virginia, so that it trends in the general direction of the Chestnut Ridge anticline of Pennsylvania, or it continues eastward.
The Sequatchie Valley anticline of southern Tennessee divides the coal field of Tennessee into two segments, the Cumberland Plateau to the northwest and Walden Ridge to the southeast. Walden Ridge extends into Alabama (where it is called Sand Mountain) and has a total length of 165 miles, 75 miles of which is in Tennessee. This structure probably resulted from the shift of a flat thrust to a higher stratigraphic level, as did similar structures described by Gwinn (1964). The thrust sheet has been penetrated in deep drilling for oil exploration. The Emery River fault zone is probably the end of the thrust block and corresponds to one of the two cross faults that occur at each end of the Cumberland overthrust sheet.

The central Appalachian area is divided by these structures into the following six parts:

1. Northeastern Kentucky north of the Paint Creek-Warfield fault zone.
2. Area between the Paint Creek-Warfield fault zone and the Pine Mountain thrust fault area, Kentucky and northernmost Tennessee.
3. Cumberland overthrust block, Virginia, Kentucky, and northern Tennessee.
4. Northern Tennessee area between the southwest end of the Cumberland overthrust block and the Emery River fault zone.
5. Cumberland Plateau northwest of Sequatchie Valley anticline.

The lattermost two zones extend southwestward into Alabama, but a gentle transverse arch near the Tennessee River restricts the Pennsylvanian rocks to small outliers on plateau remnants and, thus, largely divides the coal field in Tennessee from that in Alabama.

The Black Warrior basin of Alabama and Mississippi constitutes a large part of the southern Appalachian area, but the southeastern part of the region includes eight other distinct areas in which Pennsylvanian rocks occur. Listed in order from the northeast, these are

1. Mountain ridges in Chatooga and Floyd Counties, Ga.
2. Lookout Mountain syncline, in Alabama, Georgia, and a very small part of Tennessee.
3. Sand Mountain, Alabama and Georgia. This is the southwestward extension of Walden Ridge in Tennessee (discussed under the Middle Appalachian and not further discussed here).
4. The southward extension of the Cumberland Plateau of Tennessee, also referred to as part of the Plateau coal field of northeastern Alabama, west of the Sequatchie Valley anticline and north of the Tennessee River.
5. Blount Mountain, a syncline downfaulted on the northeast, in Blount, Etowah, and St. Clair Counties, Ala.
6. The Cahaba coal field southeast of Birmingham, Ala.
7. The Coosa coal field southeast of the Cahaba coal field, Alabama.
8. The Talladega slate belt of metamorphic rocks in Clay County, Ala., which includes the Erin Shale, a small remnant of presumably Pennsylvanian strata. The slate belt lies about 30 miles southeast of the Coosa coal field.

The Black Warrior basin (Mellen, 1947; McCalley, 1898, 1900; Metzger, 1965) forms a large triangular area bounded on the north by the southern flank of the Nashville dome and on the southeast by Appalachian folds and thrusts of the Sequatchie Valley (Browns Valley) anticline and faulted anticlines of the Birmingham-Bessemer area. On the southwest it appears to be limited by concealed thrusts of the Ouachita fault system, which may cross the Mississippi river to join the Black Warrior basin with the basin north of the Ouachita thrusts of Arkansas. Strata of the Black Warrior basin dip southward, and they thicken in the same direction mostly because of additions at the top, but probably also because of thickening of units within the section.

**PALEOGEOLOGY**

**UNITS UNDERLYING PENNSYLVANIAN**

Pennsylvanian strata overlie rocks of Mississippian age almost everywhere in the Appalachian region from northern and northeastern Pennsylvania to the Black Warrior basin (fig. 7) in Alabama and Mississippi (pl. 2). However, they rest on Upper Devonian strata of the Oswayo Formation in a few small outliers of southwestern New York and in Pennsylvania where strata of the Catskill Formation extend along the northeastern margin of the Northern Anthracite field.

Throughout the Appalachian region from Pennsylvania to Mississippi, uppermost Mississippian rocks are youngest toward the southeastern margin of the region, compared with areas on the northwest side. The paleogeologic map (pl. 2) differentiates by color the Mississippian rocks of Kinderhook, Osage, Meramec, and early, middle, and late Chester age. The Kinderhook strata immediately underlie the Pennsylvanian in Warren and McKean Counties (fig. 5) of northern Pennsylvania and in northeastern Ohio. Strata of Osage age underlie Pennsylvanian rocks along the northwest border of the Appalachian coal field (fig. 9) from central southern Ohio northeast to an area of numerous Pennsylvanian outliers in Potter, Tioga, and Bradford Counties of northeastern Pennsylvania. They also underlie Pennsylvanian strata in about one-half of the bituminous field in Ohio and one-third of the field in Pennsylvania.
Strata of the Meramec Series underlie Pennsylvanian rocks from Rockcastle County, eastern Kentucky (fig. 6), northeastward to the Northern Anthracite field as a continuous, generally narrow band. Facies changes in Pennsylvania between carbonate rocks to the west and terrigenous sediments farther east make difficult the correlation of Meramec strata across much of the State.

Lower and middle Chester rocks, Girkin Formation and Homberg Dolomite, form a subcrop band recognizable from the west border of the coal field in southernmost Kentucky and adjacent parts of northern Tennessee northeastward to the vicinity of Pittsburgh, Pa. This succession is easily recognized because of a prominent limestone, known as the upper part of the Greenbrier Limestone in West Virginia and Pennsylvania, the Maxville Limestone in Ohio, the Glen Dean Limestone and Golconda Formation in eastern Kentucky, and the Bangor Limestone in Alabama. Northeast of Pittsburg, Pa., Chester rocks of the Mauch Chunk Formation generally contain no limestone and have not been subdivided on the map.

Upper Chester rocks are included in the Mauch Chunk Formation in Pennsylvania and northern West Virginia. They are referred to, in ascending order, as the Hinton, Princeton, and Bluestone Formations of the Pennington Group in southern West Virginia and, locally, in Virginia. Elsewhere in Virginia, and in Kentucky, Tennessee, Georgia, Alabama, and Mississippi, they constitute the Pennington Formation.

In Alabama, carbonate rocks of Chester age in the Bangor Limestone change facies southeastward and are represented by mudstone in the Floyd Shale, the upper part of which is correlated with the Homberg Formation (lower and middle Chester) and is said to underlie the Pennington Formation (Weller and others, 1948). Strata referred to the Floyd Shale in the Cahaba and Coosa coal fields probably include correlatives of the Pennington. Mudstone directly underlying the Pennsylvanian in central Alabama is mostly of late Chester age.

**NORTHERN APPALACHIAN AREA**

A rock sequence known as the Maxville Limestone in Ohio and the Greenbrier Limestone in West Virginia.

*The identification of pre-Pennsylvanian formations in the Appalachian region north of lat 39°45' was supplied by Wallace de Witt (written commun., 1964). His identifications are at variance with formalizational identifications on many older well logs.*
Figure 6.—Geographic features and localities in the central and southern Appalachian region.
crops out in Ohio and in north-central West Virginia. In the parlance of well drillers, this sequence is called the Big lime. Overlying the limestone is a thin shaly mudstone unit locally referred to as the Pencil Cave, and above it, a limestone, generally about 10 feet thick, known informally as the Little lime, probably of Glen Dean age. Units designated as Big lime, Pencil Cave, and Little lime are commonly reported in oil-well logs, and the top of this sequence has been adopted as the base of the upper Chester in parts of plate 2.

The Cuyahoga Group includes elements of both Kinderhook and Osage age (Wallace de Witt, written commun., 1964); so, where the Cuyahoga underlies the Pennsylvanian in northern Ohio and central northern Pennsylvania, the Kinderhook-Osage contact cannot be located accurately on the paleogeologic map, and the two series are mapped as a unit.

In Pennsylvania, some beds in the Pocono Formation are red brown and resemble the red mudstones of the Mauch Chunk Formation. Because the diagnostic carbonate zone—the Burgoo Sandstone (locally a member of the Pocono)—that elsewhere separates the Mauch Chunk from the Pocono is absent in the Northern Anthracite field and other parts of northeastern Pennsylvania, the Pocono and Mauch Chunk Formations are difficult to distinguish in that region.

CENTRAL APPALACHIAN AREA

Pennington strata—shown by Winkler (1941) to be equivalent to the Kinkaid Limestone, which is the youngest formation of Chester age in the Mississippi Valley section—lie immediately below the Pennsylvanian in southern Tennessee, whereas successively older strata of the Pennington immediately underlie the Pennsylvanian northward. In northeastern Kentucky, Chester, Meramec, or Osage strata underlie the Pennsylvanian in outcrops.

Upper Mississippian strata thicken southeastward in Kentucky from a featheredge in northern Kentucky to about 750 feet near the southeast border of the State. Several of the limestone units of Chester age along the northwest border of the coal field wedge out eastward, and detrital sediments take their place. For this reason, little is known about the correlation of strata directly below the Pennsylvanian in southern West Virginia, Virginia, and Tennessee with the sequence of the Mississippi Valley. Here, strata probably a few hundred feet thick overlie the Kinkaid Limestone, which in Illinois and western Kentucky is the topmost formation in the Chester Series.

The Glen Dean Limestone of mid-Chester age underlies rocks of Pennsylvanian age in the Short Mountain outlier in Tennessee west of the coal field, but at all other localities strata referred to the Pennington directly underlie the Pennsylvanian.

SOUTHERN APPALACHIAN AREA

In northern Alabama, the Pennsylvanian Pottsville Formation is underlain by a wedge of mudstone and sandstone, from a few feet to 600 feet thick, assigned to the Parkwood Formation (Culbertson, 1963). In its typical area in the Cahaba coal field, the Parkwood, which is as much as 2,000 feet thick, is of Pennsylvanian age in the upper part and of Mississippian age in the lower part (Butts, 1940, p. 11). Thus, in the Cahaba and Coosa fields, the Parkwood represents continuous deposition across the Mississippian-Pennsylvanian boundary. In the Black Warrior basin and Plateau field area to the north, the upper part of the Parkwood presumably is Pennsylvanian, but the position of the boundary is uncertain. A fauna from the Parkwood on Blount Mountain, where the formation is 300 feet thick, consists of a mixture of Pennsylvanian and Mississippian forms (Butts, 1926, p. 206). In a cored well near locality 81, Chickasaw County, northern Mississippi, the 200-300 feet of strata partly equivalent to the Parkwood contains an invertebrate fauna identified as “Upper Mississippian (Chester) with Pennsylvanian elements” (Welch, 1959).

In northern Alabama, the Pennington Formation of Late Mississippian (Chester) age unconformably underlies the Parkwood Formation or, where the Parkwood is missing, the Pottsville Formation. These units, in turn, are underlain by the Bangor Limestone, also of Chester age (Welch, 1959). Southward, southeastward, and westward, the Bangor Limestone and locally the Pennington Formation interfinger with and grade into the Floyd Shale, a dark-shale sequence containing some sandstone beds (Welch, 1959; Thomas, 1967).

The presence of the Knox Dolomite (Lower Ordovician) directly below Jurassic rocks south of the Black Warrior basin suggests that Ordovician strata there had been thrust northward over Pennsylvanian strata during the Ouachita orogeny, probably in late Paleozoic time. Other wells in central Mississippi penetrate moderately metamorphosed slaty rocks beneath the Jurassic. Such wells are at localities 104 and 106 in Grenada County, Miss. (pl. 1). No fossil evidence is available to establish the age of these rocks, but the lithology resembles that of the subsurface low-rank metamorphic rocks in Texas called “Ouachita facies.” These rocks may be Pennsylvanian, Mississippian, or older.

LOWER BOUNDARY OF PENNSYLVANIAN

Uplift and truncation of the northwest border of the Appalachian geosyncline evidently occurred before the beginning of Pennsylvanian deposition, and the truncation seems to be deeper in the northern part of the region than elsewhere. Deposition was virtually con-
FIGURE 7.—Tectonic elements in the central and southern Appalachian region in Pennsylvanian and post-Pennsylvanian time. Thrusts in Alabama are boundary faults of blocks containing the Cahaba and Coosa coal fields (fig. 9).

Continuous along the southeast border; the Parkwood Formation in Alabama and Mississippi is believed to be Mississippian in its lower part and Pennsylvanian in the upper part (pl. 13). Other areas in which only minor disconformities occur between the two systems probably are in southern West Virginia and adjoining parts of western Virginia. An intertonguing of the Pennsylvanian and Mississippian Lee Formation with the Mississippian Pennington Formation is recorded (Englund and Smith, 1960; Englund, 1964, p. B32; Englund and DeLaney, 1966) between Cumberland Gap and Pennington Gap in westernmost Virginia and southeastern Kentucky. This relationship was determined by the physical tracing of units along Cumberland Mountain between the two gaps.

An intertonguing in northeastern Kentucky, recorded by Sheppard and Dobrovolny (1963), is similar to that in Virginia and southeastern Kentucky. The exposed Mississippian in the Kentucky area is of late Chester age, and perhaps the Lee Formation in that area is of Chester age.

In the northern part of the Appalachian geosyncline, Mississippian rocks, prior to their burial by Pennsylvanian sediments, probably were dissected into a fairly rough terrane comparable to that in the Illinois basin, as described by Siever (1951). The courses of some
eressional valleys in northern Ohio, filled with the Sharon Sandstone (or Conglomerate) Member of the Pottsville Formation (subinterval A1), have been mapped by Fuller (1955). Two fluvial systems draining southward discharged coarse gravels along the north border of the geosyncline. One in northeastern Ohio was described by Fuller (1955). The other drained southwestward into northwest-central Pennsylvania and formed the Olean Conglomerate Member of the Pottsville (Meckel, 1967). These drainage systems brought into the area many Devonian chert pebbles.

In central-eastern Kentucky (Rockcastle County), a major river valley existed at the time of initial Pennsylvanian sedimentation and extended westward across the saddle between the Cincinnati arch and Nashville dome (fig. 7) to the coal field in western Kentucky. This valley was incised more than 100 feet below the surrounding plain in eastern Kentucky and nearly 300 feet below in western Kentucky (Hagan, 1942).

The Upper Mississippian Big Lime in the northern Appalachian region was more resistant to erosion than were the sandstones and mudstones both above (Mauch Chunk Formation) and below (Pocono, Burgoo, and Logan Formations, and Cuyahoga Group). This carbonate rock unit formed a Niagara-type cuesta, with a northwestward-facing escarpment, between lowlands developed on the overlying and underlying detrital rocks. Until early Middle Pennsylvanian time, the escarpment extended as a divide from near Zanesville, Ohio, northeastward toward Pittsburgh, Pa., separating two basins of deposition (fig. 8). In central and eastern Pennsylvania a dividing ridge between basins probably did not exist in Early Pennsylvanian time. Resistant carbonate rocks of Meramec and early Chester age are absent. The somewhat older Burgoo Sandstone Member of the Pocono Formation and, in Ohio, the Black Hand Sandstone are both locally massive, but these rocks do not seem to have had the influence on topography that the Upper Mississippian carbonate beds had in western Pennsylvania and the adjacent parts of Ohio.

**INTERVAL A FORMATIONS INCLUDED**

Formations assigned to interval A in the northern part of the Appalachian region include the Olean Conglomerate Member of the Pottsville Formation of
southern New York, the Tumbling Run and Schuylkill Members of the Pottsville Formation of the Anthracite fields in Pennsylvania, and the Sharon Member of the Pottsville of the bituminous fields of Pennsylvania, Ohio, and Maryland (pl. 13). Interval A in West Virginia includes the New River and Pocahontas Formations of the Pottsville Group, and in Virginia and eastern Kentucky includes the Lee Formation, although its boundaries in these areas may not exactly coincide with those in areas where the New River occurs. Interval A in Tennessee includes the Crab Orchard Mountains and Gizzard Formations. In Alabama and Mississippi it includes the lower part of the Pottsville Formation and all of the Parkwood Formation, even though the lower part, in places, is Mississippian in age. The Erin Shale (phyllite) of Alabama may contain rocks assignable to interval A.

Interval A in the Appalachian region is divided into subinterval A₁, below, partially equivalent to the Springer Formation of the mid-continent area, and subinterval A₂, above, approximately correlative with the upper part of the Morrow Series. Separate maps have been prepared for subintervals A₁ and A₂ (pls. 4, 5); discussion of these maps follows the discussion of interval A (pl. 3).

**STRATIGRAPHIC RELATIONS**

Interval A in the Appalachian geosyncline is identified largely by fossil floral association. A succession of 14 floral zones in the upper Paleozoic is recognized (Read and Mamay, 1964). Floral zones 4-6 characterize interval A, and zone 7 corresponds to the base of interval B. Using West Virginia as a reference locality, Read and Mamay limited floral zone 4 to the Pocahontas Formation of Virginia and the Lykens Valley Nos. 5, 6, and 7 coals of the Southern Anthracite field, Pennsylvania (subinterval A₁). Floral zone 5 is limited to the lower part of the New River Formation of West Virginia and the Lykens Valley No. 4 coal of the Southern and Western Middle Anthracite fields. Zone 6 is in the upper part of the New River Formation and in the Schuylkill Member of the Pottsville Formation in the Anthracite fields. Rocks of floral zones 5 and 6 constitute subinterval A₂. A floral assemblage of zone 6, characterized by *Martioperis pygmaea, Neuropteris tennesseae*, and other forms, is common above the Lykens Valley Nos. 2 and 3 coals of the Schuylkill Member of the type Pottsville Formation in the Southern Anthracite field, Pennsylvania (fig. 9), the Sharon coal of western Pennsylvania and Ohio, the Sewell coal of West Virginia, and, possibly, the Sewanee coal of Tennessee and the Mary Lee coal of the Warrior coal field, Alabama (table 2); this floral assemblage also is in the Baldwin coal group in the Bloyd Shale of the Morrow Series in Arkansas.

**UPPER BOUNDARY OF INTERVAL A**

A large number of units in strata of interval A in West Virginia have received names, but all these units are nontectonic. The stratigraphic surface used to separate interval A from interval B throughout this field, therefore, probably is not a time plane. The tops of fairly widespread massive sandstone members of several formations (table 2)—including the Nuttall Sandstone Bed of the Sewell Member in West Virginia, the Bee Rock Sandstone Member in Virginia, the Corbin Sandstone Member in eastern Kentucky, and the Rockcastle Conglomerate Member in Tennessee—were used for the boundary.

In West Virginia, the uppermost part of the New River Formation is not very fossiliferous; consequently, the boundary between subinterval A₁ and interval B is difficult to pick consistently. Most subsurface records are drillers’ logs of oil wells in which interval A is traced only with difficulty. Sandstones of the New River Formation (subinterval A₂) are generally coarser grained and more permeable than those of the overlying Kanawha Formation (interval B), and statements, such as “salt sand” or “hole full of water,” in drillers’ reports necessarily had to be relied upon to distinguish interval A.

**THICKNESS TRENDS**

Thickness variations on the map of interval A undivided (pl. 3A) are influenced by the relation of subintervals A₁ and A₂ to the surfaces on which they were deposited. Subinterval A₁ is restricted in the Appalachian region to small areas near the southeast and south borders, where little or no break in sedimentation occurred between it and underlying Mississippian strata. Subinterval A₂ is much more widespread and is separated from Mississippian strata by a distinct unconformity. In a large part of the Appalachian region, subinterval A₂ is the only part of interval A represented. Thickness variations in the northern, central, and southern Appalachian areas are described separately.

**NORTHERN APPALACHIAN AREA**

Interval A has an irregular thickness; it reaches a maximum thickness of 243 feet in Portage County, Ohio, but is generally less than 100 feet thick. In northeastern Ohio, northern Pennsylvania, and southwestern New York, interval A consists of the Sharon Conglomerate Member of the Pottsville Formation and its correlative, the Olean Conglomerate Member, which have a scattered distribution. These conglomerates were formed in low places on an erosional surface and include stream-channel and delta deposits. No strata of interval A are recognized in the area from east-central Ohio east and northeastward to the east border of the bituminous coal field in central-southern Pennsylvania.

South of the belt in which interval A is missing, a conglomerate that is equivalent to the Sharon Conglomerate Member and associated nonconglomeratic
Figure 9.—Pennsylvanian coal regions and coal fields in the Eastern United States mentioned in text.

beds occurs in outcrops along the eastern and western margins of the Appalachian basin. These strata are progressively thicker southward through Maryland and northern and central West Virginia, attaining a
maximum thickness of 1,000 feet near the 38th parallel. Between the two outcrop belts, interval A is deeply buried by younger Pennsylvanian and Permian strata. In southern Ohio and northeastern Kentucky, sandstones of interval A are lenticular and are probably channel deposits.

In the Southern and Western Middle Anthracite fields, interval A has a maximum thickness of more than 1,000 feet. It thins within a short distance northward, disappearing along the northern margin of the Eastern Middle field, and is absent in the Northern field.

CENTRAL APPALACHIAN AREA

Interval A thickens southeastward from lat. 38° N. in southern West Virginia and Kentucky. Maximum thicknesses of nearly 2,000 feet occur in Wyoming and McDowell Counties, W. Va. (fig. 6). In eastern Kentucky the interval ranges in thickness from a few feet locally, in the northernmost counties, to 1,600 feet in Harlan County, near the Virginia line. Outcrops extend westward discontinuously from the coal field of eastern Kentucky to that of western Kentucky along the southern border of the Jessamine dome of the Cincinnati arch (fig. 7). In Tennessee interval A thickens from 126 feet in the Short Mountain outlier of Cannon County, at the east border of the Nashville dome, to about 1,500 feet along the southeastern border of Pennsylvanian strata (pl. 3A). In some places in West Virginia and Maryland, thicknesses of interval A are anomalously thin compared with those in nearby areas because locally the upper parts have been removed by erosion. Maximum thicknesses generally are near the southeast border of the coal field where the interval terminates abruptly against a post-Pennsylvanian thrust fault.

SOUTHERN APPALACHIAN AREA

Near and a little south of the southern Tennessee line a gentle upwarp crosses the Appalachian plateaus. Erosion has left less than 500 feet of strata of interval A on the crest of the fold, except in the Lookout Mountain syncline of Georgia and Alabama, where a maximum of 1,100 feet remains. About 12 miles south of the Tennessee line and northwest of the Sequatchie Valley anticline, Pennsylvanian outliers of interval A cap a series of high knobs. Maximum preserved thickness in this outcrop belt is in Jackson County, Ala., north of the Tennessee River, where 306 feet is exposed.

Farther south the section thickens greatly because rock units are progressively added to the column across the Black Warrior basin of Alabama and Mississippi. The thickness of rocks assignable to interval A is in some doubt because of incomplete data concerning fossil floras, but it is estimated at 4,750+ feet at a well in Sumter County, Ala. This well failed to reach the base of the Pennsylvanian. East of the Birmingham region, two narrow coal fields—Cahaba to the northwest and Coosa to the southeast—furnish floras that give information on the thickness of interval A. The maximum thickness tentatively assigned to interval A in the Cahaba coal field, in Bibb County, is 7,980 feet. In the Coosa coal field, in Shelby County, the maximum is 7,616 feet. In Clay County, Ala., the Erin Shale, which is metamorphosed to phyllite, is reported (Stearns and Mitchum, 1962, p. 84) to contain about 2,000 feet of Pennsylvanian (?) rocks; however, available fossil data are inadequate to determine whether any of these rocks belong to interval A.

LITHOFACIES TRENDS

The Appalachian and Black Warrior basins were sites where predominantly detrital sediments were deposited during interval A (pl. 3B). The entire column in each of these basins may be visualized as a pile of superposed detrital wedges, some of which thicken toward the source and others basinward, wherever downwarping made room for a large quantity of sediment. Sediment types range from fine clay to coarse conglomerate and include coal. Although conglomerate is not common, it makes up a larger proportion of interval A than of any succeeding intervals.

Some sandstone bodies maintain persistent thicknesses across moderate-size areas; but, because mostly they are the product of deposition in stream channels, delta distributaries, or barrier islands, they are generally lenticular. The total proportion of sandstone from place to place in the interval differs greatly. Twenty-four separate sandstones, 36 mudstones, 29 coals, and 4 underclays from West Virginia in the New River and Pocahontas Formations (interval A) were listed by Reger (1931a, p. 224-227). Many blanket sandstones occur in Alabama; one channel-fill sandstone, shown by Culbertson (1964, fig. 6), is several miles wide at places.

The distance between points of stratigraphic control in this region averages 6 miles. Many primary sedimentary features, such as channels of large rivers, delta distributaries, and barrier islands, are less than 1 mile wide. Thus, facies patterns are greatly generalized, and local trends cannot be clearly shown.

NORTHERN APPALACHIAN AREA

Interval A in northern Ohio, northern Pennsylvania, and southwestern New York is thin and generally includes only a single conglomerate or sandstone associated with shaly mudstones and coal. In most of southern New York and adjoining McKean County, Pa., and much of the area of northernmost Ohio, more than 80 percent of the interval is sandstone or conglomerate. Locally, however, the interval is mostly mudstone, probably because of deposition in interdistributary areas of deltas. Along the west border of the Appalachian coal field in Holmes, Coshocton, Perry, and
TABLE 2.—Stratigraphic units mentioned in text for interval A but not shown on

<table>
<thead>
<tr>
<th>Interval or subinterval</th>
<th>Western Pennsylvania</th>
<th>Pennsylvania Anthracite fields</th>
<th>Ohio</th>
<th>West Virginia</th>
<th>Virginia</th>
<th>Eastern Kentucky</th>
<th>Tennessee</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>Lower Connquenessing Sandstone Member</td>
<td>Potomac Formation</td>
<td>Potomac Sandstone Member</td>
<td>Nuttall Sandstone Bed, Sewell Member</td>
<td>Bee Rock Sandstone Member</td>
<td>Olive Hill Clay Bed</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sharon iron ore</td>
<td>Potomac Formation</td>
<td>Potomac Sandstone Member</td>
<td>Hartridge Shale</td>
<td>Lee Formation</td>
<td>Rockcastle Conglomerate Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sharon coal</td>
<td>Potomac Formation</td>
<td>Potomac Sandstone Member</td>
<td>Skelt Shale</td>
<td>Lee Formation</td>
<td>Crab Orchard Members Formation</td>
<td></td>
</tr>
<tr>
<td>A₂</td>
<td>Sharon Sandstone (or Conglomerate) Member</td>
<td>Shawnee Valley No. 2 and 3 coals</td>
<td>Jackson Shaft coal</td>
<td>Sewell coal</td>
<td>Sewee coal</td>
<td>Sewee coal</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lykens Valley No. 4 coal</td>
<td>Potomac Formation</td>
<td>Priceville Member</td>
<td>Beckley coal</td>
<td>Battle Creek coal</td>
<td>Gizzard and Crab Orchard Members Formation</td>
<td></td>
</tr>
<tr>
<td>A₁</td>
<td>Lykens Valley Nos. 5, 6, and 7 coals</td>
<td>Pocahontas Formation</td>
<td>Pocahontas No. 3 coal</td>
<td>Squire Jim coal</td>
<td>Pocahontas No. 3 coal</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Hocking Counties, Ohio, narrow belts of outcrop of interval A are composed mostly of shaly mudstone.

In southern Ohio, central and northern West Virginia, and Maryland, sandstone generally is more common than mudstone. Farther to the north and northwest, where interval A is thin, it contains only one or two sandy units, but sandstone may form more than 80 percent of the total. Toward the east border of the main coal field and the Georges Creek syncline of Maryland and West Virginia, mudstone locally exceeds sandstone in thickness. In this area the interval is thick and contains several sandstones interbedded with mudstones and coals.

The anomaly of a high sandstone percentage away from the source, which probably was to the east, may have resulted from unfavorable use of rather poor drillers' logs in which interval A was recognized largely on the basis of a coarse, highly permeable sandstone. Mudstones of the same age are not readily differentiated from those of interval B. Another factor which has a bearing on the cause of this anomaly is that subinterval A₁, the Pocahontas Formation, is restricted largely to the southeastern part of the area, and in that region it is more shaly than subinterval A₂, the New River Formation.

CENTRAL APPALACHIAN AREA

In southern West Virginia and western Virginia, sandstone is more abundant than mudstone, but only in a few places does sandstone compose as much as 80 percent of the total. In a substantial area in Mingo and Logan Counties, southern West Virginia, however, most wells for which records are available do not penetrate as deep as interval A, and the facies is poorly known.

In eastern Kentucky most logs of interval A show 80 percent or more sandstone, whereas in adjoining Virginia most records show 50-80 percent sandstone. The Virginia records are almost all logs based on systematically collected samples, whereas the Kentucky records are largely based on less reliable drillers' logs, which apparently err in reporting too much sandstone. Linear belts in Kentucky showing 50-80 percent sandstone, separating belts showing more than 80 percent sandstone, may merely be the result of differences in the quality of records. The outliers between the coal fields of eastern Kentucky and western Kentucky are composed almost entirely of sandstone but contain some quartz-pebble conglomerate.

Much of the Pennsylvanian System in the
Appalachian coal field has been extensively tested by the diamond drill for minable coals. Interval A does not contain much coal in the central Appalachian region, except in southern West Virginia, Virginia, and southern Tennessee, so it has been largely neglected in drilling. The quality of subsurface descriptions for this interval, therefore, is commonly poorer than that for higher intervals.

In Tennessee, except in the northeastern counties, interval A is represented in most of the area occupied by the Pennsylvanian System. Prominent sandstones of the Crab Orchard Mountains and Gizzard Formations in interval A cap most of the Cumberland Plateau and Walden Ridge, but the percentage of sandstones in the interval is generally less than 80 percent. The mudstone content is greatest in the west and may be attributed to a source of sediment east of Tennessee.

SOUTHERN APPALACHIAN AREA

The highest proportions of sandstone of interval A in the southern Appalachian area are in northeastern Alabama and northwestern Georgia, especially in outliers (pl. 3B), where only thin remnants of interval A are preserved. Farther west, in Mississippi, a rather large percentage of dark silty mudstone occurs, as shown by sample logs.

A thick succession of Pennsylvanian strata in west-central Alabama and east-central Mississippi evidently was deposited in a downwarped basin that trapped all grades of sediment and formed the maximum thickness of interval A in the Eastern United States. Marine fossiliferous zones of interval A are more common in Alabama and Mississippi than they are farther north in the Appalachian region. The distribution of coal in interval A throughout the Appalachian basin is discussed in a special section on coal in part II of this publication.

SOURCES, ENVIRONMENTS OF DEPOSITION, AND PALEOTECTONIC IMPLICATIONS

Source areas for the sediment probably lay to the east, as shown by measurements of the crossbedding directions in sandstones of the northern Appalachians. These source areas were probably parts of a chain of island uplifts, as suggested by Kay (1952), rather than a continuous highland of Appalachia, as formerly thought. Much of the mudstone may have accumulated in interdistributary areas of deltas on piedmonts, on flood plains, or in lakes. A few thin marine strata are also present.

Because the locations and conditions of sediment accumulation in subinterval A₁ were very different from those in A₂, detailed analysis of these subjects is given under those divisions of interval A.

SUBINTERVAL A₁ FORMATIONS INCLUDED

In the Anthracite fields the lower part (floral zone 4) of the Tumbling Run Member at the base of the Pottsville Formation is referred to subinterval A₁. In southern West Virginia the Pocahontas Formation, which is also referred to zone 4, is considered as subinterval A₁. In adjoining Virginia and eastern Kentucky, correlative strata are referred to the lower part of the Lee Formation.

In Alabama, the lower part of the Pottsville Formation (below the Black Creek coal group, or the equivalent Harkness coal bed) and the underlying Parkwood Formation of Mississippian and Pennsylvanian age are referred to subinterval A₁.²

²After the author's death, the following note was contributed by W. C. Culbertson to suggest an alternative interpretation based on recently obtained data of the age relationships of the Lower Pennsylvanian strata included in this report in Alabama and Tennessee: In Alabama, the lower part of the Pottsville Formation includes floral zone 5, considered part of subinterval A₁. The concept that it contained only floral zone 4 and was older than the Pennsylvanian rocks of Tennessee seems to have arisen from a series of revisions and oversights. David White is quoted by Charles Butts (1926, p. 210) as assigning the lower part of the Pottsville Formation (below the Black Creek coal group and Harkness coal bed) to "lower Pottsville," but floral zones diagnostic of "lower Pottsville" in Alabama were not specified. However, in West Virginia, the Pocahontas Formation had been assigned by White to the "lower Pottsville," and the Pocahontas is now known to contain only floral zone 4. The assumption that "lower Pottsville" was therefore restricted to zone 4 overlooked the fact that the Pocahontas had been redefined since White's time. The Pocahontas Formation of White (Continued)
In Mississippi, the Pennsylvanian strata are concealed beneath Mesozoic sedimentary rocks, and they are generally referred to as either the Pottsville Formation or "the Pennsylvanian rocks." An attempt has been made to project units referred to subinterval A1 in Alabama into Mississippi by correlation of electric-log patterns.

Floral zone 4 of Read and Mamay (1964) is found in the lower Lykens Valley coal beds (lower part of Tumbling Run Member) in the Southern Anthracite field of Pennsylvania and in the Pocahontas Formation of West Virginia and its equivalent in the lower part of the Lee Formation of Tazewell County, Va. The Lee Formation—once considered to be entirely Pennsylvanian in age, but probably also in part Mississippian—inter­tongues in westernmost Virginia with the Pennington Formation, considered to be of Mississippian age (Englend and Smith, 1960).

No well-established floral-zone fossils mark the top or base of subinterval A1; therefore, strata combined on isopach and lithofacies maps of this interval may not represent exactly the same span of time in all places. For this reason, at some places undifferentiated formations or groups have been referred to subinterval A1. The top of the "lower Pottsville" was tentatively placed at the Black Creek coal group by White, according to Butts (1927).

THICKNESS TRENDS

ANTHRACITE FIELDS AREA, PENNSYLVANIA

The Tumbling Run Member of the Pottsville Formation has a maximum thickness of more than 1,000 feet at its outcrop in Schuylkill County (pl. 4A), but it thins northward rather abruptly, and within 20 miles the formation is absent. Oldest Pennsylvanian strata are absent between the Southern Anthracite field and Green­brier and Nicholas Counties, W. Va., a distance of about 280 miles, probably because of erosion after folding in Permian time.

CENTRAL APPALACHIAN AREA

The area underlain by strata of the Pocahontas Formation and its equivalents in interval A extends

(Continued)

southwestward from Greenbrier and Nicholas Counties, southern West Virginia, probably to a little beyond Cumberland Gap in Claiborne County, Tenn. This area is about 180 miles from northeast to southwest and has a maximum width of 60 miles. In its outcrop area the Pocahontas Formation is readily traced, and it has been extensively prospected by drill for the valuable Pocahon­tas No. 3 coal bed. Northward, the strata dip under fairly thick cover, and the No. 3 coal bed thins and eventual­ly wedges out along a somewhat indefinite line, as deter­mined from drillers' logs of oil or gas wells. A maximum thickness of 1,090 feet is recorded in Buchanan County, Va. Erosion of upper parts of the formation in the out­crop area produced a reversal in the thickness trend in the outcrop belt. The base of the Pocahontas Formation near the southern West Virginia-Virginia line is drawn about 50 feet below the Squire Jim coal bed (table 2), the lowest Pennsylvanian coal, and just above the highest red shales of the Pennington Formation of Mississippian age.

In southeastern Kentucky the part of the Lee Forma­tion assigned to subinterval A1 is deeply buried, but its north border, in Pike County and adjoining counties, appears to be marked by rather abrupt thinning, as shown in a series of deep subsurface records.

In Tennessee the basal strata of Pennsylvanian age commonly do not contain plants of the floral zone 4 of Read. In a few localities (pl. 4A) rocks are questionably referred to subinterval A1; but the subinterval may be almost unrepresented in Tennessee.

SOUTHERN APPALACHIAN AREA

The northern limit of subinterval A1 in Alabama is not known. The thin succession of rocks capping plateau remnants north of the Tennessee River consists largely or entirely of the lower part of subinterval A2. South of the river both intervals are present in strata that dip uniformly southwestward. In the Black Warrior basin, strata tentatively referred to subinterval A1 thicken southeastward to a maximum of more than 1,700 feet near Birmingham. The Parkwood Formation crops out in the Cahaba coal field, and it and the lower part of the Pottsville Formation, assigned to subinterval A1, generally range in thickness from 3,000 to more than 5,000 feet. In the Coosa field maximum thicknesses of slightly more than 5,000 feet are assumed for subinterval A1. The Cahaba and Coosa fields are separated from the Black Warrior basin and from each other by overthrust faults, anticlines, and other structures.

The thickness of subinterval A1 in the southern and western part of the Black Warrior basin was determined by a comparison of electric logs of wells, starting in Marion County, Ala. Interval A1 appears to thicken southwest across the basin nearly to its south edge, where it is abruptly terminated. It attains thicknesses of
1,700 ± feet at locality 105 in Jefferson County, Ala., 1,780 ± feet at locality 113 in Winston County, Miss., and 1,605 ± feet at locality 159 in Sumter County, Ala.

LITHOFACIES TRENDS

ANTHRACITE FIELDS AREA

The surface and subsurface records used for differentiating subinterval A1 in the Anthracite field are based on measurements of outcrops and exposures in mine tunnels and are considered to be accurate. Sandstone exceeds mudstone in all but two of the locations studied (pl. 4B). The largest proportions of sandstone (more than 80 percent) are found toward the southwest, northwest, and eastern margins of the Southern and Western Middle fields.

In the lower part of interval A, red silty mudstones resembling those in the underlying Mauch Chunk Formation alternate repeatedly with conglomerates typical of the Pottsville. The distribution of coal beds in subinterval A1 is discussed in the chapter by H. R. Wanless in part II of this publication.

CENTRAL APPALACHIAN AREA

Subinterval A1 in the central part of the area contains 50-80 percent sandstone. The sandstone is higher in percentage in Dickenson County and adjoining counties, Virginia, and locally toward the northwest in Mingo and Boone Counties, W. Va. The interval includes more mudstone than sandstone to the east in Mercer, Summers, Greenbrier, and Fayette Counties, W. Va. The high percentage of sandstone to the northwest may result from the fact that the drillers' logs, which are the main source of information in that area, are of inferior quality. The distribution of the coals in the Pocahontas Formation is discussed in the chapter by H. R. Wanless in part II of this publication.

SOUTHERN APPALACHIAN AREA

Although subinterval A1 varies greatly in thickness in Alabama and Mississippi, its gross lithology varies only from somewhat more sandstone than mudstone to somewhat less. Basins that include thousands of feet of strata generally contain more mudstone than sandstone, as would be expected in areas that had been actively downwarped. Sandstone generally exceeds mudstone in north-central Alabama. In Mississippi much dark-gray flaky micaceous mudstone alternates with sandstone which is finer grained than that in the outcrop area in Alabama.

Possibly some of the phyllite described as Erin Shale in Clay County, Ala., southeast of the Coosa field, may belong to subinterval A1, but it has yielded only Lepidodendron sp., and its age is known no closer than "Carboniferous, probably Pennsylvanian." Interval A1 contains little minable coal in the southern Appalachians.

SOURCES AND ENVIRONMENTS OF DEPOSITION

In northeastern Pennsylvania and southern West Virginia, where nearly 1,000 feet of sediment is preserved, a local source area probably was elevated near the end of Mississippian time, but the degrees of erosion fluctuated, perhaps as a result of climatic cycles. The coarser texture of the sediment in the Pennsylvania area (Southern Anthracite field) compared with the texture of sediment deposited in West Virginia suggests that remnants of subinterval A1 in Pennsylvania were closer to the source, which lay to the south, and that the land supplying the detritus had higher relief than land furnishing detritus to southern West Virginia.

Land supplying sediment to the West Virginia-Virginia area must have lain southeast of Bluefield, W. Va., in what is now southern Virginia and southern North Carolina. Immediately south or southwest of this area, lowlands may have provided detritus for a thin succession of strata in the area to the west, although most of such strata have been removed by post-Pennsylvanian erosion.

Variations in the sandstone-mudstone ratio in the Appalachian basin probably indicate shifting courses of the alluvial channels or fans in which most of the sandstone was accumulated, as well as variations in relief and distance from source areas. The source areas were probably to the southeast, and, because of the coarseness of sediment, are estimated to have been within 50 miles of the basin of deposition.

The high proportion of sandstone in Dickenson County, Va., may mark the position of an important drainage course along which sand and some gravel were transported from a southeastern source area. The more shaly succession farther east may indicate a lower or more distant source of sediments, or it may have accumulated in an area more removed from principal channels.

In north-central Alabama, moderately coarse grained rock suggests a nearby land, probably east or southeast of present outcrops. The pattern of grain-size distribution is consistent with an easterly source for sediments, also indicated by crossbedding measurements made in Alabama, Georgia, and Tennessee (Schlee, 1963, p. 1448). Conglomerate is less plentiful and less coarse than in the Pocahontas Formation of the central Appalachian area; thus, source areas were probably lower and slopes more gradual in the Alabama-Mississippi area than farther north.

PALEOTECTONIC IMPLICATIONS

Downwarping was greatest near the north and south ends of the Appalachian region. The 1,000-foot isopach in Alabama seems to mark the transition from a shelf
area on the north to a geosyncline on the south. Absence or relative thinness of rocks of subinterval A₂ in the central Appalachian area suggests less downwarping or perhaps less elevation of sources to the east or southeast.

**SUBINTERVAL A₂**

**FORMATIONS INCLUDED**

Subinterval A₂ includes: The upper part of the Tumbling Run Member and the Schuykill Member of the Pottsville Formation of the Southern and Middle Anthracite fields, the Pottsville Formation (Olean Conglomerate Member) of southern New York, the lower part of the Pottsville Formation below the base of the Massillon Sandstone Member or Lower Connoquenessing Sandstone Member (table 2) of Ohio and northern Pennsylvania, the New River Formation of West Virginia, the upper part of the Lee Formation in Virginia and eastern Kentucky, the Crab Orchard Mountains and Gizzard Formations of Tennessee and Georgia, and the middle part of the Pottsville Formation in Alabama and Mississippi.

**STRATIGRAPHIC RELATIONS**

Subinterval A₂, which corresponds to the upper part of the Morrow Series of the midecontinent region, is best correlated regionally by its compression floras. Floral zone 5 of Read and Mamay (1964), characterized by *Marioperitis pottsvilleae* and *Ancemites* spp., occurs in the lower part of the interval at the position of the *Lykens Valley* No. 4 coal of the Southern Anthracite field, Pennsylvania, and the Quinimont Shale Member of the New River, including the *Beckley* and *Fire Creek* coal beds, of southern West Virginia. Floral zone 5 occurs near the Battle Creek coal bed in the Gizzard Formation in Tennessee and probably in the *Black Creek* coal group in the Pottsville Formation in the Warrior coal field of Alabama. It has not been reported in subinterval A₁ in southern New York, northwestern Pennsylvania, and northern Ohio, where most of subinterval A₁ is probably younger than this floral zone.

Floral zone 6, characterized by *Marioperitis pygmaea*, *Neuropeters tennesseana*, and other plant compressions, is in *Lykens Valley* coal beds No. 2 and No. 3 of the type Pottsville section, Southern Anthracite field, Pennsylvania, and in the *Sharon* coal of Pennsylvania and northern Ohio. The *Jackson Shaft* coal of southern Ohio also contains this flora. In the middle Appalachian area floral zone 6 is well displayed at the positions of the *Sewell* coal of West Virginia. It is also known in eastern Kentucky, as at Cumberland Falls. In the southern Appalachian area this flora is found just above the *Secomee* coal on Lookout Mountain, Ga., and was reported above the *Mary Lee* coal of the Warrior field by White (in Butts, 1927, p. 15). The *Wadsworth* coal of the Cahaba field, which was correlated by Butts (1927), has yielded this flora.

**UPPER BOUNDARY OF SUBINTERVAL A₂**

Throughout much of the Appalachian region, the boundary between subinterval A₂ and interval B is drawn on lithologic, rather than paleontologic, grounds. The upper parts of the New River Formation and the Lee Formation of West Virginia and Virginia and the Crab Orchard Mountains Formation of Tennessee include generally thick massive coarse-grained pebbly highly permeable sandstones which contrast with finer grained, less permeable sandstones in overlying strata. The highest very permeable sandstones are the Nuttall Sandstone Bed of the Sewell Member of the New River of West Virginia, the Bee Rock Sandstone Member of the Lee of Virginia, the Corbin Sandstone Member of the Lee of southeastern Kentucky, and the *Rockcastle Conglomerate Member* of the Crab Orchard Mountains Formation of Tennessee. Although the upper boundary of subinterval A₂ is placed some distance above strata yielding fossil plants of floral zone 6, distinctive floras are not common near the boundary just described.

In northeasternmost Kentucky only, the upper part of the Lee is excluded from subinterval A₂; it is mapped with interval B because it overlies the *Oliver Hill Clay Bed*, which probably correlates with the *Sciotoville Clay Member* at the base of interval B in southern Ohio. Because the *Oliver Hill Clay Bed* disappears southward beneath the massive sandstones referred to the upper part of the Lee, a considerable area in northeasternmost Kentucky is mapped as having no strata of subinterval A₂. The sandstone of the Lee Formation overlying the clay in Kentucky may be a correlative of the *Massillon Sandstone Member* of northern Ohio and the *Lower Connoquenessing Sandstone Member* of northwestern Pennsylvania, which are considered to be the basal beds of interval B in those areas.

In the Alabama coal basins the sandstones of subinterval A₂ are less easily distinguished from those overlying them than they are in Tennessee and northward. Fossil plants suggest that the *Brookwood coal group*, the highest outcropping strata of the Warrior coal field, belongs to subinterval A₂. In the subsurface Black Warrior basin of western Alabama and Mississippi the position of the *Brookwood coal group* was traced on electric logs, and higher strata were referred to interval B. Small parts of the thick sequences in the Cahaba and Coosa fields are assigned to interval B. Tentatively, the base of the *Yeshic* coal of the Cahaba field and that of the *Brewer* coal of the Coosa field are used as A₂-B boundaries. The stratigraphic position of the Erin Shale (phyllite), of probable Early Pennsylvania-
APPALACHIAN REGION

nian age, in Clay County, Ala., has not been determined, but these metamorphic rocks may include strata of sub-interval A₂.

THICKNESS TRENDS

In a large part of the Appalachian region the oldest Pennsylvanian rocks belong in subinterval A₂ (pl. 5A). Their thickness trends have been described under interval A. Subinterval A₂ overlies A₁ only in the Southern and Middle Anthracite fields, Pennsylvania; in an area in southern West Virginia, Virginia, and southeastern Kentucky that probably extends a short distance into Tennessee; and in a large area in Alabama and Mississippi.

ANTHRACITE FIELDS AREA

The Schuykill Member of the Pottsville in outcrops near the north border of the field is about 100 feet thick. It thickens southward and southwestward to about 300 feet near the east end of the Southern Anthracite field, and to nearly 700 feet in the “swallowtail” bifurcation of the southwestern part of the field.

CENTRAL APPALACHIAN AREA

Subinterval A₂, which comprises the New River Formation in West Virginia and the upper part of the Lee Formation of Kentucky and Tennessee thickens to the south from about 450 feet in central West Virginia. The proportions of the total thickness of the Lee Formation in southeastern Kentucky and the western counties of Virginia (pl. 13) that should be assigned to subintervals A₁ and A₂ are somewhat questionable.

SOUTHERN APPALACHIAN AREA

Subinterval A₂ in its outcrop belt in northern Alabama and Mississippi thickens southward from a thin edge to 500 feet within a distance of 10-30 miles. The greatest southward rate of thickening is in Chickasaw County, Miss., between localities 88 and 90 (pl. 1), where subinterval A₁ increases from a thin edge to 2,850 feet in a distance of about 8 miles. In Walker and Tuscaloosa Counties, Ala., the subinterval thickens in a distance of 20 miles from 502 feet at locality 81 to 2,120 feet at locality 158.

In part of this area the subinterval is beveled by post-Pennsylvanian erosion; however, the original thickness of the deposits clearly was greatest toward the south. Southward thickening for all the Pennsylvanian rocks in northern Alabama is shown by Metzger (1965). The thickest complete section recorded for subinterval A₂ is 3,740 feet at locality 107, Clay County, Miss. Southward from the narrow belt of rapid thickness changes, subinterval A₂ seems to be very uniform in thickness, although control is sparse and is confined to localities 132 and 159 in Pickens and Sumter Counties, Ala., respectively.

In the Cahaba coal field the strata reach a maximum thickness of 2,760 feet at locality 131, eastern Jefferson County, Ala. In the Coosa coal field, subinterval A₂ has a maximum thickness of more than 2,500 feet at locality 120 in Shelby County, Ala. In both these basins an abrupt increase in thickness is largely the result of the addition of higher beds of interval A₂ in a southeastern direction.

LITHOFACIES TRENDS

Features of the lithofacies map (pl. 5B) described here are limited to the areas in which subinterval A₂ is underlain by subinterval A₁. Elsewhere A₂ is equivalent to A₁, and the lithofacies have been described previously.

ANTHRACITE FIELDS AREA

In all localities in the Middle and Southern Anthracite fields, subinterval A₂ is composed of more than 80 percent sandstone and conglomerate. The sandstones generally are very pebbly.

CENTRAL APPALACHIAN AREA

Subinterval A₂ contains mostly mudstone and sandstone in the central Appalachian region. Sandstone makes up more than 80 percent of the subinterval in the northwestern part, as in Kanawha and Boone Counties, W. Va., and mudstone is predominant to the southeast, as in Mercer and McDowell Counties, W. Va. This distribution of mudstone and sandstone seems anomalous, because crossbedding studies show that much of the sediment was derived from lands to the southeast.

The highest percentages of sandstone correspond to areas where the subinterval is thin. Apparently, mud was deposited in some deeply downwarped areas and bypassed intervening areas. This feature is illustrated by such localities as those in McDowell and Nicholas Counties, W. Va., where subinterval A₂ contains more mudstone than sandstone, that are separated by such areas as those in Raleigh County, W. Va., and Dickenson County, Va., where the interval contains more sandstone.

SOUTHERN APPALACHIAN AREA

Subinterval A₂ generally contains a higher percentage of mudstone in the southern Appalachian area than it does farther north. More than 80 percent sandstone and conglomerate is found locally in the southeastern part of the Cahaba field and in the thin northernmost outcrop areas south of the Tennessee River in Franklin, Lawrence, Morgan, and Marshall Counties, Ala. Subinterval A₂ is thinner and has more sandstone in northeastern Mississippi than in areas farther south, where the subinterval is thick and dark flaky silty mudstone is predominant.

---

3Edits' note: Recent mapping (Welch, 1971) shows a normal fault of large displacement between localities 88 and 90, which apparently accounts for the abrupt thickening.

4According to W. C. Colburn, who described this core, subinterval A₁ should be about 1,320 feet thick and A₂ about 1,600 feet thick at this locality.
SOURCES AND ENVIRONMENTS OF DEPOSITION

SOURCES

Throughout the Appalachian region subinterval \( A_2 \) contains a substantial amount of crossbedded sandstone. Inclinations of the crossbeds have been studied systematically for information on current direction and direction to the sources of sediment. In northern Ohio (Fuller, 1955) and in northern Pennsylvania and southern New York (Meckel, 1967), sources for the sandstones and conglomerates of the Sharon and Olean lay to the north in northern New York and eastern Canada. This interpretation is supported by the presence of fragments of Middle Devonian fossils, such as occur in the Onondaga and Hamilton Formations of New York (Fuller, 1955, p. 171-172). Areas to the east and southeast were the sources of the conspicuous pebbly sandstones and conglomerates of the Anthracite fields (Wood and others, 1963; Meckel, 1967). These pebbles are probably the largest of this age in the Appalachian area and were derived from source areas about 50 miles to the southeast, near the present site of Philadelphia.

Farther south in the Appalachian region, sources of pebbly or crossbedded sandstones of subinterval \( A_2 \) in West Virginia, eastern Kentucky, Virginia, Tennessee, Georgia, and Alabama were probably to the southeast, perhaps in part in areas now occupied by the Blue Ridge and Piedmont of Virginia, the Carolinas, and Georgia. An easterly source in the northern part of this belt is evidenced by regional studies by Mitchum (1954), and by the presence of pebbles of bluish vein quartz in some conglomerates, resembling vein quartz in the metamorphic terrain of the Virginia Piedmont. The source areas to the east and southeast extended from near Philadelphia, Pa., southeastward into western Georgia, and were probably uplands of varying height, drained by major streams, or they may have been a series of mountainous island "wells" like those which now flank the Pacific basin. The areas where the percentage of sandstone is high may occupy the positions of former streams from highlands to the southeast, or they may have been opposite the mountainous islands of an archipelago, such as that proposed by Kay (1952) and Eardley (1947). The presence of a coarse conglomerate in the southern part of the Cahaba coal field, Alabama (Culbertson, 1964), suggests that nearby highlands related to the Ouachita uplift were an important source of sediments for central Alabama during subinterval \( A_2 \).

ENVIRONMENTS OF DEPOSITION

In addition to the conglomerate, sandstone, and mudstone mentioned previously, subinterval \( A_2 \) in the Appalachian region includes numerous coal beds and underclays and a few iron ore and limestone beds. Numerous cyclic alternations occur in most parts of the Appalachian region; therefore, the predominant sandstone and mudstone units do not generally form thick units, but alternate with and are separated by coal beds, underclays, and other types of sediment.

Table 3 shows the frequencies of various lithologies in representative stratigraphic sections for subinterval \( A_2 \). Marine environments are indicated by marine faunas in the Sharon iron ore of Ohio (Morningstar, 1922), the Skelton Shale in the New River Formation of southern West Virginia, a basal Pennsylvanian zone near Bon Air, Tenn. (Wanless, 1946, p. 22), and numerous zones in the Warrior and Cahaba fields, Alabama (Culbertson, 1964, p. B20, B36). Probably a large part of the thick succession of rocks in subinterval \( A_2 \) in Mississippi is marine. Nonmarine, probably lacustrine or lagoonal, faunas are found in the Hartridge Shale in the New River Formation, southern West Virginia (Reger, 1931a, p. 225).

The conglomerates, sandstones, and mudstones form a complex of detrital wedges. In some areas sand and mud accumulated without appreciable interruption, but elsewhere deposition of detritus ceased periodically, and coal beds were deposited. Most coal beds of subinterval \( A_2 \) are very lenticular, but some are very widespread, as illustrated by the Fire Creek, Beckley, and Sewell of southern West Virginia; the Battle Creek and Seneca of Tennessee; the Black Creek, Mary Lee, and Pratt of the Warrior field, Alabama; and the Wadsworth, Thompson, and Helena of the Cahaba field, Alabama. Almost all sedimentary rocks of subinterval \( A_2 \) in Tennessee, except coal beds and underclays, have been interpreted as marine (Wilson and Stearns, 1960). A marine origin for these rocks seems unlikely, however, as the patterns of distribution of mudstone and sandstone resemble those of fans, channels, flood plains, and deltas in other areas.

The sea is believed to have transgressed periodically from the southwest through Mississippi. Rocks in West Virginia and Virginia show less evidence of marine deposition than those farther southwest. It is not known whether the seaway in which the Sharon iron ore of Ohio was formed connected with marine waters to the southwest, or whether it extended from the west or northwest at the time that the Sellers Limestone Member (table 11) of the Caseyville Formation formed in southern Illinois or limestone in the Saginaw Formation formed in Michigan.

Cyclic sedimentation is plainly recognizable in subinterval \( A_2 \), but the cyclothemicsuccessions are less uniform regionally than those of higher intervals.

PALEOTECTONIC IMPLICATIONS

Coarse sediment entered many parts of the Appalachian region at one or several times during the deposition of subinterval \( A_2 \). Indeed, except in Alabama...
and Mississippi, the coarsest sediments of the entire sequence of Pennsylvanian rocks in the region are in this interval and indicate the highest energy conditions. This active erosion and sedimentation resulted from contemporar y tectonism in various parts of the Appalachian positive element. Whether the alternations between coarse pebbly sandstones and fire clays and coal beds resulted from intermittent renewal of tectonism, as suggested by J. M. Weller (1956), or were controlled by climatic changes, as proposed by Wanless and Shepard (1936) and Beerbower (1961), is not unequivocally proved. Probably both factors influenced sedimentary successions of this interval.

**INTERVAL B FORMATIONS INCLUDED**

In the Anthracite fields of Pennsylvania the upper part of the typical Potsville has been designated the Sharp Mountain Member and is here referred to interval B (Atoka age; pl. 13). Although this assignment is debatable in several grounds, the thickness of strata involved is small, and the Sharp Mountain Member is a part of the type Potsville. On the other hand, plant-compression floras from shaly beds of the predominantly conglomeratic sandstone succession of the Sharp Mountain Member suggest that these strata may correspond to rocks of early Allegheny or Des Moines age (interval C) in western Pennsylvania (C. B. Read, written commun., 1938), and, therefore, rocks of Atoka age may not be represented in the Southern and Middle Anthracite fields.1

In the bituminous coal fields of Pennsylvania, Ohio, and Maryland, including outliers in northern Pennsylvania and the Broad Top and Georges Creek synclines (fig. 8), interval B consists of the Potsville Formation, except where the Sharon Sandstone or Conglomerate Member and associated strata are present; the Sharon is assigned to interval A. In large parts of Pennsylvania and Ohio, the initial Pennsylvanian sediments were post-Sharon in age and belong to interval B. In the southern part of West Virginia the Pottsville, here designated as a group, thickens greatly; the Kanawha Formation represents the upper part of the Pottsville and makes up interval B. In the Kanawha River valley, strata at the level of the Coalburg and Stockton coals are included in the Kanawha Formation but are believed to be equivalent to the lower part of the Allegheny strata of Pennsylvania and are placed in interval C. In eastern Kentucky, north of Pine Mountain, interval B is composed chiefly of the Breathitt Formation, but in the northeasternmost counties, the interval also includes the Olive Hill Clay Bed at the base of interval B and the overlying strata, both of which are referred to the Lee Formation. The Breathitt Formation contains some strata of Des Moines or Allegheny age in its uppermost part, upward from the Francis or Hindman coal.

Southeast of the Pine Mountain thrust fault (fig. 7) in Kentucky, strata above the Lee Formation are, in ascending order, the Hance, Mingo, Catron, and Hignite Formations of the Breathitt Group. These are referred to interval B. In the Virginia part of the Middleboro syncline southeast of the Pine Mountain thrust fault, the Norton, Gladeville, and Wise Formations represent a greatly thickened interval B. In Tennessee interval B contains, in ascending order, the Crooked Fork, Slatestone, Indian Bluff, Graves Gap, Red Oak Mountain, Vowell Mountain, and Cross Mountain Groups (Wilson and others, 1956). The Crooked Fork Group is believed to include the equivalent of the Corbin Sandstone Member of the Lee Formation (table 2), which is the top of interval A in Kentucky, and younger strata up to the Wartburg Sandstone, equivalent to the lower Breathitt Formation of Kentucky.

Interval B has been entirely eroded from Georgia. In Alabama, strata of the Pottsville Formation including and younger than the Yeshic coal of the Cahaba coal field and the Breuer coal of the Coosa coal field are referred to interval B on the basis of tentative cor-

---

1Editor's note: According to Gordon Wood, U.S. Geological Survey, current evidence suggests that in the Anthracite area the Sharp Mountain is equivalent in age to rocks elsewhere that are in the lower part of interval C. Interval B should be considered as missing, or possibly interval B should include strata of uppermost Schuylkill and lowermost Sharp Mountain Members.

### Table 3.—Number of repetitions, by area, of lithologic types in subinterval $A_2$ of Appalachian region

<table>
<thead>
<tr>
<th></th>
<th>1 Ohio and Pennsylvania</th>
<th>2 Southern West Virginia</th>
<th>3 Big Stone Gap, Va.</th>
<th>4 Stearns and Cumberland Falls, Ky.</th>
<th>5 Graysville, Tenn.</th>
<th>6 Warrior field, Alabama</th>
<th>7 Cahaba field, Alabama</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mudstone</td>
<td>3</td>
<td>20</td>
<td>17</td>
<td>16</td>
<td>17</td>
<td>22</td>
<td>34</td>
</tr>
<tr>
<td>Underclay</td>
<td>1</td>
<td>5</td>
<td>6</td>
<td>4</td>
<td>9</td>
<td>?</td>
<td>?</td>
</tr>
<tr>
<td>Sandstone</td>
<td>2</td>
<td>13</td>
<td>11</td>
<td>10</td>
<td>12</td>
<td>20</td>
<td>34</td>
</tr>
<tr>
<td>Conglomerate</td>
<td>2</td>
<td>5</td>
<td>6</td>
<td>3</td>
<td>2</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Coal beds</td>
<td>1</td>
<td>17</td>
<td>10</td>
<td>4</td>
<td>11</td>
<td>19</td>
<td>12</td>
</tr>
<tr>
<td>Limestone</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>1</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>Iron ore</td>
<td>2</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>
relations by Butts (1927, p. 15). Strata referred to interval B in the subsurface of the Black Warrior basin in western Alabama and Mississippi have not been classified or much studied. They have generally been called Pottsville Formation, or merely Pennsylvanian. Palynological studies show that beds of Kanawha age are included in the Pottsville in the subsurface of Mississippi (Upshaw, 1967, p. 16), and these beds are accordingly assigned to interval B.

**STRATIGRAPHIC RELATIONS**

Interval B includes floral zones 7 and 8 of Read and Mamay (1964). Zone 7 is characterized by *Megaleopteris* spp., which are found in the basal part of interval B. Zone 8 is characterized by *Neuropteris tenifolia*, distributed widely through the major part of the interval. In large parts of the United States, marine strata of this interval are characterized by the foraminiferal zones of *Profusulinella* in the lower part and of *Fusulinella* in the upper part (Thompson, 1960, p. 111). *Profusulinella* has not been reported from the Appalachian coal field, but rocks of the lower part of interval B can be identified by their fossil floras.

*Fusulinella*, as represented by *F. iowensis* and *F. iowensis* var. *stouti*, occurs in Ohio in the Boggs and Lower and Upper Mercer Limestone Members of the Pottsville (Thompson, 1936; Smyth, 1957). These forms are especially useful for interregional correlation, as they are also found in the Seville Limestone Member of Spoon Formation and Curlew Limestone Member of the Tradewater Formation of the Illinois basin and in the Verde Limestone Member of the Saginaw Formation of the Michigan basin. The upper limit of interval B in the Southwest and in the Rocky Mountain States is placed just below strata yielding the most primitive *Fusulinina* and *Wedekindellina maturna*. These forms are not known in the Appalachian region, and, presumably, their stratigraphic position is a little lower than that of *Fusulinella iowensis* (R. Hollingworth, written commun., 1963). In the Appalachian region the upper boundary of interval B is placed a little above the Upper and Lower Mercer Limestone Members, which yield *Fusulinella iowensis*.

Most of the fusulinid localities known in the Appalachian coal field are in Ohio, but, by the use of floral zones and by tracing widespread coals, several correlations can be established with fair confidence with strata in Pennsylvania, Maryland, West Virginia, and eastern Kentucky.

Studies of small spores isolated from coals have been useful for correlation. Cropp (1963) studied spore succession in representative coals throughout the column in Tennessee. David White (quoted by Glenn, 1925) reported that floras near the very top of the Pennsylvania of Tennessee in Cross Mountain, Anderson County (fig. 6), are equivalent to those in the Mercer coals of Ohio and Pennsylvania. Thus, interval B includes the youngest Pennsylvanian strata in Tennessee.

A boundary was tentatively drawn by White (quoted by Butts, 1927, p. 14) between middle and upper Pottsville strata in the Cahaba coal field of Alabama, and rocks containing his "Upper Pottsville" floras are referred to interval B. Outcropping strata in the Black Warrior basin of Alabama do not include floras younger than the "middle Pottsville" according to White (quoted by Butts, 1927, p. 15).

In the subsurface Warrior field of western Alabama and Mississippi, a tentative tracing of units was made from electric-log patterns of resistivity and self-potential curves from the youngest strata outcropping in Tuscaloosa County, Ala. It was assumed that the Brookwood coal group, the youngest outcropping strata of the basin, belonged at or near the top of interval A. Younger unexposed strata, including the youngest Pennsylvanian rocks present, were referred to interval B.

**UPPER BOUNDARY OF INTERVAL B**

The upper boundary of interval B in the northern Appalachian area is placed at the base of the Brookville coal or at the top of its underlying clay or sandstone, called at some places the Homewood Sandstone Member of the Pottsville (table 4). Because *Fusulinella iowensis* is found a little below this position in the Upper Mercer Limestone Member and because *Fusulina leei* occurs in the Putnam Hill Limestone Member of the Allegheny Formation, which is the roof of the Brookville coal of Ohio (Smyth, 1957, p. 267), the boundary in Ohio corresponds closely to that used in the Illinois and Michigan basins. The position of the *Brookville coal*, however, is traced with much less confidence in Pennsylvania, Maryland, West Virginia, and eastern Kentucky, where fusulinids do not occur in marine strata. In all these areas strata near the *Brookville coal* have less continuity than in Ohio. In Pennsylvania, the *Brookville coal* is not confidently identified even a few miles away from the type outcrop near Brookville.

It has been proposed that the Pottsville-Allegheny boundary in the bituminous coal field of Pennsylvania should be raised from the *Brookville coal* to the *Lower Kittanning coal* (Ashley, 1945; Cheney and others, 1945, p. 147) because (1) readily identifiable physical evidence for the present boundary is lacking at many places; (2) the top of the Pottsville at its type section in the Southern Anthracite field is believed to be equivalent of the *Lower Kittanning coal*; and (3) no pronounced floral break occurs near the *Brookville coal*. For these reasons, the boundary between intervals B and C may not be placed consistently in all parts of Pennsylvania, Virginia, and eastern Kentucky.
Table 4.—Stratigraphic units mentioned in text for interval B but not shown on plate 13: Appalachian region

<table>
<thead>
<tr>
<th>Interval</th>
<th>Western Pennsylvania</th>
<th>Ohio</th>
<th>West Virginia</th>
<th>Virginia</th>
<th>Eastern Kentucky</th>
<th>Tennessee</th>
<th>Alabama</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>Allegheny Formation Lower Kittanning coal</td>
<td>Allegheny Formation Brookville coal</td>
<td>Putnam Hill Limestone Member</td>
<td>Brookville coal</td>
<td>Coalburg and Stockton coals</td>
<td></td>
<td>Francis or Hindman coal</td>
</tr>
<tr>
<td></td>
<td>Homewood Sandstone Member</td>
<td>Homewood Sandstone Member</td>
<td>Tionesta coal</td>
<td>Tionesta coal</td>
<td>Upper Mercer Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Upper Mercer Limestone Member</td>
<td>Upper Mercer Limestone Member</td>
<td>Bedord coal</td>
<td></td>
<td>Lower Mercer Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mercer coal</td>
<td>Mercer coal</td>
<td>Boggs Limestone Member</td>
<td></td>
<td>Quakertown coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower Connoquenessing Sandstone Member</td>
<td>Massillon Sandstone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>Pottsville Formation Quakertown coal</td>
<td>Pottsville Formation Quakertown coal</td>
<td>Massillon Sandstone Member</td>
<td></td>
<td>Cedar Grove coal</td>
<td></td>
<td>Phillips coal</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Campbell Creek coal</td>
<td>Chilton coal</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Eagle Shale Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Lower Connoquenessing Sandstone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A</td>
<td>Sharon Sandstone Member</td>
<td>Sharon Sandstone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sciotoville Clay Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1 Upper part of Lee Formation placed in interval B in northeasternmost Kentucky only.

The study of spore floras from successive coal beds near the Pottsville-Allegheny boundary in eastern Kentucky by Kosanke (1974) confirmed that the Princess No. 6 coal, Kentucky, correlates with the Lower Kittanning coal. An important change in the spore-pollen assemblage of the Princess No. 6 coal is recognized.

In Tennessee, Mississippi, and Alabama the top of interval B is at the contact with unconformably overlying Mesozoic rocks.

**THICKNESS TRENDS**

Interval B ranges in thickness from less than 50 feet to a few localities in southeastern Ohio to 7,640 feet at locality 113 in Winston County, Miss. (pl. 64). For convenience in describing thickness trends, the Appalachian region is divided into the northern, central, and southern Appalachian areas.

**NORTHERN APPALACHIAN AREA**

The northern part of the Appalachian coal field includes, in addition to the main bituminous coal field, the Anthracite fields and other outliers in northern Pennsylvania, the Broad Top coal field in southern Pennsylvania, and the Georges Creek syncline in southern Pennsylvania, Maryland, and eastern West Virginia. In the northern Appalachian area as thus defined, thickness of interval B ranges from less than 50 feet in northern Ohio and northern Pennsylvania to 1,287 feet near lat 38° N., in Fayette County, W. Va.

Interval B had as its depositional centers two basinlike areas, the first in northern Ohio and Pennsylvania, and the second in southern Ohio and West Virginia. A broad belt between these basins received no sediment during the early part of interval B. By the end of interval B, however, the two depositional basins had joined. In a belt extending eastward from localities 268 and 270 in southern Muskingum County, Ohio (fig. 5), to locality 408, northeast of Pittsburgh, Allegheny County, Pa., interval B is locally more than 100 feet thinner than it is in areas both to the north and south. The thinning reflects topographic relief on the unconformably underlyng Mississippian rocks.

Interval B in the Anthracite fields ranges from less
than 100 to about 300 feet in thickness, increasing southward (pl. 6A). In the Broad Top field the interval is as much as 147 feet thick and rests directly on pre-Pennsylvanian rocks. In the Georges Creek syncline, interval B ranges from less than 100 to more than 400 feet and thickens southwestward.

Local variations in thickness in the northern Appalachian area are due to (1) topographic irregularities on the pre-Pennsylvanian surface; (2) differential compaction of sediments; (3) uncertainty regarding the exact upper or lower boundary of the interval, as where the Lower Conococheessing Sandstone Member of the Pottsville (interval B) of western Pennsylvania directly overlies the Burgoo Sandstone of the Pocono Group (Mississippian); (4) poor quality of records in drillers' logs, especially in the deeper part of the basin where interval B is more than 1,000 feet below the surface; and (5) greater downwarping of the basin to the south during the deposition of interval B.

CENTRAL APPALACHIAN AREA

The thickness of interval B in the central Appalachian area ranges from about 100 feet in northern Kentucky, where only the lower part is present, to about 3,200 feet in Big Black Mountain (fig. 6) on the Kentucky-Virginia line near Big Stone Gap, Va. The average thickness in northeastern Kentucky, where interval B is overlain by interval C, is about 300 feet. The maximum thickness in West Virginia is 1,765 feet at locality 590 in Raleigh County and in Tennessee it is 2,824 feet at locality 41 on Cross Mountain in Anderson County.

Abrupt thickening of interval B in the vicinity of the Paint Creek-Warfield fault zone of West Virginia and Kentucky and near the Pine Mountain thrust fault in Virginia, Kentucky, and Tennessee has been described by Wanless (1946, p. 4). The thickness of the interval ranges from about 300 to 600 feet north of the Paint Creek fault zone, from 500 to 2,400 feet between the Paint Creek and Pine Mountain faults in Kentucky, and from 2,500 to 3,200 feet south of the Pine Mountain thrust.

SOUTHERN APPALACHIAN AREA

Correlation of strata assigned to interval B in the Coosa and Cahaba coal fields in Alabama and in the Black Warrior basin in Mississippi is uncertain. Moreover, it is not certain that these rocks correspond closely to those in interval B of the northern and central Appalachian areas. Presumably, most of the strata in the Coosa coal field belong to interval A, but in the southern part of the field, at Alabama locality 120, a thickness of 380 feet is referred to interval B. A thick sequence of strata outcropping in the Wattsville basin in the northern part of the Coosa field is interpreted as belonging entirely in interval A. In the southern part of the Cahaba field at locality 124, a maximum of 2,060 feet is tentatively assigned to interval B.

In the Black Warrior basin, interval B seems to have been eroded from the area north of Monroe, Clay, and Chickasaw Counties, Miss. If electric-log correlations are correct, interval B is 3,800 feet thick within about 7 miles of its northern limit at locality 103, Clay County, Miss., and is thicker than 1,500 feet in nearly all well records where the interval is present. A well in Kemper County, Miss. (locality 119), shows interval B to be 6,140 feet thick, and one in Winston County shows it to be 7,640 + feet thick. Spores in samples from the well at locality 103, Clay County, show correlatives of rocks assigned here to interval B (Cropp, 1960).

LITHOFACIES TRENDS

Interval B in the Appalachian region consists principally of sandstone and mudstone (pl. 6B). The interval contains 50-80 percent sandstone along the southeast border of the coal field in West Virginia and Virginia, about 20-40 percent along the northwest border in Kentucky and Tennessee, and less than 20 percent in much of its extent in north-central Mississippi. At no place do coal, limestone, and iron deposits compose as much as 20 percent of the total. These rock types are present, however, and the coals are mined extensively in part of the region. Interval B contains very little conglomerate except in the Anthracite fields of Pennsylvania and in the Cahaba field of Alabama.

The lithofacies map (pl. 6B) shows many variations in lithology within the Appalachian region and reflects the numerous factors that affected the accumulation of this complex succession of strata. General trends in variation as between sandstone and mudstone are not apparent on most parts of the map, however, because interval B in vertical sequence includes a large number of more or less widespread rock units, each with its own pattern of variation.

Table 5 shows the number of separate sandstones, mudstones, underclays, coals, and limestones at representative localities in 10 areas and clearly indicates the extraordinary lithologic complexity of interval B in the Appalachian region.

The sandstones of interval B are generally less conglomeratic, finer grained, and less quartzose than those of interval A, except in the Anthracite and southern Cahaba fields. Commonly, interval B sandstone contains interstitial clay, mica, and rock fragments, although some of the thick massive sandstones in Virginia and southern West Virginia and the Massillon and Lower Conococheessing Sandstone Members of the Pottsville of northern Ohio and Pennsylvania are nearly as quartzose as those of interval A. Some sandstone in interval B contains locally derived pebbles of mudstone or ironstone.

Interval B includes many gray mudstones containing rich compression floras, commonly the roof beds of coals; regularly bedded light-gray siltstones and
mudstones; dark-gray to black mudstones, like the Eagle Shale Member of the Kanawha of West Virginia; calcareous fossiliferous mudstones, like the Kendrick Shale Member of the Breathitt Formation of Kentucky; and valuable refractory underclays like the Sciotoville Clay Member of southern Ohio and the Olive Hill Clay Bed of northeastern Kentucky.

Coal beds are infrequent and very thin in interval B in the Anthracite field and in the Pittsburgh area, Pennsylvania. The three coal beds near Pittsburgh (table 5) have a total thickness of 8 inches. Several widely traced coal beds, such as the Quakertown, Mercer, Bedford, and Tionesta, occur in both Ohio and northern Pennsylvania, but thickness of each bed averages less than 1 foot.

Where interval B thickens abruptly south of the Paint Creek-Warfield fault trend, the number of coal beds increases to 35 in West Virginia and 29 in Kentucky and includes such extensively mined coals as the Eagle, Campbell Creek, Cedar Grove, and Chilton of West Virginia and the Jellico, Blue Gem, Elkhorn, Fire Clay, and Hazard of Kentucky. These individual coal beds may be split into two or several benches by clay, mudstone, or sandstone partings that are not very persistent. One extraordinary parting in the Chilton coal of West Virginia, the Fire Clay coal of Kentucky, the Phillips coal of Virginia, and the Windrock coal of Tennessee is composed of dark-gray flint fire clay 4-6 inches thick (Wanless, 1952, p. 170-172). This is the best marker bed for regional correlation in interval B in the central Appalachian area.

Coal beds are numerous and extensively split in Virginia and Tennessee. In table 5 the number of beds for the Cahaba field, Alabama, would be more than doubled by showing all of the separate benches. The 14 coal beds shown are generally thin and interbedded with coarse sandstone and conglomerate. In the electric log of the Kemper County, Miss., well (locality 119), several thin zones of very low resistivity suggest underclays. Probably coal beds are also present, but none is thick enough to show the high resistivity of thicker coals elsewhere. In Clay County, Miss., 40 coals were recognized by Cropp (1960) in a sample set studied for microspores.

In interval B limestones are uncommon and thin. Only one 8-inch limestone occurs in the 3,500-foot section of interval B in Big Black Mountain. The most widespread is the Lower Mercer Limestone Member of the Pottsville Formation of Ohio and Pennsylvania, which is correlated with the Magoffin Limestone Member of the Breathitt Formation of Kentucky, the Winnifrede Limestone Member of the Kanawha Formation of West Virginia, and the fossil limestone of Virginia. Several calcareous fossiliferous mudstones, especially in eastern Kentucky, include very large sporadic limestone concretions, locally septarian, containing a few marine fossils. Limestones are generally argillaceous and grade laterally into calcareous mudstones.

Interval B contains a few persistent named beds of ironstone or “iron ore,” particularly in Ohio where they were quarried for iron during the 19th century. Also, ironstone nodules occur in dark mudstones that are not counted separately in table 5.

A thin bed of black chert occurs locally at the position of the Upper Mercer Limestone Member in southern Ohio.

**SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES**

During interval B time the Appalachian and Black Warrior basins were bordered on the southeast by tec-
tonically formed highlands that probably extended from the vicinity of Philadelphia, Pa., to Georgia. These highlands must have been the principal sources of detrital sediments. They evidently did not constitute a continuous upland like the traditional land of Appalachia, but, rather, a series of uplands and lowlands, as postulated by Kay (1951, p. 32-33) for an Appalachian eugeosyncline. Areas nearest the higher lands or islands received more and coarser sediments, including conspicuous metamorphic rock fragments and basic minerals. Sources consisting of ancient sedimentary rocks north of Pennsylvania and Ohio probably were resonsible for the quartzose sandstone of the Lower Conocoquenessing and Massillon (Meckel, 1964).

The coarsest detritus accumulations of interval B in the Appalachian region probably were from sources near Philadelphia, Pa., where pebbles as much as 5 inches in diameter are not rare, and from south of the Cahaba coal field in central Alabama (Culbertson, 1964, p. B60). Furthermore, the Southern Anthracite field in Pennsylvania and the southern Cahaba field, Alabama, are the only areas listed in table 5 that show a large proportion of conglomerates. The unique conglomerates of the Cahaba field were described by Butts (1940).

Other highlands probably lay southeast of southern West Virginia and western Virginia, as shown by the large number of thick sandstones in interval B in those areas (Cavorac and others, 1963). At least 20 of the 53 sandstones in interval B at Big Black Mountain, Va. and Ky. (table 5) are more than 50 feet thick each, but only 2 of the 31 sandstones at Frozenhead Mountain, Tenn., are as thick, although both localities are near possible sources of sediment at the southeastern margin of the Appalachian coal field and are only 100 miles apart. It may be concluded that source highlands southeast of Virginia were nearer, higher, or yielded coarser detritus than those east of Tennessee.

In the Black Warrior basin, Mississippi, sandstone generally does not occur in the basal 1,800 feet of the interval. Evidently this area was relatively distant from sources of sandy sediment but received a large amount of very fine detritus.

**Environments of Deposition**

All parts of the Appalachian region—except a belt across central Ohio, central and southern Pennsylvania, and northern West Virginia—had accumulated some Pennsylvanian sediments during interval A, and pre-Pennsylvanian topographic irregularities had been eliminated. In most of the region the controlling pattern of sedimentation during interval B was cyclic and involved alternate dominance of detrital influx and of marine transgression during times of reduced erosion of source areas. This cyclic deposition was responsible for the frequently repeated sequence of sandstone, underyclay or "seat rock," coal, and mudstone, with or without evidence of marine inundation.

Interval B in the Appalachian region consists largely of fluvial and deltaic deposits that accumulated on a surface of very low relief near sea level. A shallow sea at times inundated large parts of the region. At other times most areas were covered by stagnant swamps in which large amounts of vegetation accumulated, now preserved as coal.

Cyclic sequences, although recognizable in vertical successions at most localities, cannot be traced laterally as far or as easily as in the Illinois basin of the northern midcontinent region, apparently owing to two major causes: (1) the Appalachian region was much nearer sources of detrital sediment than were the areas farther west, and sequences were subject to interruption by detrital wedges of fluvialite or deltaic sediment at many different times and places; and (2) differential downwarp of adjacent areas produced different contemporary environments in nearby localities.

Sediment accumulation kept roughly in balance with subsidence. The downwarped areas accumulated thicker sequences of more rapidly deposited sediment than did some adjacent areas which stood topographically higher and on which sediment was not deposited for appreciable periods. As proof of rapid sedimentation, fossil casts of tree trunks, in erect position and as much as 15 feet high, have been observed in southern West Virginia and eastern Kentucky.

A second evidence of rapid sediment accumulation near the borders of deposition is found in the coal beds. In the Illinois basin and in Ohio, most coal beds do not contain detrital partings, but in Kentucky, southern West Virginia, Virginia, and Tennessee, most coal beds are divided by one or more beds of sandstone, mudstone, or underclay. Interbedded detritus commonly contains compressions of roots, stems, and foliage of contemporary plants. The very common splitting of coal beds suggests that many of the coal swamps were somewhat local and that most parts were sufficiently near river channels to receive fine-grained detrital sediment washed in by floods.

In many thick sections, some units are so widespread as to be excellent key beds for correlation. In eastern Kentucky such beds include the Fire Clay coal, the Kendrick Shale Member, and the Magoffin Limestone Member of the Brethitt Formation. The coal bed is identified nearly everywhere by a parting of flint fire clay, and two marine zones occur in rather thick mudstone successions that generally are not interrupted by sandstone or coal beds. This presence of a limited number of widespread units in a very heterogeneous column probably is the result of fairly abrupt marine transgressions.

Source uplands ceased to provide much coarse
detritus during times of the major marine transgressions, perhaps because the climate temporarily became more humid and upland areas were protected from erosion by a denser growth of vegetation. The flint clay parting resembles “tonstein” partings of western Europe, which are thin beds with great lateral extent (L. R. Moore, University of Sheffield, written commun., 1967). The origin of tonstein bands has not been fully resolved, although some have proposed that they are volcanic-ash layers.

Probably both fluvial and deltaic sandstones are present in interval B. The frontal parts of sandy deltas were formed in marine waters, and marine fossils are found in the sandstones in some places; it seems unlikely, however, that the sandstones generally were distributed by marine currents, as proposed by Wilson and Stearns (1960) for sandstones in Tennessee.

Mudstones, sandstones, or underclays containing root zones form seat rocks beneath coal beds in the Appalachian region (Huddle and Patterson, 1961). The typical and valuable underclays of this region, including the kaolinitic Sciotoville Clay Member and Olive Hill Clay Bed, resulted from the leaching and mineralogical reorganization of a mixed assemblage of clays accumulated as soils.

Most of the coal beds record a period of transition from a forested emergent flatland to an inundated area with water-laid sediment of lake, lagoon, or marine environments. Details of configuration of the individual coal basins were no doubt controlled by the construction of suitable level platforms by alluvial or deltaic aggradation, but the development of a widespread coal swamp seems to indicate decreased influx of terrigenous sediment, rise in sea level, and climatic conditions favorable for luxuriant growth of vegetation.

Most of the Appalachian region was remote from the areas of persistent marine inundation of the South-Central and Southwestern United States during much of interval B time. The region was, however, briefly covered by marine waters at least several times during the interval, and four marine fossiliferous zones are recorded from northern Ohio. Of these, the Lower Mercer Limestone Member and its correlatives probably are present throughout most of the Appalachian region as far south as Tennessee and Virginia, and 10 separate marine inundations occurred in southern West Virginia during interval B (Reger, 1931a). Marine fossils have been found, locally, above most of the 29 coal beds of interval B in eastern Kentucky (J. W. Huddle, written commun., 1962). Five or more marine zones have been observed in northern Tennessee.

Interval B in the Black Warrior basin in Mississippi includes large amounts of soft flaky dark mudstone that probably accumulated in marine or brackish water. The seas generally transgressed northeastward through the Appalachian region from Mississippi to northern Ohio and northwestern Pennsylvania, but they did not reach eastward as far as the Anthracite fields (Stearns and Mitchum, 1962, p. 85). Marine fossil-bearing strata are not continuous throughout the region; the shoreline probably was extremely irregular because of low-lying alluvial plains and deltas that maintained themselves above water level at many places.

Numerous mudstones that contain abundant remains of freshwater pelecypods probably record lacustrine environments (Lucas, 1957). These mudstones may or may not be contemporaneous with marine mudstones deposited in other parts of the Appalachian region. Some of the pelecypods that have been reported probably record brackish-water rather than freshwater habitats (Eager and Turner, 1967).

The repetitive oscillations of the shoreline produced corresponding alterations in environments during interval B time. In Ohio, 12 environmental cycles, of which about half are recognized in the northern part, were recorded by Stout (1931, p. 215-216). In West Virginia, 29 cycles in interval B were noted by Reger (1931a, p. 234-235). The thick section in Big Black Mountain, Va. and Ky., shows 43 cycles (Eby, 1923). In general, each cycle records a fluvial period commonly marked by deposition of sandstone, followed in turn by a period of arrested sedimentation and weathering marked by underclay; a brief period of gradual inundation by the sea, marked successively by deposition of coal and lacustrine, lagoonal, or marine mudstones or limestones; and finally a withdrawal of the sea and a return to deltaic or fluvial environments in which mud and, later, sand were deposited.

Cyclic sedimentation has been explained by diastrophism (Weller, 1956), by eustatic changes related to worldwide oscillations of sea level (Wanless and Shepard, 1936; Wheeler and Murray, 1957), and by climatic changes (Beerbower, 1961). The problems of Pennsylvanian cyclic sedimentation are discussed in more detail in another part of this publication.

**PALEOTECTONIC IMPLICATIONS**

The northern Appalachian area was generally stable during interval B time. At the beginning of the interval, two separate basins were receiving sediment from the north and the southeast; these merged with burial of the dividing upland. Comparison of the sequence deposited in the basins with the much thinner sequence on the intervening high area suggests numerous nondepositional episodes or unconformities on the divide, although unconformities are not evident in local outcrops.

Monoclinal folding occurred along hinge lines (1) at the position of the subsequently active Paint Creek-Warfield fault zone in West Virginia and Kentucky, (2) in the position of the Pine Mountain thrust belt in Ken-
tucky, and (3) at the east edge of the deeper part of the Black Warrior basin in Mississippi, as shown by abrupt changes in the thickness of rocks on opposite sides of the folds. These folds were precursors of later thrusting or sharp folding during the Appalachian orogeny of Permian time. Areas to the southeast of the Appalachian region must have been repeatedly rejuvenated by uplift during interval B to provide sediments for the numerous wedges of detritus that advanced westward during deposition of the interval.

In southeastern Kentucky interval B is thicker than to the north, partly because subsidence began earlier in the south and was more rapid and less frequently interrupted than to the north. Evidence of the earlier start is furnished by the stratigraphic positions of the Fire Clay coal bed—within 50 feet or less of the interval base in northeastern Kentucky, but nearly 1,000 feet above the base in southern Kentucky. Evidence of more rapid accumulation is furnished by (1) the increase in vertical distance between widespread key beds where sandstone wedges enter the succession of strata to the south and (2) the preservation of erect fossil tree stumps, 15-20 feet high, at localities in southern West Virginia, Kentucky, and western Virginia.

Cyclic sedimentation of the Appalachian region, according to Weller (1956), may record periodic uplift of the source area adjoining the depositional basin, alternating with penepelation or subsidence. However, a tectonic cause for 40 or 50 such alternations during one of five divisions of Pennsylvanian time is difficult to visualize.

**INTERVAL C**

**FORMATIONS INCLUDED**

Interval C includes rocks generally placed in the Allegheny Formation or Group in the Appalachian region. Interval C contains the well-known and widespread Brookville, Clarion, Kittanning, and Freeport coal beds (table 6), which were named for locations in western Pennsylvania. These names have been applied widely in Pennsylvania, Ohio, Maryland, and West Virginia to coal beds and associated clays, mudstones, and sandstones. The terms Clarion, Kittanning, and Freeport have been proposed as names for formations within the Allegheny Group (Williams, 1960, p. 910). In central West Virginia the upper part of the Kanawha Formation above the base of the Coalburg coal also is referred to interval C.

In the Anthracite fields the name Llewellyn Formation has been applied to the major coal-bearing sequence above the Pottsville Formation (Wood and others, 1963, p. 6). The Llewellyn Formation in its lower part includes rocks equivalent to the Allegheny Formation of the bituminous coal field. The Sharp Mountain Member of the Pottsville in the Anthracite fields contains in its upper part plant fossils that suggest a possible early Allegheny age, but the member is referred to interval B for reasons stated in the discussion of interval B. The equivalents of the Kittanning and Freeport of western Pennsylvania are represented in the Llewellyn Formation as are the lower Conemaugh strata referable to interval D.

In eastern Kentucky north of Pine Mountain (fig. 6) the Breathitt Formation contains strata referred to interval C. In Big Black Mountain on the Kentucky-Virginia line strata assigned to interval C are called the Harlan Sandstone, and in the Log Mountains area of Kentucky, the Bryson Formation. Interval C is absent south of the Kentucky-Tennessee line.

**STRATIGRAPHIC RELATIONS**

The Allegheny Formation was originally defined to include an interval from the base of the Brookville coal to the top of the Upper Freeport coal. Rocks above and below are characterized by little or no commercial coal in the areas of their typical exposures. The formation boundaries are difficult to recognize accurately beyond the limits of the Brookville and Upper Freeport coals.

Faunal markers are of limited use in consistently defining the interval. The genus Fusulina is believed to be limited to approximately the time of interval C. It has been identified in two marine limestones—the Putnam Hill and Vanport in the lower part of the Allegheny Formation of Ohio. Where these limestones are traceable into other parts of the Appalachian region, they serve as excellent key beds.

Floral zones 9 and 10 characterize the Allegheny Formation (Read and Mamay, 1964). Zone 9, marked by Neuropteris rarinervis, is typical of the lower Allegheny, and zone 10, containing Neuropteris flecuosa and Pecopteris spp., of the upper Allegheny and lower Conemaugh (interval D). Although many roof shales of coal beds in the Appalachian region contain well-preserved plant compressions, most of the floras associated with particular coal beds are not yet described. Spores of the Allegheny Formation have been studied (Cross, 1947; Habib, 1965; Gray, 1965), but distinctive spores for various parts of interval C in the Appalachian region are not yet well known.

In the Anthracite fields many minable coals occur in the Llewellyn Formation, and it is uncertain which of them is correlative with the Upper Freeport (table 6). The boundary between intervals C and D is picked arbitrarily within the productive measures.

In southern West Virginia, southeastern Kentucky, and western Virginia, some of the coals are believed to be equivalent to strata in the Allegheny Formation farther north and, therefore, are indicative of interval C.

---

*In this publication individual authors may have used either Fusulina or Beedoea in reference to certain fusulinids of Desmoines age. Beedoea now is recognized as the proper name for these fusulinids. See Beedoea Galloway emend. Jable (1967, p. 656).*
### Table 6.—Stratigraphic units mentioned in text for interval C but not shown on plate 13: Appalachian region

<table>
<thead>
<tr>
<th>Interval</th>
<th>Pennsylvania–Ohio</th>
<th>Pennsylvania Anthracite fields</th>
<th>West Virginia</th>
<th>Kentucky</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Mahoning Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Upper Freeport coal</td>
<td>Holmes coal</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Upper Freeport Sandstone Member</td>
<td>(Equivalent unit in Pennsylvania–Ohio area not known)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower Freeport coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower Freeport Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Johnstown Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Middle Kittanning coal</td>
<td>Mammoth coal</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Columbiana Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Strasburg Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lower Kittanning coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kittanning Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Vanport Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Clarion coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Upper Clarion Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Clarion Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Zaleski Flint Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Putnam Hill Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td></td>
<td>Brookville coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Notes:**
- **Llanoyn Formation:** Location and extent of unit not clearly defined.
- **Kilgore? Flint:** Presence and extent unconfirmed at this time.
- **Sharp Mountain Member:** Stratigraphic detail not specified.
as defined in that area. Although lack of precise criteria makes correlation uncertain, current studies of plant spores should aid in achieving more uniform boundaries in this area (R. M. Konske, written commun., 1965).

**UPPER BOUNDARY OF INTERVAL C**

The top of the *Upper Freeport coal bed*, where present and recognized, is used as the upper boundary of interval C. At some other places the base of the overlying Mahoning Sandstone Member in the basal part of the Conemaugh Formation is used as the upper boundary. In the Southern Anthracite field the exact equivalent of the *Upper Freeport coal bed* is uncertain, and the *Holmes coal* has been arbitrarily chosen as the upper boundary of interval C.

In Kentucky south of Lawrence County, in southern West Virginia, and in Virginia, no younger Pennsylvanian rocks overlie interval C.

**THICKNESS TRENDS**

Interval C is nearly 100 feet thick, possibly much more, in the Anthracite fields (pl. 7A), and is as much as 1,189 feet in Big Black Mountain near the Kentucky-Virginia line. Throughout much of the northwestern half of the bituminous field the total thickness is 200-350 feet. Farther southeast the thickness is more variable, probably because of differential compaction.

The rate of southeastward thickening of interval C is less marked than that of intervals A or B, which may indicate a decrease in the rate of downwarp in deeper parts of the basin. However, interval C has been eroded from parts of southern West Virginia, Virginia, and Tennessee, whereas intervals A and B attain their maximum thicknesses in these areas.

In Maryland and Pennsylvania the present-day bituminous coal field is divided by three parallel northeast-trending anticlines (fig. 8). A decrease in thickness of interval C on the Chestnut Ridge anticline suggests that moderate differential uplift or downwarp was in progress during the time this interval was deposited.

**LITHOFACIES TRENDS**

Interval C in the Appalachian region includes conglomerate, sandstone, mudstone, underclay, coal, limestone, and flint. Ferruginous concretions are a minor but locally important constituent of the shale and sandstone. In all areas the conglomerate, sandstone, and mudstone exceed the nondetrital sediments by a ratio of more than four to one.

A prominent belt in which sandstone exceeds mudstone by more than four to one is in Nicholas, Clay, Kanawha, Putnam, and Braxton Counties, south-central West Virginia (fig. 5), where locally most of the interval is composed of a succession of massive sandstones, collectively called *Charleston Sandstone* by Campbell and Mendenhall (1896, p. 488, 508). A belt of somewhat similar facies occurs farther south in Fayette, Boone, Logan, and Mingo Counties, W. Va., and in Pike County, Ky., where only the lower part of the interval is present. Likewise, in Portage County, northern Ohio, and in parts of northern Pennsylvania, only the lower part of the interval is present and here also the proportion of sandstone is high.

The areas that have the highest proportions of mudstone are mainly near the northwest border of the Appalachian region, especially in Tuscarawas, Coshocton, and Muskingum Counties, Ohio. Interval C, in general, contains more mudstone than sandstone near the southeast border of the region in Pennsylvania and more sandstone than mudstone on the southeast border of the region in Maryland and West Virginia. The Broad Top coal field (fig. 9) of southern Pennsylvania has a fairly large proportion of mudstone compared to nearby areas.

Conglomerates containing pebbles of quartz and metamorphic rock are common throughout interval C in the southern part of the Anthracite fields. In southern and central West Virginia several beds of pebbly sandstone are composed largely of rounded quartz fragments.

Sandstones in interval C include lenticular bodies, some of which are convex downward and others convex upward; they also include many widespread beds of almost uniform thickness. These widespread sandstones occur at six stratigraphic positions in the northern part of the field. Named from the base up, they are the Clarion and *Upper Clarion, Kittanning, Strasburg, Lower and Upper Freeport Sandstone Members* of the Allegheny Formation, respectively.

Interval C includes laminated gray noncalcareous shaly mudstone containing fossil plants, calcareous shaly mudstone containing marine fossils, silty mudstone or siltstone, dark-gray to black soft shaly mudstones, and underclay. The most common type of mudstone in the interval is gray, noncalcareous, and shaly. It locally contains fossil floras. Locally in northeastern Kentucky the upper part of the interval includes red and mottled red and gray mudstones.

Coal beds are widely distributed, but the aggregate thickness of coal in the interval is generally small, except in the Southern Anthracite field. There, the bottom split of the *Mammoth coal zone* averages about 40 feet thick and in places is as much as 80 feet thick. The bituminous field in Pennsylvania includes six widespread coal beds—the *Brookville, Clarion, Lower Kittanning, Middle Kittanning, Lower Freeport*, and *Upper Freeport*—and several thinner and more localized beds. Individual coal beds are more widespread than those of interval A or interval B in the Appalachian region but are less widespread than coal beds of interval C in the Illinois basin.
Limestone makes up three thin but widespread units. In ascending order, these are the Putnam Hill, Vanport, and Colombiana. The Putnam Hill and the Colombiana Limestone Members are more conspicuous in Ohio than in areas farther east, but the Vanport Limestone Member is also widespread in northern Pennsylvania. A freshwater limestone, the Johnstown Member, occurs locally in south-central Pennsylvania.

Flint or chert makes up a small part of interval C in a few areas. Black flints known as the Zaleski, Kanawha Black, Flint Ridge, and Kilgore Flints occur in southern Ohio, western central West Virginia, and northern Kentucky. Gray chert on Flint Ridge in Licking County, Ohio, is part of the Vanport Limestone Member.

Ferruginous concretions, made up of siderite, are present but less common than in interval B. They are prominent at the stratigraphic horizon of the Vanport Limestone Member.

Because each of the major lithologies of interval C is repeated vertically several times, the lithofacies patterns on the map (pl. 7B) fail to show the regional distribution of many sedimentary units representing particular brief periods of time.

SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES

Patterns of distribution of detrital sediments in the Appalachian region appear to give a clear picture of directions of sediment flow during interval C time. The distribution of areas with large proportions of sandstone suggests that the principal uplands yielding sediments lay southeast of the region and that the more southern source areas were higher or lay near the basin of deposition. Belts in which the proportion of sandstone is very high, as in the Charleston, W. Va., area, represent sites of either recurrent or continuous channelways of major distributary streams in large delta complexes. Minor drainage courses, represented by a high proportion of sandstone, that occur in the lower part of the interval in northern Ohio and Pennsylvania probably indicate that some sources of sediment lay north of the Appalachian region. These sources, however, were less important than those to the southeast. In nearly all parts of the Anthracite fields, sandstone is more abundant than mudstone, and fairly coarse conglomerate is common. Nearby source areas of moderate relief are therefore inferred.

Widespread detrital units in the lower Allegheny Formation had their principal sources southeast of West Virginia, probably in the present Blue Ridge and Piedmont areas of Virginia (Baroffio, 1964). Other less important sources of sediment were east of central Pennsylvania and north of the present coal field of Ohio and Pennsylvania, in what is now New York or Ontario. The source area in eastern Pennsylvania may have supplied sediment westward toward the bituminous field and northward toward the Anthracite fields; however, the latter were closer to the source, which has been tentatively located in the area near the present site of Philadelphia (G. H. Wood, 1963, written commun.). At the beginning of Allegheny time, nearly all the detrital sediment was derived from southeasterly sources, but by middle Allegheny time, Ohio and western Pennsylvania received significant amounts of sediment from the north, and east-central Pennsylvania received detritus from the east (Baroffio, 1964; Cavorac and others, 1963).

ENVIRONMENTS OF DEPOSITION

Many contrasts in lithology that indicate differences in environment characterize interval C, as well as intervals A and B. The Allegheny Formation of Ohio was divided by Stout (1931, p. 214-216) into 12 sedimentary cycles, of which 5 are traceable through large parts of the Appalachian region and 7 are recognizable only in small areas. In a generalized section of this formation, 212 feet thick, he shows 6 sandstones, 9 mudstones, of which 2 contain marine faunas, 13 coals, 9 underclays, 3 marine limestones, 3 nonmarine limestones, 3 iron deposits, and 1 flint.

The following is a general history of the environmental changes during a representative widespread cycle that is considered to have begun during the accumulation of sandy sediments:

Highlands southeast of the basin and probably about 100 miles distant from it provided detritus that was carried northwestward across the coal basin. The southeast border of the basin was a piedmont plain made up of coalescing alluvial aprons. Near the middle of the basin, gradients were low, and flood-plain deposits were prevalent. At this time the sea lay west of the present coal basin, nearly all parts of which were emergent.

At a late stage the amount of sand entering the basin diminished, and finer detrital particles of silt or clay were widely deposited. The change in type of sediment may have been due to a climatic change which allowed the growth of a vegetative cover that protected slopes in the highlands from excessive erosion. Sediments accumulated slowly, leaching of alkalies and of alkali earths took place, and the clays became kaolinitic. Refractory kaolinitic clays are more common toward the northern part of the basin, so it is assumed that clays were exposed to weathering for a longer time in that area than farther southward in the basin.

At a late stage, sea level apparently rose, and the shoreline moved toward the west border of the present coal basin in Ohio. Rising water tables on the land probably slowed oxidation of plant detritus. Lime mud accumulated in ponds in small depressions, and coal swamps developed locally near the outer borders of
deltas and in lagoons behind barrier beaches. During a few of the cycles, coal swamps spread over a large part of the basin. Toward the southeast, sediment that entered the basin spread out over the coal swamps, as shown by partings of claystone or sandstone that divide many coal beds near the southeastern margin of the basin in Pennsylvania and West Virginia.

The sea eventually spread over the swamps, and the vegetation that had accumulated was buried with mud or marl. Farther from shore, or at times when little mud was introduced into the basin, lime mud accumulated. In lagoons and inlets iron carbonate was deposited as siderite, or siliceous muds formed chert or flint.

Later the sea withdrew, prodelta muds buried the lime mud, the exposed areas gradually reverted to a plain, and a new cycle started.

Adjustment of the faunas to the changing environments of a drowning topography has been described by Williams (1960). Marine mudstones have been distinguished from freshwater mudstones in the Allegheny rocks by using trace elements (Degens and others, 1957).

Relief of the surface of deposition in the Appalachian region during all of interval C was very low. All pre-Pennsylvanian hills and valleys had been buried before the beginning of the interval, and the surface must have been very low and flat. Differential downwarping of the southern part of the region was less than during the times of the earlier intervals.

**PALEOTECTONIC IMPLICATIONS**

In western Pennsylvania, strata of middle Allegheny age are strongly contorted because of penecontemporaneous deformation (Ferm and Huddle, 1955; Williams and others, 1964). Contorted beds are reported in 50 percent of the outcrops in some areas. The distorted structures are limited at the top and bottom by undisturbed strata. This contortion is interpreted as having occurred along streambanks (Williams and others, 1964), and probably has no paleotectonic implications. The most widespread zone of excessive deformation is near the position of the *Middle Kittanning coal bed* (middle Allegheny), and it extends east-west for about 30 miles between Brookville and Rochester in western Pennsylvania.

Thinning of the Allegheny Formation suggests that incipient folding may have taken place along the Chestnut Ridge anticline (fig. 8) and nearby anticlines during interval C time. If so, however, the amplitude of folding was only a few tens of feet. The data are inconclusive, moreover, and the conclusions stated here have been strongly questioned by some geologists.

A general southward thickening of interval C, particularly in southern West Virginia and in the Big Black Mountain area of Virginia, probably indicates that downwarping was somewhat greater in the southern part of the northern Appalachian basin than in the remainder of the basin.

Local uplifts to the north and east of Ohio and Pennsylvania and perhaps a gentle tilting of the whole basin that partially reversed northwest-dipping paleoslopes are indicated for the later part of interval C time by an increase of detrital sediment in the northern part of the basin.

The periodic influxes of large quantities of sand-size sediment into the basin may have resulted from a series of local uplifts of source areas. An alternative nontectonic explanation is that the quantity of sediment eroded and transported depended on the extent of the vegetative cover on upland areas, with much less sand or silt being eroded from forested than from semiarid upland areas.

Six marine transgressions recognized in the Appalachian basin during Allegheny time may have been due either to alternate subsidence and uplift or to eustatic rise and fall of sea level.

**INTERVAL D FORMATIONS INCLUDED**

Strata referred to interval D in the Appalachian coal basin comprise the lower part (locally all) of the Conemaugh Formation or Group. These strata were originally known as the Lower Barren Measures (Platt, 1875, p. 8). The Conemaugh recently has been assigned group status where it can be divided into two units, the Glenshaw and the overlying Casselman Formations (Flint, 1965, p. 70-73). The upper part of the Casselman is assigned to interval E. In contrast with the underlying Allegheny Formation and the overlying Monongahela Group or Formation, the Conemaugh contains little coal of minable thickness.

**STRATIGRAPHIC RELATIONS**

The Conemaugh Formation is defined as extending from the top of the *Upper Freeport coal bed* up to the base of the *Pittsburgh coal bed*. The part of the Conemaugh assigned to interval D correlates throughout the United States with rocks containing the more primitive forms of the fusulinid genus *Triticites*, which is referred to the subgenus *Kansanella* (Thompson, 1957). Fusulinids belonging to early forms of *Triticites* are in the Brush Creek, Cambridge, and Ames Limestone Members of the Glenshaw and *Gaysport Limestone* in the Casselman (table 7) in Ohio (Thompson, 1936; Smyth, 1957). The highest limestone strata containing fusulinids, the Ames and *Gaysport*, seem to correlate with strata in the midcontinent region a little below the Missouri-Virgil boundary, which is used as the interval D-interval E boundary in that region.

In the Anthracite fields a horizon equivalent to that of the *Upper Freeport coal bed*, which forms the boundary
between the Allegheny and Conemaugh in the bituminous fields, has not yet been identified with certainty; the Llewellyn Formation includes strata currently referable to interval D as well as those assigned to interval C (Wood and others, 1963, p. 6). The upper part of the Llewellyn Formation, from the top of the Holmes coal bed to the top of the Pennsylvanian in the Anthracite fields, is referred to interval D, according to assignments by G. H. Wood (written commun., 1971), who has revised the data provided by Wanless. The possibility that a thin, upper part of the formation is of Monongahela age is indicated on plate 13 and in table 7. Differences in thicknesses assigned to the Upper Pennsylvanian in the Anthracite fields by Wanless and by Wood are shown in table 8.

Marine faunas have not been discovered in the upper part of the Conemaugh Group of the northern Appalachians; for this reason age assignments are uncertain. The base of the Morgantown Sandstone Member of the Casselman Formation, a widespread and locally well developed sandstone unit, has been used arbitrarily for the interval D-interval E boundary. The lower two-thirds or three-fourths of the Conemaugh Group of the northern Appalachian region is the base of the Morgan town Sandstone Member of the Casselman Formation, a widespread and locally well developed sandstone unit, has been used arbitrarily for the interval D-interval E boundary. The lower two-thirds or three-fourths of the Conemaugh, which includes all known marine faunas in the formation, is referred to interval D, and the highest third or fourth is referred to interval E. Marine limestones are less persistent near the southeastern part of the outcrop area, and in several localities in West Virginia the boundary is approximated with less certainty than elsewhere.

Interval D is characterized in its lower part by fossil plant compressions of floral zone 10 (Read and Mamay, 1964, p. K10-K11), which is distinguished by Neuropteris flexuosa and various species of Pecopteris, and above this by the lower part of floral zone 11, which contains Lescuropteris spp. and also extends up into interval E. Few lists of plants from the Conemaugh have been published, and little study has been made of spores from coals and mudstones, so greatest reliance is placed on the marine faunas, particularly the fusulinids.

UPPER BOUNDARY OF INTERVAL D

The boundary between intervals D and E in the northern Appalachian region is the base of the Morgantown Sandstone Member of the Casselman Formation. This boundary lies slightly above the Skelly Limestone, which is the youngest marine unit of the Appalachian region; the underlying Ames and Gaysport, below the Skelly, contain fusulinids equivalent to those of very late Missouri age in the midcontinent region.

THICKNESS TRENDS

Interval D, in areas where it is present in its originally deposited thickness, ranges in thickness from 180 feet at locality 418 (Lawrence County) in southern Ohio to a little more than 500 feet at localities 440 (Cambria County)

---

**Table 8.** Differing assignments of Upper Pennsylvanian sequences to intervals D and E in the Anthracite fields of Pennsylvania

<table>
<thead>
<tr>
<th>Loc. (pl. 1)</th>
<th>Interval D</th>
<th>Interval E</th>
<th>Interval D</th>
</tr>
</thead>
<tbody>
<tr>
<td>75</td>
<td>166</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>89</td>
<td>91</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>107</td>
<td>235</td>
<td>--</td>
<td>260</td>
</tr>
<tr>
<td>127</td>
<td>263</td>
<td>--</td>
<td>220</td>
</tr>
<tr>
<td>142</td>
<td>--</td>
<td>--</td>
<td>99</td>
</tr>
<tr>
<td>143</td>
<td>231</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>165</td>
<td>424</td>
<td>--</td>
<td>448</td>
</tr>
<tr>
<td>166</td>
<td>903</td>
<td>--</td>
<td>820</td>
</tr>
<tr>
<td>256</td>
<td>150</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>317</td>
<td>--</td>
<td>--</td>
<td>600</td>
</tr>
<tr>
<td>324</td>
<td>635</td>
<td>540</td>
<td>800</td>
</tr>
<tr>
<td>347</td>
<td>830</td>
<td>445</td>
<td>1,050</td>
</tr>
<tr>
<td>348</td>
<td>1,065</td>
<td>375</td>
<td>1,020</td>
</tr>
<tr>
<td>373</td>
<td>480</td>
<td>--</td>
<td>480</td>
</tr>
<tr>
<td>375</td>
<td>885</td>
<td>--</td>
<td>880</td>
</tr>
</tbody>
</table>

*No rocks assigned to interval E.*
and 461 and 478 (Washington County) in Pennsylvania and, possibly, at locality 297 (Braxton County) in West Virginia. The interval is as much as 1,050 feet thick in the Anthracite fields of Pennsylvania (table 8), although thrust faults make accurate measurements difficult in that area. Interval D has been partly removed by recent erosion in belts of various widths around the margins of depositional basins of the Appalachian region, so the present thickness of the interval is less than the original thickness in these areas. In nearly all localities where interval D is shown as less than 200 feet thick, the upper part has been eroded.

The average rate of thickness change in interval D is less than in intervals A, B, and C, but interval D is absent in the southern Appalachian area, where the older intervals show the greatest thickening. In that area, interval D has been removed by erosion.

Thickness differences near the Chestnut Ridge, Laurel Hill, and Negro Mountain anticlines and the adjacent synclines (fig. 8) are difficult to interpret because the upper parts of the interval are generally eroded from anticlinal crests. Uneroded sequences between localities 364 and 371 in Pennsylvania have the following thicknesses:

<table>
<thead>
<tr>
<th>Loc. (pl. 1)</th>
<th>Thickness (ft)</th>
<th>Structural position</th>
</tr>
</thead>
<tbody>
<tr>
<td>364</td>
<td>463</td>
<td>Basin west of Chestnut Ridge anticline.</td>
</tr>
<tr>
<td>365</td>
<td>351</td>
<td>West flank of Chestnut Ridge anticline.</td>
</tr>
<tr>
<td>367</td>
<td>246</td>
<td>East of crest of Chestnut Ridge anticline.</td>
</tr>
<tr>
<td>368</td>
<td>416</td>
<td>Ligonier basin, east of Chestnut Ridge anticline.</td>
</tr>
<tr>
<td>369</td>
<td>288</td>
<td>Near crest of Laurel Hill anticline.</td>
</tr>
<tr>
<td>371</td>
<td>396</td>
<td>Basin between Laurel Hill and Negro Mountain anticlines.</td>
</tr>
</tbody>
</table>

These relations suggest that folding was in progress during interval D time, but the validity of this conclusion has been questioned by some geologists.

Formerly, much of the Monongahela Formation (interval E) was believed to be represented by strata in the Southern Anthracite field. Recent mapping (Wood and others, 1963, p. 69), however, has revealed thrust faults that cause the Pennsylvania sequence to be repeated in this area. The youngest strata of the Southern Anthracite field appear to belong to the lower part of interval D rather than to the higher part, as was supposed before duplication by thrust faulting was discovered. Some thicknesses on plate 8A may be excessive, but others, including a maximum of a little more than 800 feet in the Northern Anthracite field, almost certainly are correct if the top of the Holmes coal bed and its correlatives are considered to be the base of interval D. The more accurate figures seem to indicate that the maximum thickness of interval D in the Anthracite fields is at least double that in the Pennsylvania bituminous field.

**LITHOFAECIES TRENDS**

Interval D in the Appalachian coal field consists principally of mudstone, sandstone, and siltstone, but includes lesser quantities of coal, underclay, and marine limestone. At all localities sandstone and mudstone exceed the other constituents by a ratio of more than four to one. Mudstone generally makes up 50–80 percent and sandstone 20–50 percent of the rocks present. Sandstone exceeds mudstone in the roughly linear areas near the southeast border of the field where lenticular channel-type sandstones replace much of the mudstone. In several areas near the northwest border of the field, mudstone exceeds sandstone by a ratio of more than four to one.

Interval D contains five named sandstone members widely distributed in Ohio (Condit, 1912, p. 20), seven in Pennsylvania (Dutcher and others, 1959, p. 69), and nine in West Virginia (Reger, 1931a, p. 229-230). Each of the sandstones is locally massive and forms a conspicuous bench or ridge; all are discontinuous and commonly grade laterally into mudstones. Each sandstone is a composite of bifurcating channel deposits. Differences in the proportions of sandstone to mudstone in interval D commonly are the result of superposition of the five to nine principal sandstones.

In the Southern Anthracite field at least 15 well-developed sandstones, several of which are conglomeratic, are present (Wood and others, 1963, p. 7).

Most mudstones of interval D in Pennsylvania and northern Ohio are gray, but a few are red brown. In central West Virginia nearly all are red brown.

Marine limestones that have been named occur at six horizons in interval D in Ohio, at three in Pennsylvania, and at four in northern West Virginia. The three most widespread, in ascending order, are the Brush Creek, Cambridge, and Ames. To the southeast these limestones wedge out. Nowhere is the aggregate thickness of the marine limestones more than 5-10 feet. One fossiliferous limestone in the Llewellyn Formation, known as the Mill Creek Limestone, is present in the Northern Anthracite field near Wilkes-Barre and, if the paleontologic identification and interpretations are correct, is equivalent to the Ames Limestone Member in the bituminous fields (Chow, 1951). Plant fossils collected from an overlying coal bed are characteristic of Read and Mamay's floral zone 10 (G. H. Wood, written commun., 1970).

Freshwater limestone occurs at three to five stratigraphic positions in interval D, but is generally thin and much less conspicuous than in overlying interval E.

In the main coal field, 8 to 12 coal beds have been reported, of which 4 are very widespread; the remainder
are relatively discontinuous. In the Anthracite fields, 24 coal beds are reported (but some are splits of other beds), of which 14 are thick enough to have been mined, at least in parts of the field.

SOURCES AND ENVIRONMENTS OF DEPOSITION

The larger proportion and the coarser texture of sandstones in the eastern and southeastern parts of the areas occupied by interval D suggest that sources of sediment lay east and southeast of the present basin, probably as discontinuous highlands within what are now the Blue Ridge and Piedmont provinces. The Cincinnati arch seems improbable as a major source of sediment because interval D is thinner and generally composed of finer grained sediment along the western margin than elsewhere in the basin.

Few studies of transport direction in sandstones of this interval have been made, but coarse detrital sediment of the Anthracite fields contains some rock fragments that probably were derived from quartz veins in the Wissahickon Schist of eastern Pennsylvania. The distribution of the Mahoning Sandstone Member of the Glenshaw near the base of interval D in northern Pennsylvania suggests that sediments reached the basin from northerly, probably Canadian, sources. Similar conclusions have been reached by mapping (Morris, 1967) of both the Mahoning and the Buffalo Sandstone Members.

ENVIRONMENTS OF DEPOSITION

A generalized section for western Pennsylvania shows 32 lithologically distinct stratigraphic units that have fairly wide distribution (Dutcher and others, 1959, p. 69), a section for Ohio shows 40 units (Stout, 1939), and a section for West Virginia shows 52 units (Rege, 1931a, p. 228-230). As each of these units required environmental conditions different from those of adjoining strata above and below, frequent environmental changes are clearly indicated throughout large parts of the Appalachian basin.

Environmental maps (Morris, 1967) of individual strata from the base of rocks assigned here to interval D to the Cambridge Limestone Member, near the middle of the interval, show a locus of deposition at about the center of the northern Appalachian basin near the southwest corner of Pennsylvania. Deltaic sand bodies at the terminus of northwest-flowing streams formed at times of major sandstone deposition. Linear sand bodies near the center of the basin are interpreted as distributary channels, commonly flanked by natural levees. The central part of the basin was the site of freshwater lakes in which limestones and marly mudstones formed. At times of major marine transgressions, marine waters replaced fresh or brackish waters. Stream valleys and delta complexes formed in smaller drainage systems that lay east, north, and northwest of the center of the basin.

Near the end of deposition of major sandstones, sediment flow diminished in quantity, and particle size decreased, perhaps because of the protection offered upland soils by a vegetative cover. As rates of deposition diminished, the areas of lowland sand accumulation supported vegetation, whose root impressions are recorded as a network in the underclays.

The surface made up of piedmont, fluvialite, and deltaic sediment probably was uneven, and the rising water in the basin at the beginning of a marine transgression likely inundated the lower parts. Freshwater limestones accumulated in lakes or ponds, and marsh vegetation accumulated to form peat in areas of shallow water, ultimately becoming the coal beds of interval D. Areas at water level and above continued to support growing vegetation but plant material normally oxidized and disappeared, leaving only carbonaceous films and underclays (seat rocks) to mark the time of deposition of coal beds elsewhere. Coal swamps formed as frequently as during the times of preceding intervals, but, for some reason not apparent, they were more localized and existed more briefly, giving this interval its "barren measures" character, so far as minable coals are concerned.

Fluctuations in sea level resulted in seven marine transgressions and subsequent regressions during the time of interval D. The sea advanced from the west. It covered Ohio and northeastern Kentucky first and remained there longest. The eastward spread of the sea was limited by advancing tongues of deltaic sediment from the southeast and by flood plains already built too high to be flooded. One of the marine transgressions, the Brush Creek, is represented in southern Ohio by an upper and a lower limestone, locally 20-40 feet apart vertically. A detrital wedge, probably of deltaic origin, separates the earlier and later marine strata (Morris, 1967). This wedge accumulated during a minor oscillation of the sea.

Mud carried into the sea by rivers was deposited with calcium carbonate to produce limestones having an earthy, argillaceous composition. The seas that deposited the Portersville and Skelly Limestones do not seem to have extended farther east than Ohio. The Gaysport sea extended a short distance into West Virginia, the Brush Creek and Cambridge seas farther into Pennsylvania and West Virginia, and the Ames sea may have extended as far east as the Northern Anthracite field.

The red color of the sediments increasing southward in West Virginia may have resulted from contemporaneous or later oxidation of iron in subaerial piedmonts or valley alluvium. On the other hand, the color may be due largely to a red regolith washed off such
lowlands as the Piedmont areas to the east or the Cincinnati arch to the west (Case, 1915, p. 80-85).

**PALEOTECTONIC IMPLICATIONS**

Areas southeast, east, and probably north of the northern Appalachian basin probably were elevated repeatedly during the time of interval D, as shown by coarse- or medium-grained detrital sediments that reached the basin intermittently from sources in those directions. The fluctuations in the supply of sediments may have been at least partly controlled by nontectonic causes, as well, including climatic fluctuations that might have produced differences in the vegetative cover and, consequently, differences in the rate of erosion on the highland.

Downwarp of the basin was greatest along the east side; the thickest sediments appear to be nearest to the source areas. Subsidence of the basin is indicated by the fact that the highest marine limestone, the Skelly, is 150-180 feet above the lowest, the Brush Creek; the basin, therefore, must have been downwarped by at least that amount in some areas. Variations in thickness of interval D across the Chestnut Ridge and Laurel Hill anticlines and intervening synclines suggest incipient folding along these features during interval D time.

**INTERVAL E**

**FORMATIONS INCLUDED**

Interval E includes about the upper one-third of the Conemaugh Formation or Group and the overlying Monongahela Formation or Group. The horizon used for the lower limit of the interval is the base of the Morgantown Sandstone Member (Flint, 1965, p. 71) of the Casselman Formation of the Conemaugh Group. The Ames Limestone Member of the Glenshaw Formation and the Gaysport Limestone in the Casselman (table 9), which occur below the Morgantown Sandstone Member, contain some fusulinids which may be indicative of early Virgil age, as recognized in the midcontinent region, but other fusulinids more clearly indicate correlation of these limestones with strata of Missouri age. The Monongahela was originally called Upper Productive Measures because it contains the thick and extensively mined Pittsburgh coal, as well as overlying mined coal beds known in Pennsylvania as the Redstone, Sewickley, Uniontown, and Waynesburg. The Monongahela is now considered to be a group where it can be divided into two formations, the Pittsburgh and the overlying Uniontown, the base of which is the Uniontown coal and the top of which is the base of the Waynesburg coal (Berryhill and Swanson, 1962).

**STRATIGRAPHIC RELATIONS**

Marine fossils are not known in the Monongahela or the upper part of the Conemaugh Formation, so fusulinids cannot be used as stratigraphic indices. Floral

**UPPER BOUNDARY OF INTERVAL E**

Strata above the Waynesburg coal bed, assigned in this publication to the Permian, were formerly termed the "Upper Barren Measures" because, like the Conemaugh Group, they contain few minable coal beds. Later they were named the Dunkard Group from a locality in Green County, Pa. (White, 1891, p. 20). The discovery of a flora including Callipteris conforma in the Dunkard strata (Fontaine and White, 1880) led to the assignment of the Dunkard to the Permian. Callipteris is the index genus of the earliest Permian floral zone (Read and Mamay, 1964, p. K12).

Validity of the identification of Callipteris has been questioned, and the position of the systemic boundary has not been supported by any independent lithologic criteria. Numerous proposals have been made that substantial parts of the Dunkard Group should be transferred to the Upper Pennsylvanian. The Dunkard

| Table 9: Stratigraphic units mentioned in text for interval E but not shown on plate 13: Appalachian region |
|---|---|---|
| **Pennsylvanian and Permian** | **Dunkard Group (part)** | **Waynesburg Formation** |
| **Waynesburg coal** | **Waynesburg coal** | **Waynesburg coal** |
| **Uniontown Formation** | **Uniontown Formation** | **Uniontown coal** |
| **Sewickley coal** | **Sewickley coal** | **Sewickley coal** |
| **Redstone coal** | **Redstone coal** | **Redstone coal** |
| **Pittsburgh coal** | **Pittsburgh coal** | **Pittsburgh coal** |
| **Morgantown Sandstone Member** | **Morgantown Sandstone Member** | **Morgantown Sandstone Member** |
| **Gaysport Limestone** | **Gaysport Limestone** | **Gaysport Limestone** |
| **Ames Limestone Member** | **Ames Limestone Member** | **Ames Limestone Member** |
Group is divided into three formations—the Waynesburg, the Washington, and the Greene, in ascending order. The Waynesburg, including the Waynesburg coal bed, was reassigned to the Pennsylvanian-Permian by Berryhill and Swanson (1962), and all of the Dunkard may be Pennsylvanian according to some geologists (for example, Clendening, 1969). All the Dunkard Group was mapped with the Permian by McKee, Oriel, and others (1967).

Correlation of the systemic boundary between the Appalachian and midcontinent regions is very difficult. Freshwater invertebrates and fish remains that characterize the latest Pennsylvanian in the Appalachian region are absent in the midcontinent region. The Appalachian freshwater floras have not been studied in sufficient detail to evaluate their relative Pennsylvanian and Permian affinities. Ultimately, it may be necessary to determine the Permian-Pennsylvanian boundary in this region by a comparison with nonmarine facies outside the conterminous United States.

**THICKNESS TRENDS**

Where the entire sequence of interval E is present, thicknesses range from about 400 feet near the northwestern margin of the region in Ohio to a maximum of 807 ± feet in Calhoun County, W. Va. (pl. 9A). The rate of southeastward thickening is comparable with that of interval D. Local variations probably result from differing proportions of less compactible sandstones and freshwater limestones in relation to more compactible mudstones and from inconsistencies in picking the upper and lower boundaries of the interval. The Morgantown Sandstone Member, for instance, is discontinuous and can be easily confused with other locally conspicuous sandstones in the upper part of the Conemaugh Formation. The Waynesburg coal bed is also discontinuous and generally is absent in the southeastern part of the basin. In the deeper part of the basin, most of the records that include the base of interval E are oil-well drillers' logs on which neither the upper nor the lower boundary can be clearly identified. In the outlying synclinal basins east of the main bituminous coal field, the uppermost strata of the interval have been eroded at nearly all places, except possibly at locality 15, Allegany County, Md., where interval E is about 600 feet thick.

**LITHOFACIES TRENDS**

Interval E consists principally of cyclic alternations of sandstone, siltstone, mudstone, limestone, underclay, and coal beds, which constitute 40-60 distinctive units in various parts of Ohio, Pennsylvania, and West Virginia. Limestones are sufficiently thick and numerous to make up more than 20 percent of the total thickness of the interval in a broad belt in the northwest half of the region, northwest of a line extending from Washington County, Ohio, northeastward to Fayette County, Pa., and including the northern part of the outlier in Somerset County, Pa. (pl. 9B). Limestone composes as much as 50 percent of the total thickness at two localities in Washington County, Pa.

Limestone generally is less than 20 percent of the total thickness north of the northwestern belt from Noble County, Ohio, northeastward to Westmoreland and Indiana Counties, Pa. In much of this region, however, the upper part of the interval, which elsewhere contains much limestone, has been eroded. Along the western outcrop belt of interval E, the percentage of limestone decreases southward; limestone is largely replaced in the south by red sandstone and mudstone.

In areas where the limestone makes up less than 20 percent of the total, mudstone generally exceeds sandstone in ratios ranging from 1:1 to 4:1. At a few places, as in Kanawha County, W. Va., near the south border of interval E, the proportion of sandstone exceeds that of mudstone. At most of these localities, the upper part of interval E has been eroded, but the higher proportion of sandstone may also indicate a southern source of detrital sediment.

A large number of well records for deeper parts of the basin are of such poor quality that the information could not be used in making the lithofacies map (pl. 9B). Therefore, the distribution of control points used is less dense on this map than on the isopach map.

Stratigraphic units in the sequence assigned here to interval E were described in detail for Ohio by Condit (1912, p. 20) and by Stout (1939), for West Virginia by Reger (1931a), and for Pennsylvania by Berryhill (1963) and by Dutcher, Ferm, Flint, and Williams (1959). Upper Pennsylvanian and overlying Permian rocks at many places in the Appalachian region were described by Cross, Smith, and Arkle (1950) and by Arkle (1959).

Generalized geologic sections for interval E in Ohio, western Pennsylvania, and West Virginia contain great numbers of different rock units. From 9 to 12 sandstones, 13 to 23 mudstones, 7 to 10 limestones, and 11 to 17 coal beds are recognized in these States. The greatest number to be differentiated and named is in West Virginia.

The sandstones in interval E range from thin flat-beded fine-grained sheets to massive crossbedded, fine- to medium-grained lenses that contain a few quartz pebbles. The thin sandstones are commoner in Ohio and northern Pennsylvania, and the more massive ones are prevalent to the southeast in Pennsylvania and to the south in West Virginia. In the northern and western parts of the basin, sandstones are generally gray, but in the southern part, some are red brown (Arkle, 1959, p. 122-123; Berryhill, 1963).
Shaly mudstones overlie coal beds or coal horizons in many places and commonly are grayish black and carbonaceous. Many of those in the northern part of the region contain fossil-plant compressions. Other mudstones are generally gray and, in the northern part of the basin, are locally calcareous. Farther south most of the mudstones are red or mottled red and greenish gray. Root zones are found in underclays beneath some coal beds but are less numerous than in the older intervals in the same region.

Limestone units range from massive single beds, 2-3 feet thick, to 50-foot sequences of limestone intercalated with calcareous greenish-gray claystone and mudstone. Some beds are unfossiliferous, but others contain abundant invertebrates and fish remains. Limestone commonly occurs below coal or underclay.

Coal beds in the Monongahela Formation or Group are more persistent than most of those in the underlying Conemaugh, and one, the Pittsburgh coal bed, is probably more extensively mined than any other coal bed in the Appalachian field (Cross, 1954). The coal beds are generally divided by one or more detrital partings. These coals wedge out southward into red beds that characterize interval E in the southern part of West Virginia and Ohio.

Many coal and limestone units of interval E are sufficiently widespread to be easily traced and correlated in the northern part of the region, but lensing red mudstones and massive sandstones dominate the sequences to the south and make tracing impossible in many areas. Because of irregular lateral distribution sandstones are especially difficult to trace. Such difficulties apply to the limiting units of interval E, the Morgantown Sandstone Member of the Casselman Formation and the Waynesburg coal bed, and inconsistencies probably occur in picking these boundaries at some localities.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

**SOURCES**

Sandstones in interval E include both sheetlike horizontally bedded fine-grained sandstones and localized channel-type sandstones. Both types are recognized in nearly all parts of the basin, but little is known concerning the current directions indicated by their foreset beds. Although the direction of the crossbedding dips locally seems to be a reliable indicator of current direction, at some localities the orientation of crossbedding is recorded to vary as much as 180° (Berryhill, 1963, p. 80) in a single succession of beds.

Concentration of sand along the south border of the area of interval E occurrence in West Virginia and Ohio (pl. 9B) suggests that streams flowing into the basin had southern or southwestern sources (Arkl, 1959, p. 123).

In Belmont County, Ohio, patterns made by channel sandstones resemble the distributary-stream systems of deltas. The orientation of the channels indicates that the streams had source areas to the north and that they discharged southward into standing bodies of water. In this area Berryhill (1963, p. 87) identified a major southwestwardly oriented drowned valley that nearly paralleled the present Ohio valley and received drainage from tributaries from the north, east, and southeast. Southwestern Pennsylvania may have received drainage from the east, and West Virginia, principally from the southeast.

**ENVIRONMENTS OF DEPOSITION**

Throughout much of the Appalachian region one environment followed another in a rhythmic pattern that was repeated many times. During Late Pennsylvanian and Early Permian sedimentation, streams apparently introduced sediment from one or more directions as deltaic deposits that were built outward into standing bodies of water (Cross and others, 1950; Arkl, 1959; Beerbower, 1961; Berryhill, 1963).

Parts of interval E accumulated in lakes and ponds, and deltaic platforms provided the sites for abundant growth of land vegetation that subsequently was preserved in well-developed moderately extensive coal beds. The absence of coal and the occurrence of red beds in central West Virginia suggest that, in that region, land surfaces were dry enough for the oxidation of vegetable detritus and of clay minerals.

Limestone and associated calcareous fossiliferous mudstone formed in large bodies of freshwater. Occasional occurrences of the inarticulate brachiopod Lingula and the worm Spirorbis in the limestone beds suggest connections at times with a distant sea, or perhaps the survival of these animals in lakes or freshwater lagoons cut off from the sea.

The limestone beds of interval E probably formed at times of widespread marine transgression farther west (Cross and others, 1950; Berryhill, 1963). Because they represent some of the most extensive occurrences of freshwater-lake limestones in the United States, they are of particular interest. Suncracks and fragmentation of some beds by currents suggest that the depth of water in the lakes was generally less than 15 feet and that, periodically, the water drained away or evaporated. The lakes shifted position from time to time as a result of tilting or other deformation of the basin, fluctuation of water level, and filling of the lakes by deltaic sediment.

The cyclic character of the stratigraphic succession emphasized by Cross, Smith, and Arkl (1950), Arkl (1959), Beerbower (1961), and Berryhill (1963) suggests that some form of rhythm controlled the periodicity of sediment influx from surrounding highlands and the degree of spread of the lakes. The nature of cyclic
sedimentation is critically discussed by Beerbower (1961) for the Dunkard strata which overlie interval E and which have patterns of lithologic succession resembling those of the Upper Pennsylvanian, and by Berryhill (1963) for the Upper Pennsylvanian of Belmont County, Ohio.

Climatic changes, particularly in amount, type and distribution of rainfall, seem the most reasonable explanation for the cyclic repetition of lake and delta deposits in interval E. Expansion of the sea over large areas of the interior of the continent would result in increased rainfall and denser vegetation in the bordering land areas, and contraction of the sea would be expected to produce the reverse effects.

**PALEOTECTONIC IMPLICATIONS**

Downwarping in the northern Appalachian region was sufficient for the accumulation of several hundred feet of sediment of fluvial and lacustrine origin. As many as 16 cyclic alternations of beds are recognized in interval E (Reger, 1931a, p. 234; Berryhill, 1963). These are interpreted as the result of climatic cycles in a basin of relatively uniform subsidence. An intermittently accelerated influx of sediment may have resulted from the earth movements in surrounding highlands, or from climatic changes, as proposed by Beerbower (1961).

**TOTAL THICKNESS OF PENNSYLVANIAN ROCKS**

**THICKNESS TRENDS**

**NORTHERN APPALACHIAN AREA**

All intervals of the Pennsylvanian are represented, at least in part, in the northern Appalachian area. In eastern Ohio, southwestern Pennsylvania, and northern West Virginia, the Dunkard Group, referred to the Permian System, overlies rocks of Pennsylvanian age. Although a conspicuous unconformity occurs at the base of the Pennsylvanian, the upper limit of the system as used here is an arbitrary boundary at the base of the *Waynesburg coal bed* (Berryhill and Swanson, 1962). The Waynesburg Formation, which constitutes the lower part of the Dunkard Group above the definitely Pennsylvanian strata, is generally listed as Pennsylvanian and Permian in age, although the entire Dunkard Group is considered Pennsylvanian by some geologists (Clendenning, 1969).

Maximum thicknesses of Pennsylvanian strata are about 1,500 feet in Ohio; 1,500 feet in the center of the basin in southwestern Pennsylvania, where the system is overlain by the Dunkard Group; 1,400 feet in the small Broad Top field; 1,400 feet in the Georges Creek syncline, Maryland; 2,500 feet in the Northern Anthracite field; 2,600 feet in the Western Middle Anthracite field; and 5,100 feet in the Southern Anthracite field. The high rate of change in thickness from west to east (intervals A, B, and C) is shown diagrammatically in figure 10.

---

**Figure 10.**—Diagrammatic section showing relative thickness of Lower and Middle Pennsylvanian rocks (intervals A, B, and C) in the western and eastern parts of the northern Appalachian region. Contacts dashed where restored in areas of erosion of Pennsylvanian rocks; locality points are the same as on plate 1.

Except where the Dunkard Group overlies Pennsylvanian rocks, the present thicknesses of Pennsylvanian rocks generally do not represent original total thicknesses. The northern Appalachian area is maturely dissected throughout; narrow ridges commonly rise 500-1,000 feet above adjacent valleys. Pleistocene glacial drift overlies the Pennsylvanian in parts of northeastern Ohio and northern Pennsylvania. As most of the drill holes used in compiling the total isopach map (pl. 11) were not located on hilltops, many sections, indicated by the diamond pattern on the map, are combined from information obtained from one or more outcrops or coal-test borings and from deeper oil-exploration wells.

As shown on the isopach map (pl. 11), the thickness of the Pennsylvanian rocks in the northern Appalachian area ranges from zero at the margins of the basin to about 5,100 feet in the Southern Anthracite coal field near Pottsville, Pa.* The isopachs have been drawn at 500-foot intervals because local thicknesses are influenced by the depth of dissection. Thickness variations are less marked on the total isopach map than on the maps of the individual intervals because the lower intervals thicken greatly near the southern border of the northern Appalachian area, and the remnants of intervals D and E thin by erosion in the same localities where the lower intervals thicken. The map shows a gradual

---

*Thicknesses for the total Pennsylvania in the Anthracite fields, given here and in succeeding paragraphs as well as on plate 11, were supplied by J. B. Roen (written comm., 1970) and G. H. Wood, Jr. (written comm., 1971). Corresponding figures, supplied at an earlier date by H. R. Wanless, were approximately 3,600 feet for the Southern Anthracite field, 2,000 feet for the Western Middle field, and 1,300 feet for the Northern field.
southward thickening of the total Pennsylvanian in the main basin to about 2,500 feet along high ridges at drainage divides in Boone County, W. Va. Here intervals D and E are eroded, so the thick Pennsylvanian section is the result of the abrupt southward thickening of intervals A and B.

Nearly all this area was a subsiding platform during Pennsylvanian time, but strata in the Anthracite fields represent facies of the Appalachian miogeosyncline (geosynclinal sediments that have no associated volcanic rocks).

Along the north and west borders of the northern Appalachian area Pennsylvanian rocks are comparatively thin, and the 500-foot isopach is 20-70 miles from the margin of the area. Along the eastern margin of the field, and along the margins of the Chestnut Ridge, Laurel Hill, and Negro Mountain anticlines, and in the Georges Creek, Broad Top, and Anthracite fields, strata are commonly upturned sharply, and, therefore, thicknesses of from 500 to more than 1,000 feet may be found within a few miles of the present erosional margin. In the Southern Anthracite field, Pennsylvanian rocks are 5,100 feet thick, and many strata are steeply dipping or vertical. No attempt is made to show thickness variations along the margins of anticlinal folds.

In a large part of the Appalachian folded belt, much thicker and more complete sections of the Paleozoic sequence occur in synclines than in adjacent anticlines (Cooper, 1957). Differential downwarping of synclinal areas may have taken place frequently during Paleozoic time.

CENTRAL APPALACHIAN AREA

At no place in the area of southern West Virginia, eastern Kentucky, Virginia, and Tennessee are the Pennsylvanian rocks overlain by Permian or other younger strata, so they have everywhere been exposed to erosion; none of the thicknesses shown on the isopach map (pl. 11) represent the original thickness of the sediments. Interval E has been entirely eroded from this area, although it covers parts of West Virginia nearly to the border of northeastern Kentucky. Interval D is present in only parts of four counties in northeastern Kentucky. Interval C is absent in Tennessee. The central Appalachian area, therefore, is characterized by great thickening of intervals A and B, which together total less than 100 feet in parts of northern Ohio and Pennsylvania but increase to more than 5,000 feet in the mountains near Big Stone Gap, Va.10

In northeastern Kentucky, Pennsylvania rocks thicken gradually from the west border of the area eastward toward Louisa, Lawrence County, where, in the axis of the major syncline of the basin, most of interval D has escaped erosion, and the total section is about 1,400 feet thick.

Between the Paint Creek-Warfield fault zone and the Pine Mountain thrust fault area, the general succession of intervals A and B is thicker than in the northeastern Kentucky zone and reaches a maximum of about 3,000 feet in Pike County, eastern Kentucky. In southern West Virginia the maximum thickness is about 2,800 feet in Fayette County.

The Cumberland overthrust block shows a markedly thicker succession than that of the area northwest of the Pine Mountain fault (Wanless, 1946, p. 4). The block is believed to have moved northwestward at least 10 miles in an area where the section thickens abruptly, as shown by differences in the vertical distances between key beds on the two sides of this fault. The greatest thickness, more than 5,000 feet, is in Big Black Mountain on the Kentucky-Virginia line. The succession is also thick in the Log Mountains near the Kentucky-Tennessee line, where a maximum of about 4,000 feet is reported (Ashley and Glenn, 1906, p. 85-86).

Southwest of the Cumberland overthrust block is a rather small triangular area in Anderson, Campbell, Scott, and Morgan Counties, Tenn., in which strata are nearly flat lying to within 1 mile of the fault contact with overthrust Ordovician strata to the southeast. This deeply dissected area includes only strata of intervals A, and B. In Cross Mountain, the highest ridge, Pennsylvanian rocks are approximately 4,000 feet thick. At the southern margin of this area, interval B terminates in an escarpment, and west and southwest of the area the plateau is largely undissected and is capped by resistant sandstones of interval A, and the basal part of interval B.

In the Cumberland Plateau of southern Tennessee, strata dip gently southeastward from their northwestern margin to turn up abruptly on the northwest side of the faulted Sequatchie Valley anticline. The maximum thickness of Pennsylvanian strata occurs generally within 1-2 miles of the upturned margin of the Cumberland Plateau. Here the maximum thickness, composed wholly of interval A, is about 1,200 feet.

Walden Ridge, east of the Sequatchie Valley, is similar to the Cumberland Plateau in that strata dip southeastward from the margin of the valley nearly to the upturned edge at the southeastern margin of the coal field in Tennessee. Here the maximum thickness is about 1,800 feet. The same stratum caps Walden Ridge as caps the Cumberland Plateau, but the section thickens about 50 percent in the 12-15 miles of Walden Ridge.

---

10 Information received from R. L. Miller (written commun., 1971) after the author's completion of maps and sections indicated that the total thickness of Pennsylvanian strata for locality 94 near Big Stone Gap, Va., is approximately 5,600 feet, distributed as follows: Lee Formation, about 600 feet (interval A); Norton Formation, 900 feet, Gladstone Sandstone, 45 feet, and Wise Formation, 2,440 feet (interval B); Harlan Sandstone, about 600 feet (interval C).
SOUTHERN APPALACHIAN AREA

All Pennsylvanian strata of the southern Appalachian area are believed to belong to intervals A and B. Rocks of these intervals have a maximum thickness of more than 10,000 feet in Shelby County, Ala., and they are more than 12,300 feet thick in a well (loc. 113) in Winston County, Miss. Maximum thickness in northwestern Georgia is a little more than 1,000 feet.

In the northern part of the basin the thickness increases moderately at the rate of about 50 feet per mile. Farther south several wells in Chickasaw, Clay, and Lowndes Counties, Miss., show very abrupt thickening along a northwest trend at the rate of about 500 feet per mile. South of this hinge the base of the Pennsylvanian is not reached in any wells. The thickening south of the hinge line is at the rate of at least 100 feet per mile. Presumably, the thickening continues to an inferred thrust fault along the southern margin of the basin.

In the small synclinal outliers north of Rome, Ga., in Floyd and Chatooga Counties, Pennsylvanian strata are limited to mountain tops and attain a maximum thickness of about 300 feet (Hayes, 1902).

The Lookout Mountain syncline is about 85 miles long and 5-15 miles wide (McCallie, 1904). The maximum thickness of Pennsylvanian rocks is estimated to be about 1,100 feet near the axis of the syncline, but drill holes have not penetrated the entire section. Pre-Pennsylvanian strata crop out on the bordering escarpments of the mountain.

Because of a gentle east-west arch near the Tennessee-Alabama line, most of the Pennsylvanian strata have been removed by erosion from the southwestern extension of the Cumberland Plateau of Tennessee. This extension has been included within the Plateau coal field (McCalley, 1891). Pennsylvanian strata not exceeding 300 feet in thickness cap many plateau remnants. This arch, which extends about 80 miles from northeast to southwest, makes very difficult the correlation of rocks in Tennessee with corresponding rocks of the Black Warrior basin to the southwest.

A small area of Pennsylvanian strata is nearly isolated in a plateau remnant called Blount Mountain, a syncline which lies between two anticlines (Gibson, 1891). This plateau remnant contains about 4,220 feet of Pennsylvanian rocks (pls. 4A, 5A), which are exposed in the upturned flanks of the anticlines bordering the Blount Mountain plateau.

The Cahaba coal field is in the first major overthrust sheet southeast of the Black Warrior basin (Squire, 1890; Butts, 1907, 1911). The Pennsylvanian succession in the Cahaba coal field dips regularly southeast toward a bounding overthrust fault (fig. 7) and thickens from zero to more than 10,000 feet within the 8- to 12-mile width of the field. The sequence of Pennsylvanian rocks is so unlike the sequences in the adjoining Warrior and Coosa fields that correlation of coal beds and other key beds is difficult (Culbertson, 1964, p. B50).

Pennsylvanian rocks are exposed for about 90 miles along strike in the Cahaba coal field, and, at the south end of the exposed area, they are overlapped by Cretaceous rocks of the Gulf Coastal Plain. About 80 miles southwest, in Sumter County, Ala., a deep oil test well (loc. 134) appears to penetrate Pennsylvanian rocks below the Cretaceous and to extend 415 feet without reaching the base. Another well in Sumter County northeast of this one passes from Cretaceous strata directly into Ordovician(?), penetrating no Pennsylvanian rocks. These relationships suggest that the Cahaba field may extend continuously or with gaps for at least 80 miles beyond the Cretaceous overlap and that the structural high of the Birmingham area may separate the Warrior and Cahaba fields throughout that distance.

The Coosa coal field lies southeast of the Cahaba and is separated from it by another great overthrust (Gibson, 1895; Jones, 1929). The Coosa field includes several separate basins—the Yellowleaf to the south and the Wattsville and others to the north—separated by a narrow band of pre-Pennsylvanian rocks. The Pennsylvanian strata in the Coosa coal field have a marked southeasterward dip, as do those in the Cahaba field, and have a maximum thickness of probably 7,000 feet.

PENNYSylvANIAN COMPONENTS OF CRYSTALLINE BELTS, EASTERN UNITED STATES

A belt of crystalline metamorphic and plutonic rocks crops out southeast of the Appalachian coal field and the folded belt of the Valley and Ridge province, from Alabama northeastward to the Canadian border in northern New England. The New England mountains, Appalachian Mountains, Blue Ridge, Great Smoky Mountains, and Piedmont are parts of this crystalline belt. The rocks were long considered Precambrian in age and were thought to constitute an old mountainous upland—Appalachia.

More recently, discovery of Paleozoic fossils in the metamorphics in New England and elsewhere has led to increasing confidence that most of these were formed from Paleozoic sediments, volcanics, or plutonics, accumulated in rapidly subsiding eugeosynclines. Most crystalline rocks in New England are now considered Ordovician to Mississippian; Cambrian has been identified in the Virginia Piedmont; and Devonian and Pennsylvanian (?) fossils have been reported in the Talladega Slate of Alabama. During the late Paleozoic, local basins apparently continued to subside and accumulate thick successions of detrital sediments.

For many years the only "Pennsylvanian" rocks known in the crystalline belt have been phyllite near...
Worcester, Mass., and phyllite known as the Erin Shale in Clay County, Ala. In both places *Lepidodendron* has been collected in the phyllite. These plants, identified by David White, were listed as Carboniferous, probably Pennsylvanian in early reports by Butts (1926, p. 217-219) and by Prouty (1923).

Most of the Talladega Slate from Clay County, Ala., northeast to the Georgia line, an area 135 miles long and 14 miles wide, is largely Mississippian and Pennsylvanian. Maximum thickness of rocks in this area is as great as 28,000 feet according to Carrington (1967).

About 40 miles southeast of Birmingham, Ala., a sequence of metasedimentary rocks of the Talladega Slate, about 11,000 to 28,000 feet thick, crops out along a northeast-trending belt 100 miles long and 8-14 miles wide. A Carboniferous age for the predominant part of the Talladega may be valid (Carrington, 1967). Rocks lying northwest of known Devonian rocks in northern Chilton County probably are, for the most part, Mississippian and Pennsylvanian in age, according to Carrington (1967). The possibility of accurately dating these rocks by fossils is small because of metamorphism.

Immediately southeast of the Erin Shale in Clay County, and separated by a thrust fault, is a much more highly metamorphosed rock, the Ashland Mica Schist, which has generally been considered to be Precambrian. It contains abundant flake graphite, which is commercially produced. This highly carbonaceous metamorphic rock may have been derived from a Pennsylvanian coal-bearing sequence. Certain granites in North Carolina have tentatively been referred to Pennsylvanian age on the basis of potassium-argon dating.

Other rock units between the Canadian border in New England and the overlap of Cretaceous sediments in Alabama also include altered sedimentary rocks, volcanics, or plutonics, some of which may be of Pennsylvanian age, as indicated by fossils or geochemical dating. Areas in which such rocks may crop out are in Alabama, Georgia, North and South Carolina, Virginia, Maryland, Pennsylvania, New Jersey, southwestern New York, and the crystalline rock uplands of New England.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

**UNITS OVERLYING PENNSYLVANIAN**

In the north-central part of the Appalachian coal field, strata of the Dunkard Group directly overlie Pennsylvanian strata of the Monogahela Group (interval E). The Pennsylvanian-Pennsylvanian boundary as employed in this study follows usage in "Paleotectonic Investigations in the Permian System" (McKee, Oriel, and others, 1967), where the system boundary is arbitrarily placed at the contact between the Dunkard and Monongahela Groups. The base of the Waynesburg coal, which is relatively widespread in the region, is considered to be the top of interval E of the Pennsylvanian System. Thus, no unconformity between the two systems is recognized in this region.

Depression of the area containing the Dunkard Group seems to have taken place as early as interval D of the Pennsylvanian, when a centripetal drainage developed into the basin from all sides (Morris, 1967). Lacustrine sediments in both the Monongahela and Washington Formations show that basin development continued from Pennsylvanian into Permian time.

In most of the Appalachian coal field north of Alabama and Mississippi, Pennsylvanian rocks are exposed at the surface, but at the north end of the geosyncline, in Pennsylvania and Ohio, Pleistocene glacial deposits overlie them. Typical examples are in the Northern Anthracite field and in various other small synclinal outliers where glacial drift covers the surface. A southwestward trend in glacial deposits continues across Ohio, in the central part of which a western limit of Pennsylvanian rocks closely approximates the east border of Pleistocene glaciation (Flint and others, 1959).

In the Cahaba and Warrior coal fields of Alabama, rocks of Pennsylvanian age pass beneath Upper Cretaceous rocks near the latitude of Tuscaloosa, Ala., and along an irregular boundary extending northwestward toward the northwest corner of the State. The Upper Cretaceous Tuscaloosa Formation was deposited with northward onlap across older Paleozoic rocks during downwarping of the Gulf of Mexico embayment. In the northern part of the Black Warrior basin in Mississippi, strata of Pennsylvanian age directly underlie Tuscaloosa strata, but farther south the Lower Cretaceous Hosston Formation occurs between rocks of Pennsylvanian age and the Tuscaloosa, as shown in several wells.

**PALEOTECTONIC IMPLICATIONS**

Sedimentation apparently continued without interruption from Pennsylvanian into Permian time in the northern Appalachian region and probably elsewhere. The Appalachian region was folded in Permian time, and the Permian Dunkard strata were removed from anticlinal areas but survived in two synclines east of the basin—one in Pennsylvania and the other in the Georges Creek syncline in Maryland (O’Harra, 1900, p. 128-129). The former extent of these rocks north and south of their present limits is unknown.

In late Paleozoic time the south end of the Black Warrior basin of Alabama was deformed probably by overthrusting; subsequently the region was beveled by erosion. Jurassic (McKee and others, 1956) and Lower Cretaceous rocks were deposited progressively northward on a level surface that had been cut across
the deformed lower Paleozoic rocks. Subsequent history has been marked by continued down warp of the Gulf of Mexico basin; thus, not only the top of the Paleozoic but also the Mesozoic and younger rocks all increase in depth southward toward the Gulf.

REFERENCES


Andrews, E. B., 1874, Report on second district—Surface geology of southeastern Ohio; geology of Washington County, Noble County, Guernsey County (southern half); Belmont County (southern half), Monroe County, Pickaway and Fairfield Counties: Ohio Geol. Survey Rept. 2, pt. 1, Geology, p. 439-600.


Northern anthracite field, Parts 1-6: Pennsylvania Geol. Survey, ser. 2, v. AA.


Cathey, B., 1955, Geology and mineral resources of the Newburgh
1880, Preliminary report on the geology of Morgan, Johnson, Magoffin, and Floyd Counties, with map: Kentucky Geol. Survey Rept. Progr. 6, p. 315-338.
1961, Well-sample descriptions in northwestern Penn-
Fette, C. R., and Stephenson, R. C., 1946, Oil and gas developments in the
North Strabane area, Washington County, Pennsylvania with a chapter on Core analysis determinations of diamond core from
6120, 401 p.
Flint, N. K., 1951, Geology of Perry County: Ohio Div. Geol. Survey
Bull. 48, 234 p.
Flint, R. F., and others, 1959, Glacial map of the United States east of the
Rocky Mountains: Geol. Soc. America.
Fols, F. J., 1912, Coals of the region drained by the Quicksand Creeks in Breathitt, Floyd, and Knott Counties [Kentucky]: Kentucky
Fontaine, W. M., and White, I. C., 1880, The Permian or upper Carboniferous flora of West Virginia and southwestern Pennsyl-
Freeman, L. B., 1951, Regional aspects of Silurian and Devonian stratigraphy in Kentucky: Kentucky Geol. Survey, ser. 9, Bull. 6, 565 p.
Gardner, J. H., 1912, Preliminary report on the economic geology of the
Hartford quadrangle: Kentucky Geol. Survey Bull. 20, ser. 27, 33 p.
____, 1913, The Broadtop coal field of Huntington, Bedford, and
Gibson, A. M., 1891, Report on the coal measures of Blount County, in
215.
____, 1895, Report upon the Coosa coal field: Alabama Geol. Survey.
143 p., maps.
Giles, A. W., 1921, The geology and coal resources of Dickenson County,
Virginia: Virginia Geol. Survey Bull. 21, 224 p.
____, 1925, The geology and coal resources of the coal-bearing portion of
Glenn, L. C., 1903, Devonic and Carbonic formations of southwestern
____, 1922, The geology and coals of Webster County: Kentucky Geol.
Survey, ser. 6, v. 5, 249 p.
____, 1925, The northern Tennessee coal field included in Anderson,
33-B, 478 p.
Graeber, C. K., and Foose, R. M., 1942, Geology and mineral resources
Gray, L. R., 1965, Palynology of four Allegheny coals, northern App-
Grimsley, G. P., 1907, Ohio, Brooke, and Hancock Counties: West
____, 1910, Pleasants, Wood, and Ritchie Counties: West Virginia
Geol. Survey, County Repts. and Maps, 352 p.
Gwinn, V. E., 1964, Thin-skinned tectonics in the plateau and northwestern valley-and-ridge provinces of the central Ap-
Habib, Daniel, 1965, Distribution of miospore assemblages in the
Hagan, W. W., 1942, Geology of the Cub Run quadrangle, Kentucky:
Harmsberger, T. K., 1919, The geology and coal resources of the coal-
bearing portion of Tazewell County, Virginia: Virginia Geol.
Hauser, R. E., 1932, Geology and mineral resources of the Paintsville
quadrangle: Kentucky Geol. Survey, ser. 9, Bull. 13, 80 p.
Hayes, C. W., 1902, Description of the Rome quadrangle, Georgia-
____, 1911, Wirt, Roane, and Calhoun Counties: West Virginia Geol.
Survey, County Repts., 573 p.
____, 1912, Doddridge and Harrison Counties: West Virginia Geol.
Survey, County Repts., 712 p.
____, 1917, Braxton and Clay Counties: West Virginia Geol. Survey,
County Repts., 883 p.
Hennen, R. V., and Gathrop, R. M., 1915, Wyoming and McDowell
Hennen, R. V., and Reger, D. B., 1913, Marion, Monongalia, and Taylor
____, 1914a, Preston County: West Virginia Geol. Survey, County
Repts., 556 p.
____, 1914b, Logan and Mingo Counties: West Virginia Geol. Survey,
County Repts., 776 p.
Hickok, W. O., 4th and Moyer, F. T., 1940, Geology and mineral resources
Hinds, Henry, 1918, The geology and coal resources of Buchanan
Hodge, J. M., 1910, Report on the coals of the Three Forks of the Kent-
cucky River: Kentucky Geol. Survey Bull. 11, 280 p.
____, 1912, Report on the upper Cumberland coal field—The region
drained by Poor and Clover Forks in Harlan and Letcher Coun-
____, 1918, The coals of Goose Creek and its tributaries: Kentucky
Huddle, J. W., and Patterson, S. H., 1961, Origin of Pennsylvania under-
Hughes, H. H., 1933, Freeport quadrangle: geology and mineral resources:
Hunt, C. B., Briggs, G. H., Jr., Munyan, A. C., and Wesley, G. R., 1937,
Coal deposits of Pike County, Kentucky: U.S. Geol. Survey Bull.
876, 92 p.
Hutchinson, F. M., 1912, Report on the geology and coals of the Cen-
tral City, Madisonville, Calhoun, and Newburg quadrangles in
Muhlenberg, Hopkins, Ohio, McLean, Webster, Daviess, and
Ishii, Kenichi, 1957, On the so-called Frasulina: Proc. Japan Acad., v. 33,
no. 10, p. 652-656.
Jillson, W. R., 1919, The Kendrick shale; a new calcareous fossil
horizon in the coal measures of eastern Kentucky: Kentucky Dept. Geol. and Forestry, ser. 5 [of Kentucky Geol. Survey], Mineral and
Forest Resources of Kentucky, v. 1, no. 2, p. 96-104.


———, 1898, Map of the Warrior coal basin with columnar sections of formations so far as it carries workable coals: Alabama Geol. Survey.


McKee, E. D., Oriel, S. S., and others, 1967, Paleotectonic in-


Miller, A. M., 1919, Coals of the lower measures along the western border of the eastern coal field: Kentucky Geol. Survey Bull. 12, 83 p.


O'Harra, C. C., 1900, The geology of Allegany County: Maryland Geol. Survey, Allegany County, p. 57-163.


Prouty, W. F., 1923, Geology and mineral resources of Clay County, with special reference to the graphite industry: Alabama Geol. Survey County Rept. 1, 190 p.


Rothrock, H. E., 1949, Geology and coal resources of the northeast part of the Coosa coal field, St. Clair County, Alabama: Alabama Geol. Survey Bull. 61, pt. 1, 163 p.


1918, Geology of Muskingum County: Ohio Geol. Survey Bull. 21, 351 p.


Swartz, C. K., 1922, Distribution and stratigraphy of the coal measures of Maryland; Correlation of the coal measures of Maryland; The coal basins of Maryland: Maryland Geol. Survey, v. 11, p. 35-126.


Welch, S. W., 1959, Mississippian rocks of the northern part of the Black Warrior basin, Alabama and Mississippi: U.S. Geol. Survey Oil and Gas Inv. Chart OC-62.


Michigan Basin Region

By HAROLD R. WANLESS and GERALD L. SHIDELER

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN
THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES
OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-D
CONTENTS

Abstract ..................................................... 63
Region defined .................................................. 63
Paleogeology ................................................... 64
Units underlying Pennsylvanian System .................. 64
Paleotectonic implications ................................. 64
Interval A ...................................................... 64
Formations included ............................................ 64
Upper boundary of interval A ............................... 64
Thickness trends ............................................... 65
Lithofacies trends ............................................ 65
Sources and environments of deposition .................. 65
Paleotectonic implications .................................. 66
Interval B ...................................................... 66
Formations included ............................................ 66
Stratigraphic relations ....................................... 66
Upper boundary of interval B ............................... 67
Thickness trends ............................................... 67
Lithofacies trends ............................................ 68
Sources and environments of deposition .................. 69
Paleotectonic implications .................................. 69
Interval C ...................................................... 69
Formations included ............................................ 69
Stratigraphic relations ....................................... 69
Upper boundary of interval C ............................... 70
Thickness trends ............................................... 70
Lithofacies trends ............................................ 70
Sources and environments of deposition .................. 70
Paleotectonic implications .................................. 70
Total thickness of Pennsylvanian rocks .................. 70
Geologic units directly above Pennsylvanian System .... 70
Units overlying Pennsylvanian System .................... 70
Paleotectonic implications .................................. 70
References ...................................................... 70

ILLUSTRATION

[For listing of plates (in separate case) see volume "Contents"]

Figure 11. Map showing geographic features and localities and Pennsylvanian tectonic elements in the Michigan basin region mentioned in the text ........................................... 64

TABLE

Table 10. Positions of stratigraphic units mentioned in text but not shown on plate 13: Intervals A, B, and C, Michigan basin .......... 65
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

MICHIGAN BASIN REGION

By Harold R. Wanless and Gerald L. Shideler

ABSTRACT

The Michigan basin was a subsiding area at various times during the Paleozoic Era. It evidently was folded shortly before the beginning of Pennsylvanian sedimentation, in the time of subinterval A. The Bayport Limestone of Late Mississippian (Meramec) age, which extensively underlies Pennsylvanian strata in the basin, was eroded in places, exposing the underlying Michigan Formation of Meramec and Osage age. In addition, the Howell anticline, which constitutes one of the major structural features within the Michigan basin, was elevated and stripped of both the Bayport and Michigan Formations, as well as the underlying Marshall Formation, except along its margins. The Coldwater Shale of Early Mississippian (Kinderhook) age underlies the central part of the structure. This elevated area was not buried until 400 feet of Lower Pennsylvanian sediment had accumulated around it.

The Pennsylvanian section of Michigan appears to be comprised predominantly of alluvial and deltaic deposits. Sediment sources for this basin appear to have been older Paleozoic strata, as indicated by the quartzose sandstones which contain frequently rounded and some frosted sand grains, and a stable heavy-mineral suite.

During interval A time (approximately late Morrow), a river system from the northeast reached the northeast border of the basin near the present site of Saginaw Bay and distributed sediments along the east and northeast sides. The upper surfaces of sandstones slope to the west and southwest, apparently reflecting the regional palaeoslope on which a delta complex prograded across the basin. No outlet for sediment introduced into the basin seems to have been available. Early in interval A, a fusulinid limestone was deposited, which suggests a marine connection with western seas. During and after the time of interval A, local deposits of dolomite, gypsum, and red mudstones formed in the western and northernwestern parts of the basin; presumably at times of low precipitation, the waters in remnant pools became hypersaline. Coal accumulated along the east and south margins of the basin. More than 500 feet of detrital sediment accumulated along the courses of distributary channels on the east and northeast sides of the basin.

During deposition of interval B, probably mainly in Atoka time, much less sand was introduced, and the abundance of dark-gray to black mudstone indicates that the submerged parts of the basin were poorly aerated. Several thin coal beds are present. One marine limestone is present, the Verne Limestone Member of the Saginaw Formation, and marine or brackish-water faunas are also found within mudstones in five of seven cyclic sequences of rocks. The limestone appears to be a correlative of the Sevillo Limestone Member of the Spoon Formation in Illinois and of the Lower Mercer Limestone Member of the Pottsville Formation in Ohio, indicating that marine straits, in some way, connected these three coal basins late in interval B time. The connecting deposits, now eroded, probably were in northwestern Indiana. Maximum thickness of interval B is 184 feet.

Massive sandstones interval C (lower Des Moines) resemble those in interval A. Erosion has removed at least the upper part of the original deposits of interval C throughout the basin, and the entire interval is missing in much of the basin except in two belts that trend west and southwest from the head of Saginaw Bay. Intervals B and C tend to be thick in areas where interval A is thin.

In post-Pennsylvanian time, the Pennsylvanian section was extensively eroded. During the Jurassic Period, red sandstone, red mudstone, and gypsum deposits formed on a plain of low relief, particularly on the west side of the basin. Later, a second erosional surface was developed and subsequently covered with unusually thick deposits of Pleistocene glacial drift, attaining a maximum thickness of 900 feet in the northwestern part of the basin.

REGION DEFINED

The Pennsylvanian rocks of the Michigan basin, which are in the central part of the southern peninsula of Michigan, cover an area having maximum dimensions of 145 miles north-south and 130 miles east-west. The basin as defined by the Pennsylvanian rocks is, in general, slightly elliptical, shallow, and structurally simple. Its most notable intrabasinal structure is the northwest-trending Howell anticline on its southeast margin. The basin presumably was bordered on its western flank during much of Pennsylvanian time by a shallow shelf and coastal plain. However, these marginal features, if once present, have been obscured by post-Pennsylvanian erosion which has completely isolated Pennsylvanian strata in Michigan from Pennsylvanian deposits of adjacent areas. Maximum thickness of Pennsylvanian strata remaining within the basin is approximately 700 feet.

Exposures of Pennsylvanian strata are extremely limited in the Michigan basin because of the thick cover of Pleistocene glacial drift. The principal outcrop area is near the town of Grand Ledge, Eaton County (Kelly, 1933). Most of the information about thickness and lithology of the Pennsylvanian rocks has been derived from sample logs of oil-test borings in the subsurface-geology files of the University of Michigan.


**PALEOGEOLOGY**

**UNITS UNDERLYING PENNSYLVANIAN SYSTEM**

Strata that directly underlie the Pennsylvanian System in the Michigan basin are entirely Mississippian in age (pl. 2). Across much of the central part of the basin and along parts of its rim the Bayport Limestone, correlated with the St. Louis Limestone (Meramec age) of the midcontinent region, is the immediate substratum. In scattered patches throughout the central basin and along much of its rim, the Michigan Formation of Meramec and Osage age is the substratum. This formation elsewhere immediately underlies the Bayport and is believed to be equivalent to the Salem and Warsaw Limestones (Meramec age) and Keokuk and Burlington Limestones (Osage age). Except in a few scattered localities, the only older substrata directly beneath the Pennsylvanian rocks are in the area of the Howell anticline (fig. 11), on the southeast flank of the basin. Along the flanks of the anticline, as well as in the scattered localities, the Napoleon Sandstone Member of the Marshall Formation of early Osage (Fern Glen Limestone equivalent) age underlies the Pennsylvanian, whereas in the center of the uplift the Coldwater Shale of Kinderhook age is the substratum.

**PALEOTECTONIC IMPLICATIONS**

After deposition of the Bayport Limestone, the basin underwent slight deformation. The Bayport was stripped from anticlinal folds in parts of the basin, thus exposing the underlying Michigan Formation. Along the Howell anticline about 1,000 feet of structural movement occurred and the elevation was sufficient to permit removal of the Bayport and Michigan Formations and the Napoleon Sandstone Member of the Marshall Formation to a depth of 500-800 feet. At the beginning of Pennsylvanian sedimentation during Morrow time, the anticlinal area apparently stood above the surrounding depositional plain as a monadnock several hundred feet high, but it was buried by the end of interval A deposition.

**INTERVAL A FORMATIONS INCLUDED**

The lower part of the Saginaw Formation in the Michigan basin, from a few feet above the **Saginaw coal** down to the base of the Pennsylvanian section, is assigned to interval A (pl. 13; table 10). The name Parm a is sometimes applied to a sandstone unit in the basal part of the Saginaw Formation, and the unit is now considered to be a member of the Saginaw.

In the Saginaw area on the east side of the basin, coal operations were formerly carried on by shaft mining. In that area, numerous coal test borings were put down, and two important coal beds were named, the **Saginaw (lower)** and **Verne** (table 10). In some places, the **Verne coal bed** has a roof of fossiliferous marine limestone.

Fossil floras, collected from both outcrops and roof shales of coal beds formerly mined underground in the Saginaw area, have been described by Arnold (1949). He recognized a succession of three floral zones, of which the lowest is in strata associated with the **Saginaw coal bed**. This flora contains Neuropteris schlehani, N. saginawensis, Aulacotheca campbelli, Sphenopteris artemisiaefolioides, and Diplophlema obtusiloba. It is correlated by Arnold (1949, p. 153) with floral zone 6 of Read and Mamay (1964), formerly called zone 3 of Read (1947). This zone occurs in the upper part of the New River Formation of West Virginia and Kentucky (subinterval A₂, pl. 5). No floras have been discovered lower than the **Saginaw coal bed**.

**Paramillerella** sp., found in a limestone near the base of the Saginaw Formation, apparently is of the same form as that found in limestones of Morrow age in Wyoming and Montana.

**UPPER BOUNDARY OF INTERVAL A**

The boundary between interval A and interval B is established a few feet above the top of the **Saginaw coal bed**, or at an estimated equivalent position in parts of the basin where the coal bed seems to be absent. Coal test borings in Saginaw County indicate that the **Saginaw coal bed** is stratigraphically near the base of a predominantly dark mudstone sequence that composes the upper part of the Pennsylvanian section throughout
TABLE 10.—Positions of stratigraphic units mentioned in text but not shown on plate 13: intervals A, B, and C, Michigan basin region

<table>
<thead>
<tr>
<th>Interval</th>
<th>Formation</th>
<th>Named unit</th>
<th>Cyclical formations of Kelly (1933, 1936)</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>Grand River</td>
<td>Verne limestone</td>
<td>Grand Group</td>
</tr>
<tr>
<td></td>
<td>Formation</td>
<td>Member</td>
<td>H</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>Saginaw</td>
<td>Verne coal</td>
<td>G</td>
</tr>
<tr>
<td></td>
<td>Formation</td>
<td>Group</td>
<td>F</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>E</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>D</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>C</td>
</tr>
<tr>
<td>A</td>
<td>Saginaw</td>
<td>coal</td>
<td>B</td>
</tr>
<tr>
<td></td>
<td>Sandstone</td>
<td>Member</td>
<td></td>
</tr>
</tbody>
</table>

much of the Michigan basin. The Verne and other thin coal beds occur within this dark mudstone sequence and are assigned to interval B, although the coals themselves are rarely recorded in sample logs of exploration wells. The base of the dark mudstone sequence served effectively as the upper boundary of interval A in parts of the basin lacking the Saginaw coal (Shideler, 1965). The roof limestone of the Verne coal bed contains a species of Fusulinella that indicates rocks of Atoka rather than Morrow age and sets an upper limit to the zone within which the Morrow-Atoka boundary can be placed.

THICKNESS TRENDS

Interval A ranges in thickness from less than 30 feet near the margins of the basin to a maximum of 563 feet at locality 147 in Midland County (pl. 3A). Isopachs, drawn at 100-foot intervals, show that the thickness increases from the margins toward the north-central part of the basin. This suggests a roughly bowl-shaped basin at the beginning of interval A time, surrounded by a shallow rim on which very little sediment accumulated.

The most abrupt thickness change in interval A is in southern Shiawassee County near the southeastern margin of the basin. Interval A is absent at locality 285 on the crest of the Howell anticline, but is represented by 491 feet at locality 287 in the basin, about 6 miles to the northwest. This contrast shows that the anticline existed as a topographic high when Pennsylvanian sedimentation began and was nearly buried by sediment at the end of interval A time.

Other thickness irregularities shown by the map suggest that the interval attained a considerable thickness in areas extending westward from the present Saginaw Bay. The broad border where the interval is thin on the west, north, and southeast margins of the basin is at a maximum distance from the source of sediment. Where the interval is thick within the basin, the deposits are largely sandstone that accumulated along the main lines of transport; more muddy, thinner sequences in the basin probably were differentially thinned by compaction. Thickness values demonstrate that interval A composes most of the Pennsylvanian section in the Michigan basin.

LITHOFACIES TRENDS

In most of the Michigan basin, interval A consists of more than 80 percent detrital rock, ranging from fine mudstone to slightly pebbly sandstone; sandstones are the predominant lithology (pl. 3B). Coal is present largely in the eastern and southern parts of the basin. Subordinate quantities of red beds, gypsum, dolomite, and limestone are also present, largely in the northwestern part of the basin. The lithofacies map (pl. 3B) exhibits an overall distribution pattern which appears to be random, illustrating the lithologic variability of interval A.

In many wells, the basal unit of interval A is the Parma Sandstone Member of the Saginaw Formation. It is 5-50 feet thick and is overlain by a brownish-gray dense marine limestone 5-10 feet thick. The sandstone is clean, quartzose, medium to coarse grained, and in many places crossbedded; locally, it is conglomeratic (Kelly, 1936).

SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES

Crossbedding measurements from the basal Pennsylvanian sandstones (Potter and Siever, 1956, p. 236) indicate that the major direction of current flow was southwestward from the Saginaw Bay region, but minor westward flow also occurred, as well as minor southward flow at the southern tip of the basin. Isopach patterns and sandstone dispersal patterns (Shideler, 1969) also indicate a southwesterly paleocurrent direction, with the primary sediment source located northeast of Saginaw Bay.

Most of the sandstone beds contain a high percentage of rounded quartz grains, some of which are frosted. In addition, they contain a stable heavy-mineral suite, consisting mainly of tourmaline and zircon. These properties indicate that the source materials were probably older Paleozoic arenites, such as the Cambrian-Ordovician sandstones or, possibly, sandstones of the Marshall Formation of Mississippian age. Such sources were available to the northeast, in areas now inundated by Lake Huron or in areas of sedimentary rock in Ontario south of the Canadian Shield. It is believed that the basin trapped and held most of the sediment introduced into it; little detrital sediment escaped to areas farther south.
ENVIROMENTS OF DEPOSITION

The thin sequence of strata composing interval A along the eastern border of the Michigan basin probably includes channel sandstones, flood-plain mudstones, and the deposits of coal swamps. The present upper surface of the major sand wedge making up interval A is inclined westward, and the sandstone beds thicken abruptly southwestward from the basin margin. This trend suggests that alluvial deposits in the northeast merged into deltaic deposits in the central part of the basin. The delta complex apparently prograded westward and southward across the basin. Before the principal development of the delta complex, some sand may have been distributed by waves or currents across the basin floor, possibly in a shallow-shelf or littoral environment, to form the basal Parma Sandstone Member.

Limestone immediately overlying the basal sandstone in the western, southwestern, and northwestern parts of the basin contains Paramillerella sp., indicating that a marine transgression took place across the basin. In the east, the Saginaw coal bed is more or less continuous above the detrital wedge in the shoreward part of the basin. It may have been deposited almost contemporaneously with the marine limestone in the west. The coal and limestone beds mark a time when little detrital sediment was introduced into the basin.

Fingerlike distributaries of the delta complex spread across the basin during interval A time, and large ponds and lagoons were cut off from the open sea. Within some of these lagoons in western parts of the basin very dark gray mudstone was deposited; in other ponds and lagoons, salinity increased because of evaporation, and carbonate and gypsum were deposited. The red beds in interval A, in some places, may have resulted from the deposition of highly oxidized detritus in slightly better drained parts of the basin; however, oxidation after deposition is also a possibility, as proposed for red beds elsewhere (Walker, 1963).

Regular cyclical successions are not obvious within interval A in the Michigan basin, as they are in other parts of the United States.

PALEOTECTONIC IMPLICATIONS

Slight subsidence probably occurred during deposition of interval A, as shown by depositional thickening westward of the strata below the Saginaw coal bed. The coal may have been deposited on a gently westward sloping surface, rather than on a level plain. Certainly differential movement during, and since, interval A time has been slight.

INTERVAL B

FORMATIONS INCLUDED

Dark mudstone and coal, referred to interval B in the Michigan basin, belong entirely to the upper part of the Saginaw Formation. The interval extends from a few feet above the Saginaw coal bed to a few feet above the Verne Limestone Member of the Saginaw. Virtually all of interval B crops out near Grand Ledge, where Kelly (1933) recognized eight cyclical formations, which he designated as formations A through H (table 10). Of these formations, A through G represent interval B; H is the basal sandstone of interval C.

STRATIGRAPHIC RELATIONS

Interval B is distributed as a rather thin assemblage throughout most of the Michigan basin. It is believed to be absent in Kent and parts of Newaygo and Montcalm Counties (fig. 11) on the western side of the basin. Interval B is also probably absent in Bay, Saginaw, Gratiot, and Ingham Counties, as well as in other smaller areas on the east side of the basin.

The thin succession in the upper part of the Saginaw Formation, here referred to interval B, is so identified on the basis of the following information. The Saginaw coal is placed near the top of interval A because the roof mudstone of the coal contains plants identical to those found in the upper part of the Lee Formation in eastern Kentucky (Arnold, 1949, p. 159). The Lower Verne coal bed, 120-150 feet above the Saginaw coal bed, is locally overlain by the marine Verne Limestone Member of the Saginaw Formation. This limestone unit contains Fusulinella sp., believed to be F. iovensis. The Verne has been correlated with an impure limestone in outcrop near Grand Ledge, described by Kelly (1933, p. 85). Kelly correlated the fauna in this limestone with the faunas in the Mercer Member of the Pottsville Formation of Ohio and the Seville Limestone Member of the Spoon Formation of Illinois, both units assigned to interval B.

Neither the Saginaw coal in interval A nor the limestone above the Verne coal in interval B is widespread in the Michigan basin; however, the strata between the two coal beds, consisting largely of dark-gray to black nonfissile mudstone, are shown by subsurface data to be widespread throughout the basin. The mudstone sequence overlies predominantly sandy rocks of interval A and underlies another succession in interval C that contains a large proportion of sandstone. The dark mudstone sequence was the principal basis for dividing strata of the Michigan basin into intervals A, B, and C. However, if the Fusulinella sp., found in the Verne Limestone Member, is a more primitive form than F. iovensis, some of the strata above the Verne Limestone Member, here referred to interval C, are probably Atoka in age and on this basis should be transferred to interval B. Arnold (1949, p. 180) listed species of plants, which he termed the "intermediate" flora, from strata near the Lower Verne coal. These plants correspond with plants of floral zone 8 of Kanawha age (Read and Mamay, 1964) and thus support the assignment of strata containing them to interval B.

Interval B in the Michigan basin is believed to corre-
spond approximately with rocks of Atoka age in the midcontinent region.

**UPPER BOUNDARY OF INTERVAL B**

In the general absence of the *Verne Limestone Member*, the top of interval B is established at the base of a major sandstone bed that has been referred to the Grand River Formation. The Grand River unconformably overlies the Saginaw Formation. At many places, the Grand River Formation is absent, and interval B is directly overlain by Upper Jurassic red beds and gypsum (A. T. Cross, written commun., 1964) or by Pleistocene glacial drift.

**THICKNESS TRENDS**

The range of thickness for interval B is from zero to 184 feet at locality 378 in Lake County (pl. 6A). Key beds for correlating are absent throughout much of the basin, and numerous inconsistencies probably occur in determining the thickness of interval B.

Variations in thickness of interval B are due principally to: the tendency for interval B to be thicker in areas where interval A is shaly and thin; and erosion of the upper parts of interval B prior to deposition of channel sandstones of interval C, Upper Jurassic red beds, or Pleistocene drift. The floor of the Michigan basin evidently had much less relief during deposition of interval B than it had at the beginning of Pennsylvanian sedimentation.

**LITHOFOACIES TRENDS**

Interval B throughout most of the Michigan basin contains more than 80 percent detrital rocks. Mudstones in the interval exceed sandstones in thickness by more than four to one throughout much of the basin. Some linear belts in the eastern and southern parts are between 50 and 80 percent mudstone. A few very small areas contain more sandstone than mudstone, but such areas generally are identified on the basis of a single well which happened to encounter a channel sandstone in interval B. The facies depicted for the interval on plate 6B are the averages of units of widely diversified lithology, as is illustrated by exposures near Grand Ledge. The interval there consists of three distinct sandstones, five mudstones, six underclays, five coals, and one limestone.

The lithology of interval B in the northern and northwestern parts of the basin contrasts with that in the southern and southeastern parts. Red beds are reported in the north and northwest, most generally in Mecosta and Lake Counties; and gypsum interbedded with the red beds is recorded in a few well logs. Limestone and dolomite are also reported from sporadic wells, largely in the northwestern part of the basin; these carbonate units are thicker than the *Verne Limestone Member* near the top of interval B in the eastern and southern parts of the basin.

Small areas, particularly in Lake, Newaygo, Wexford, and Missaukee Counties, have more than 20 percent nondetrital sediments. The abnormal succession of carbonate rocks in the northwestern counties is nowhere exposed and is known only from sample logs and drillers’ logs.

In the southern and eastern parts of the Michigan basin, coal is a minor but characteristic component of interval B, although none of the coal beds in interval B is as widespread as the *Saginaw coal* of interval A. Most interval B coals are no more than 2 feet thick and are principally in Saginaw and Ingham Counties, which are the areas having the greatest number of coal test borings records.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

A sand isolith map for interval B shows that sandstone more than 40 feet thick occurs mainly at the northern margin of the basin and along a belt extending southwestward from the south end of Saginaw Bay to Shiawassee County (Shideler, 1969, p. 1234). Throughout much of the central part of the basin, interval B contains no sandstone. Sources of sediment, therefore, appear to have been largely to the north and northeast of the basin. The sandstones of interval B, like those of interval A, are highly quartzose and have a stable heavy-mineral suite; they are believed to have been derived from older Paleozoic arenites rather than from crystalline rocks of the Canadian Shield.

Interval B contains much less sandstone than intervals A and C. Sandstones in all three intervals are similar in petrology and they probably had a common source. The smaller proportion of sandstone in interval B indicates either that the source areas had lower relief than during deposition of intervals A and C or that erosion of sand from the source areas decreased because of a change in climate. In addition, if the source areas were densely covered with vegetation, there would have been less erosion than if soils and subsoils were directly exposed. Fairly numerous coal beds in parts of interval B suggest abundant vegetation.

**ENVIRONMENTS OF DEPOSITION**

Basin filling continued during interval B time. The finer average grain size of the introduced sediment indicates that currents bringing in sediment were less vigorous than during interval A time. Mudstone and shaly limestone probably tended to accumulate in the minor depressions left between fluvial or deltaic sand deposits of interval A. The depositional environment is envisioned as a broad alluvial plain, which was periodically inundated by restricted seas (Shideler, 1965, 1969). The discontinuity of such marine or brackish-water units as the *Verne Limestone Member* and the upper and lower Lingula mudstones of Kelly (1936) near Grand Ledge suggests that during inundations the sea spread
through the tortuous channels of estuaries leaving many islands and shoals and did not at times cover the entire basin.

Much of the sediment accumulated in standing water; the predominant dark color of the mudstone suggests poor circulation in the basin. On the other hand, oxidizing conditions are suggested by red beds in the western and northwestern parts of the basin. Oxidation there may have been due to subaerial exposure, or, alternatively, it may have been diagenetic. If the oxidation were diagenetic, however, much of the organic material in dark mudstones associated with the red beds should also have been oxidized.

A paleoslope that was inclined southwestward persisted during interval B time; most coal beds formed in the eastern part of the basin, relatively high on the paleoslope. During the time that limestones of the Verne in the Michigan basin, the Seville of northern Illinois, and the Mercer of northern Ohio were deposited, the seas in all these areas were probably interconnected, but the exact positions of the seaways are not known.

Small isolated depressions in the northwestern part of the Michigan basin, from which sea water was evaporated, are shown by deposits of dolomite and gypsum that are locally present.

A pulsatory fluctuation in sea level that produced four cycles of marine transgression and regression is evident. In the outcrop of interval B near Grand Ledge, Kelly (1933, p. 83) found marine or brackish-water fossiliferous mudstone or limestone in cycles A, B, E, and F (table 10); fresh-water pelecypods in mudstone in cycles D and F; coal beds in cycles B, C, D, and F; underlay in cycles A, B, C, D, E, and F; plant-bearing mudstone in cycles A and F; and sandstone in cycles A, E, and F. Three periods of sand influx are indicated, as well as periods during which five or six soils formed that subsequently were overlain by peat in small bogs.

PALEOTECTONIC IMPLICATIONS

The Michigan basin must have been remarkably stable during interval B-time, as there is no known evidence of contemporaneous deformation within the basin. The periodic transgressions and regressions of the sea probably resulted from eustatic shifts in sea level, rather than from local diastrophic movements.

INTERVAL C

FORMATIONS INCLUDED

The youngest Pennsylvanian (and youngest Paleozoic) strata in the Michigan basin are referred to interval C. The name Saginaw Formation has been applied by many geologists to all of the Pennsylvanian rocks of Michigan, but the name Grand River Group was proposed by Kelly (1936) for a sandstone assemblage composing the upper part of the Pennsylvanian section throughout much of the basin; Grand River Formation (Cohee, 1944) is preferred by the U.S. Geological Survey for this upper unit. Interval C is equivalent to the Grand River Formation (table 10).

STRATIGRAPHIC RELATIONS

Three floral zones are recognized in the Pennsylvanian strata of Michigan (Arnold, 1949, p. 153-154). The "upper" flora was reported to contain Neuropteris rarioris, which is the diagnostic species of floral zone 9 of Read and Mamay (1964, p. K8-K10), and is characteristic of the lower part of the Allegheny Formation. Associated with this form are Sphenophyllum saxifragaefolium and S. marginatum, which are considered transitional between late Kanawha and early Allegheny time. Arnold concluded that the "upper" flora is equivalent in age to early Allegheny floras of the Appalachian region and that the rocks in which it occurs correlate with the Clarion and Kittanning Formations of Pennsylvania. Invertebrate faunas have not been reported in strata referred to interval C.

UPPER BOUNDARY OF INTERVAL C

Pleistocene glacial drift directly overlies the Grand River Formation throughout most of the Michigan basin where the formation is found. However, at some localities, especially toward the western side of the basin, Jurassic rocks (A. T. Cross, written commun., 1964) consisting of red mudstones which contain gypsum and some dolomite lie between the Grand River Formation and the Pleistocene glacial drift.

THICKNESS TRENDS

Interval C is present principally in the following two areas, which constitute only about half of the Michigan basin: (1) from the Saginaw Bay region southwestward to Calhoun and Jackson Counties near the south border of the basin; and (2) from Midland County westward to Newaygo and Lake Counties in the northwestern part of the basin. The maximum thickness is 275 feet at locality 363 in Mecosta County (pl. 7A). The thickness throughout most of the two principal areas is generally 50-100 feet, and only three drill records show a thickness of more than 200 feet.

Interval C was deposited on an uneven, locally eroded surface of interval B. Subsequent erosion of interval C occurred during both pre-Jurassic and pre-Pleistocene times; consequently, the assemblage is of a highly remnant nature. The present distribution and thickness of interval C, therefore, is not a reliable indication of the original distribution and thickness.

LITHOFAcies TRENDS

The Grand River Formation is mostly "massive, cross-bedded coarse-grained sandstones, which contain little or no mica, and which are frequently iron-stained on fresh fracture" (Kelly, 1936). The sandstone, which composes about 80 percent of the formation, commonly is
pink or red, in contrast to the gray sandstone of the underlying Saginaw Formation. Red mudstone also is present. Higher percentages of mudstone are generally found at increasing distances from Saginaw Bay.

Other constituents occur in small quantities, as reported in sample logs. Gypsum is locally associated with red beds, and one bed of gypsum in Clare County is 20 feet thick. Limestone and dolomite are found sporadically in rocks of this interval; the maximum thickness of limestone is 49 feet at locality 158 in Newaygo County. Carbonates are also known in five other separate districts. Coal, having a maximum thickness of about 2 feet, is reported in five areas, principally along the east side of the basin southwest of Saginaw Bay.

The lithofacies proportions and their lateral variations illustrated by plate 7B have limited significance because at no place is the entire interval represented. The lithofacies pattern reflects only strata of the uneroded lower part of the interval.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

**SOURCES**

The present distribution of interval C in two bands that tend to converge near Saginaw Bay, although highly remnant, suggests that the principal sources of sediment for interval C, as for interval A and to a lesser degree for interval B, lay east and northeast of Saginaw Bay. This is further supported by the greater proportion of sandstone near the bay, and by a study of sandstone dispersal patterns (Shideler, 1969). No information on the crossbedding direction of sandstones composing the Grand River Formation has been obtained.

The highly quartzose sandstones resemble those of interval A; both contain substantial amounts of rounded and occasionally frosted grains, such as are common in the Mississippian Marshall Formation of the Michigan basin and the Cambrian-Ordovician rocks of the central United States and Lake Superior region. The interval C sandstones are believed to have been derived from such earlier Paleozoic arenites.

**ENVIRONMENTS OF DEPOSITION**

Deposits of interval C, like those of earlier intervals, were chiefly alluvial or deltaic; the sandstones were deposited mainly in fluvial channels that spread out westward and southward across the basin from the Saginaw Bay area. Some channels may have been cut into previously deposited interval B strata, but most channel sandstones were probably deposited over earlier sediments without much prior erosion.

The western and northwestern parts of the basin received less sand than the southern and central parts. Some areas may have been cut off from the remainder of the basin by aggrading distributaries, somewhat as the Salton Sea was cut off from the Gulf of California by the lengthening of the delta of the Colorado River. The water in such areas became hypersaline at times, resulting in the local deposition of limestone, dolomite, and gypsum. No marine faunas have been reported from the carbonate rocks.

The red color of some of the rocks is believed to be the result of oxidation that took place after deposition, while the sediments still lay near the depositional interface (T. R. Walker, written commun., 1965). Walker considered this explanation to be especially probable if this area, like coastal regions of the Gulf of California today, was both warm and arid. The environment in which these red beds and associated gypsum were formed was less than 100 miles west or northwest of an area where coal was forming. These coal beds are very sporadic and are located mainly along the eastern side of the basin.

No regionally traceable succession of lithologic units is known for interval C in the Michigan basin. Therefore, details of sedimentation, including transgression, regression, and delta building, cannot be determined as they can in the Appalachian, Illinois, and midcontinent regions.

**PALEOTECTONIC IMPLICATIONS**

The Michigan basin apparently was fairly free from differential movements during the deposition of strata of interval C. The much larger proportion of sand in interval C, as contrasted with that of interval B, may indicate tectonic uplift of source areas to the east or northeast of the basin.

**TOTAL THICKNESS OF PENNSYLVANIAN ROCKS**

Strata of Pennsylvanian age, preserved in the central and deepest part of the Michigan basin, represent the youngest Paleozoic system in the region. Because of post-Pennsylvanian erosion, preservation of Pennsylvanian rocks is incomplete, and the total thickness once present is unknown.

Rocks of Morrow to Des Moines age, which are assigned to intervals A, B, and C, are believed to be present. The thickest remnants of Pennsylvanian rocks thus far known—650-700 feet—are in Midland and Isabella Counties (Cohee and others, 1951, sheet 4; fig. 11); however, thickness data are limited because the basin is covered with thick deposits of Pleistocene drift, which exceed 900 feet in the northwest. Outcrops are largely confined to the vicinity of Grand Ledge (Kelly, 1933, 1936); and subsurface exploration for coal has been limited mainly to the areas of thinner drift in the eastern and southern parts of the basin.

On plate 11, isopachs for the Pennsylvanian are roughly concentric. The axis of greatest thickness follows a curvilinear trend, which extends northward from Clinton County to Midland County then curves northwestward into Oseola County. Irregularities in the isopachs are attributed mainly to topography of the
pre-Pennsylvanian surface, which was produced by Late Mississippian deformation and erosion, and to post-Pennsylvanian erosion, which occurred before deposition of Pleistocene glacial drift. Some differential compaction also may have affected thickness.

The greatest local irregularity in thickness is in the area of the Howell anticline, largely in Livingston and Shiawassee Counties in the southeastern part of the basin. On the crest of the anticline (loc. 348), in an area too small to be shown on plate 11, the total thickness of Pennsylvanian strata is 55 feet; however, it increases to more than 400 feet within 10 miles to the east and to the west. Some structural control of thickness is evident on the east and south sides of the basin; there, the 100-foot isopach, located 15-25 miles from the zero line, seems to define a shelf bordering the basin. Cross sections indicate that thin sequences in the eastern and southern parts of the basin represent approximately the same part of Pennsylvanian time as is represented by the thicker deposits of the central basin. On the west and north margins of the basin, depositional thinning on a marginal shelf is less readily recognized.

Preglacial topographic irregularities at the surface of Pennsylvanian rocks were mapped by Leverett and Taylor (1915, pl. 2), who showed a westward-draining preglacial valley across the Michigan basin from Saginaw Bay to Newaygo County. A check of bedrock elevations in deep wells drilled since 1915 showed, at some places, a variance of 200-400 feet from the previously published contours. The relief of the preglacial surface is locally greater and the surface more irregular than was shown on the Leverett and Taylor (1915, p. 529) map.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

UNITS OVERLYING PENNSYLVANIAN SYSTEM

In many wells drilled for oil and gas in the Michigan basin, especially in its western and northern parts, a unit consisting of red mudstone and sandstone, gypsum, and some dolomite overlies the beveled Pennsylvanian surface. These deposits are in notable contrast to the underlying coal-bearing gray, dark-gray, or tan Saginaw Formation, or the Grand River Formation. Locally, this unit is nearly 200 feet thick, but in most places it is much thinner. The unit was called "Permo-Carboniferous Red Beds" in early papers on Michigan stratigraphy. However, spores collected from well cuttings indicate that it is of Late Jurassic (Kimmeridgian) age (A. T. Cross, written commun., 1967) and is about equivalent to the Morrison Formation of the Western United States. This Michigan occurrence of Jurassic rocks was not treated in the Jurassic folio (McKee and others, 1956), because the discovery of its Jurassic age postdates the publication of that folio. The formation probably does not crop out anywhere in the Michigan basin, but it is the immediate cover of Pennsylvanian strata in about one-third of the basin (pl. 12).

In a few wells on the west side of the basin, Jurassic rocks immediately overlie the Michigan Formation (lower middle Mississippian); whereas, a short distance to the east, at about the same elevation, they overlie strata of the Saginaw Formation.

All the Michigan basin is covered by Pleistocene glacial drift, which overlies the Jurassic red beds or the Pennsylvanian Saginaw Formation or Grand River Formation.

PALEOTECTONIC IMPLICATIONS

After Pennsylvanian deposition, the Michigan basin region was upwarped and beveled by erosion. In Jurassic time, red beds accumulated in a restricted and probably slightly depressed part of the region. The red beds contain spores shown by Cross (1966) to be similar to those in the Fort Dodge Gypsum of Iowa, which suggests that a connection may have existed between the Michigan Jurassic and the Jurassic farther west.

No post-Pennsylvanian faulting or folding is known to have occurred within the region. The cutting of preglacial valleys through the Jurassic red beds implies slight post-Jurassic upwarping.

REFERENCES


Illinois Basin Region

By HAROLD R. WANLESS

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-E
CONTENTS

Abstract .................................. 71
Region defined ................................ 71
Paleogeology .................................. 71
Units underlying Pennsylvanian ............ 71
Lower boundary of the Pennsylvanian ....... 71
Paleotectonic implications ................. 71
Interval A .................................... 72
Formations included ......................... 72
Upper boundary of interval A ................ 72
Thickness trends .............................. 72
Lithofacies trends ............................ 72
Sources and environments of deposition .. 72
Paleotectonic implications ................. 72
Interval B .................................... 73
Formations included ......................... 73
Surface of deposition ........................ 73
Stratigraphic relations ....................... 73
Upper boundary of interval B ................ 73
Thickness trends .............................. 73
Lithofacies trends ............................ 73
Sources and environments of deposition .. 73
Paleotectonic implications ................. 73
Interval C .................................... 74
Formations included ......................... 74
Stratigraphic relations ....................... 74
Interval D .................................... 75
Formations included ......................... 75
Upper boundary of interval D ................ 75
Thickness trends .............................. 75
Lithofacies trends ............................ 75
Sources and environments of deposition .. 75
Paleotectonic implications ................. 75
Interval E .................................... 76
Formations included ......................... 76
Upper boundary of interval E ................ 76
Thickness trends .............................. 76
Lithofacies trends ............................ 76
Sources and environments of deposition .. 76
Paleotectonic implications ................. 76
Reference ..................................... 76

ILLUSTRATIONS

[For listing of plates (in separate case) see volume “Contents”]

Figure 12. Map showing geographic features and localities in the Illinois basin region and adjacent areas mentioned in text .......... 73
13. Map showing tectonic elements in the Illinois basin region and adjacent areas in Pennsylvanian time ......................... 74
14. Map showing relation between mapped pre-Pennsylvanian and preglacial drainage lines in part of the Illinois basin region 92

TABLES

Tables 11-15. Stratigraphic units mentioned in text for intervals A-E but not shown on plate 13:
11. Interval A and part of interval B .................. 75
12. Interval B and parts of adjacent intervals .......... 76
13. Interval C and parts of adjacent intervals .......... 77
14. Interval D and parts of adjacent intervals .......... 78
15. Interval E .................................. 79
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

ILLINOIS BASIN REGION

By HAROLD R. WANELLES

ABSTRACT

Pennsylvanian rocks in the Illinois basin cover about three-fourths of Illinois; they are absent in the northern part of the State in a belt 80-100 miles wide, in a narrower belt at the south end of the State, and in other belts along the Mississippi and Illinois Rivers. The Pennsylvanian rocks extend into southwestern Indiana and western Kentucky and occur as small outliers separated from the Illinois exposures by the Mississippi River in the vicinity of St. Louis, Mo., and Muscatine, Iowa. The long axis of the basin from northwest to southeast is 400 miles, and the maximum width northeast from St. Louis is 240 miles. Unfortunately, each of the five States involved has a different nomenclature for many of its groups, formations, and small units of Pennsylvanian age.

The Illinois basin is separated from the midcontinent region on the west by the Ozark uplift and the Mississippi River arch and from the Appalachian region on the east by the Cincinnati arch. It is bordered on the north by the Wisconsin arch and separated from the Michigan basin by the Kankakee arch. The depositional basin during Pennsylvanian time extended southward far beyond the limits of the present structural basin.

The Illinois basin is divided into segments by two asymmetrical folds—the La Salle anticline trending north-northwest and the Duquoin monocline trending north. The steep limbs of these folds separate the Fairfield basin (the deep part of the Illinois basin) from shelf areas to the east and west. Near the present southern end of the Illinois basin the east-west Shawneetown-Rough Creek fault zone separates the narrow Moorman syncline from the rest of the basin. The greatest thickness of Pennsylvanian sediments, about 3,200 feet, is in the Moorman syncline in western Kentucky.

Before the beginning of Pennsylvanian deposition, the area that now forms the Illinois basin was elevated at its north end and depressed moderately to the south. It was beveled by erosion; consequently, near Ottawa, in northern Illinois, Pennsylvanian strata rest directly on the St. Peter Sandstone (lower Middle Ordovician), whereas in southern Illinois the St. Peter is 6,000 feet below the base of the Pennsylvanian. The La Salle anticline also was elevated and stripped of younger sediments; therefore, for nearly 200 miles along the crest of the anticline, strata underlaying the Pennsylvanian are older than those on its flanks.

Shortly before subinterval A₂ time, the whole area of the Illinois basin was emergent, and a series of streams crossed it on a predominantly southwestward slope. These streams cut valleys 200-300 feet below the surrounding plains. The largest valley entered western Kentucky near the southeast corner of the Illinois basin and carried drainage from the Appalachian area of eastern Kentucky. Another valley, referred to as the Evansville channel, paralleled the La Salle anticline on the east for more than 100 miles before curving southwestward near Evansville, Ind., providing evidence that the La Salle anticline was a topographic barrier at this time.

In northern Illinois where the substrata were dominantly carbonate rocks, caves and sinkholes had formed in Ordovician, Silurian, Devonian, and Mississippian limestones and subsequently were filled with Pennsylvanian sediments.

Sources of sediment were distant and principally northeast of the Illinois basin, probably in the same region from which detritus was carried southward to the northern Appalachian basin and westward to the Michigan basin. Because of the barrier of the La Salle anticline, much less sand was deposited west of this anticline than east of it, and the sand deposited to the west probably came from a northwestern source. Later in the Pennsylvanian, in late interval B time, some sand may have entered the basin from the southwest from sources in the Ouachita uplift; however, this interpretation has been strongly questioned by some geologists.

At the beginning of Pennsylvanian deposition, sand from the northeast was highly quartzose and was evidently reworked from older Paleozoic strata, but in interval B time, as indicated by increasing amounts of mica and of exotic rock fragments, upland sources had been stripped of their sedimentary cover, and micaceous metamorphic rocks were supplying detritus.

In the deep southeastern part of the basin, where substrata consist of Chester (Upper Mississippian) rocks, earliest sediments were deposited in and filled pre-Pennsylvanian valleys before sediments began to spread over the uplands bordering the stream valleys. In subinterval A₂ most rock units are restricted to limited areas, their distribution controlled by local pre-Pennsylvanian topography. Cyclic sedimentation is much less obvious in these deposits than in those of later intervals.

During interval B time seas entered the Illinois basin two or three times from an easterly direction, probably from the Appalachian basins. Most of the records of these transgressions are in Indiana and western Kentucky, but during at least one transgression the sea reached southern Illinois. The maximum thickness of interval B in western Kentucky is 650 feet. The uplands of the La Salle anticline and Mississippi River arch in northern and central-western Illinois continued to be eroded and weathered; surface clays that accumulated in both areas were highly kaolinitic. Encroachment of the sea on Pennsylvanian land areas had reached all parts of the basin, except the highest areas, by the end of interval B time.
Late in interval B time the valleys of northwestern Illinois began to be drowned during the first marine transgression from the west. The Spillow Limestone Member of the Spoon Formation was distributed in the estuaries of streams that evidently entered Illinois from Iowa at the north end of the Mississippi River arch. At about the same time came the latest transgression of the sea from the east; the Curlew Limestone Member of the Tradewater Formation was deposited in western Kentucky, and its correlative was deposited in southwestern Indiana. By the middle of interval C time the last of the pre-Pennsylvanian uplands of the Illinois basin had been covered with Pennsylvanian sediments. A virtually uninterrupted depositional surface extended from the Nemaha anticline of eastern Kansas and Nebraska across Missouri and Iowa to Illinois. This surface was inundated several times by marine waters, which at times extended eastward across the basin to Ohio and Pennsylvania. Differential vertical movements on the La Salle anticline and other positive elements had nearly stopped; thus, at most times there were no topographic barriers between Kansas and Pennsylvania.

During interval C time the sea made 14 transgressions and regressions across the Illinois basin. Marine limestones or mudstones generally less than 10 feet thick mark the times of maximum transgression. The marine beds directly overlie widespread coals, formed during early stages of the transgressions. Between marine inundations, mud from the northeast or north formed prodeltas on which were superposed channel sandstones, mostly deposited in the distributaries of advancing deltas. When the rate of detritus import declined, weathering on exposed surfaces produced kaolinitic clays, particularly in Indiana; in low topographic positions, as in western Illinois, long periods of nondeposition occurred in areas under a shallow cover of water where the sediment was not exposed to weathering. These successive transgressions, regressions, and delta-building episodes constitute the record of cyclic sedimentation for which Illinois is well known.

The southeastern part of the Illinois basin subsided slowly during the time of interval C. Rocks of this age are only 100-200 feet thick near St. Louis on the Mississippi River arch, whereas in southeastern Illinois they are more than 900 feet thick.

During interval D time, cyclic sedimentation similar to that of interval C continued. At least 14 marine transgressions and regressions deposited a sequence which, where complete, ranges in thickness from 400 to 950 feet. Less coal is present, and coals are thinner than in interval C.

Post-Pennsylvanian erosion has limited rocks of interval E to three areas just west of the La Salle anticline in the Fairfield basin and to one area in the Moorman syncline of western Kentucky. In the Moorman syncline 800 feet of rock record several marine transgressions and regressions. The great number of strandline oscillations recorded in intervals C, D, and E (at least 34) seems more likely explainable by world fluctuations in sea level than by local tectonism.

REGION DEFINED

The Illinois basin, formerly called the Eastern Interior coal basin (Wanless, 1962; Weller and Bell, 1937), consists of about three-quarters of Illinois and adjoining areas in southwestern Indiana and western Kentucky. Outliers north of the Mississippi River in Scott and Muscatine Counties, Iowa, and west of the Mississippi in St. Louis County, Mo., are detached remnants of the deposits of this basin. Another series of small outliers extends east from the coal field in western Kentucky toward the Appalachian coal field and is composed of sandstone deposited as detritus in a river that formerly drained westward from the Appalachian region (Miller, 1910b). A small outlier near Remington, Ind., about 30 miles north of the present basin margin, likewise is the remnant of a channel sandstone.

Unfortunately for the purposes of stratigraphic discussion, five systems of classification are in use in this basin, including the outliers across the Mississippi River to the west. Not only the formations, subgroups, and groups but also many of the beds of coal, limestone, and sandstone are differently named. The correlation, stratigraphic relations, and depositional history of the region, therefore, are difficult to describe without a cumbersome multiplicity of names and are difficult for a geologist unfamiliar with the region to comprehend.

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

The Illinois basin includes Pennsylvanian rocks in the region of Illinois, southwestern Indiana, and western Kentucky and in two small areas, one near St. Louis, Mo., and another near Muscatine, Iowa (fig. 12); rocks of these small areas are today separated from the main part of the basin to the east by the Mississippi River. Strata directly below the Pennsylvanian range from the St. Peter Sandstone, of early Middle Ordovician age, to the Kinkaid Limestone and overlying Grove Church Formation, the highest formations of the Chester Series in its type area. The map of this basin was prepared using the control points shown on plate 2 but was checked and locally modified to conform with other pre-Pennsylvanian maps prepared with a denser net of control points (Wanless, 1962; Willman and others, 1967).

Because of truncation of strata across the basin, the St. Peter Sandstone of Ordovician age is about 6,000 feet below the Pennsylvanian in southern Illinois but directly underlies it in parts of northern Illinois. South and southwest from this northern area, units directly underlying the Pennsylvanian are, successively, younger Ordovician, Silurian, Devonian, and Mississippian. The La Salle anticlinal belt (fig. 13), which has a steep west flank, trends south-southeastward across the basin and plunges southeastward. Strata directly underlying the Pennsylvanian in a strip along the anticlinal crest are older than those on either side from the vicinity of La Salle County as far south as Lawrence County, Ill. West of this fold, in southeastern and south-central Illinois, is the deep part of the Illinois basin—the Fairfield basin. In it the Kinkaid Limestone of Chester age directly underlies the Pennsylvanian across a wide area, including parts of southern Indiana and western Kentucky.

LOWER BOUNDARY OF THE PENNSYLVANIAN

Paleogeology below the basal Pennsylvanian unconformity in southern Illinois was mapped by Siever (1951). In addition to structural features, his map shows a system of pre-Pennsylvanian valleys now filled with
Lower Pennsylvanian strata. These valleys were locally incised 200 feet or more into an otherwise nearly level plain. The streams that cut them drained generally southwestward but apparently meandered within the valleys. The stream that occupied the largest valley originated in the Appalachian area and flowed westward across the Cumberland saddle between the Nashville dome and Jessamine dome (Miller, 1910b; Hagan, 1942). Another major valley, which had a southward trend, lay just east of the La Salle anticline and has been called the Evansville valley after the town of Evansville, Ind. (Potter and Desborough, 1965). The pre-Pennsylvanian paleogeology and drainage for the southern half of the Illinois basin is shown on a map by Bristol and Howard (1971).

In many places where the unit underlying the Pennsylvanian is a carbonate rock, caves and sinkholes developed, and were filled with Pennsylvanian
sylvanian time includes the presence of Late Devonian fish teeth in solution cavities within Silurian carbonate rocks and of greenish-gray shaly mudstone resembling the Kinderhook Hannibal Shale in these sinks. Such remnants may be products of an earlier time of uplift and exposure, after which the area was buried by Mississippian sediments and subsequently exhumed.

The succession of events following accumulation of the Kinkaid Limestone probably was: (1) tilting and broad folding of the basin and bordering arches; (2) subaerial exposure and beveling of the tilted pre-Pennsylvanian rocks; and (3) carving of a drainage system that locally exposed some of the oldest rocks in the basin. Apparently a substantial time elapsed in this region between the deposition of latest Mississippian and earliest Pennsylvanian strata. This hiatus may be represented in Oklahoma by rocks of Springer age, in Alabama by the Mississippian Parkwood Formation, in Virginia and southern West Virginia by the Pocahontas Formation, and in the Southern Anthracite coal field by the Tumbling Run Formation. The La Salle anticline began to rise in northern Illinois prior to deposition of the youngest Mississippian rocks within the basin, and folding continued into the Pennsylvanian.

On the west side of the Illinois basin the Mississippi River arch, which separates this basin from the Forest City basin farther west (fig. 15), probably originated in Mississippian time and continued into the Pennsylvanian. Outliers of Pennsylvanian strata bordering the arch can be correlated with strata in both basins and show that it was covered with sediment by late interval B time. The sedimentary sequence on the arch is thinner than contemporary sequences on either side, indicating that, although periodically flooded, the arch nevertheless retained a slight submarine relief during deposition. Strata of subinterval A₂ are uncommon near the arch, suggesting that the arch formed later than the La Salle anticline.

INTERVAL A
FORMATIONS INCLUDED

Strata of the Caseyville Formation of Illinois and western Kentucky are referred to interval A because they yield diagnostic floras of Morrow age. In Indiana, rocks of corresponding age occur in the Mansfield Formation, which is believed to include strata of both interval A and, at least in central-western Indiana, the lower part of interval B. In a small isolated area of northwestern Illinois and adjacent Iowa, strata referred to the Caseyville Formation because of diagnostic spores are assigned to interval A. Before the discovery of these spores all Lower Pennsylvanian strata of western Illinois had been assigned to the Abbott Formation of interval B. In the Illinois basin all of interval A is the

PALEOTECTONIC IMPLICATIONS

The Kinkaid Limestone, the youngest remaining widespread formation of Mississippian age, is preserved because it was downwarped into the Fairfield basin when the La Salle anticline was relatively elevated. Subsequently the whole area was exposed subaerially, and the drainage network, described by Siever (1951) and Bristol and Howard (1971), was developed. Still later, during subinterval A₂, the earliest Pennsylvanian sediments accumulated on the valley floors of southwestward-draining streams. From this time on, the basin seems to have been largely aggradational. Sand filled the stream valleys, and eventually, detrital deposits spread across the entire basin to form a featureless plain late in the time of interval B and during interval C. Evidence that the northern Illinois carbonate area had been exposed prior to Middle Pennsylvanian.
equivalent of subinterval A₂ in the Appalachian region, except in Orange County, Ind., where there is a local occurrence of subinterval A₁.

Various stratigraphic units of interval A in the Illinois basin (table 11) are correlated as follows: The roof shale of the Gentry coal, formerly called Battery Rock coal of southeastern Illinois, yields fossil plants of floral zone 6 (Read and Mamay, 1964), permitting approximate correlation of the Gentry with the Sharon coal of Ohio and Pennsylvania, the Sewell coal of West Virginia, and the Sevancee coal of Tennessee (table 2). Plants of floral zone 5 are found in the Lusk Shale Member of the Caseyville Formation in Johnson County, southern Illinois (fig. 12), indicating equivalence of the Lusk to the lower part of the New River Formation in subinterval A₁ of West Virginia. The Hindostan Whetstone Beds of the Mansfield Formation (Ashley, 1899, p. 1082-1083; Franklin, 1939) in Orange County, Ind., yield fossil plants belonging to floral zone 4, equivalent to the Pocahontas Formation in subinterval A₁ of southern West Virginia. The only area in the Illinois basin-eastern midcontinent region in which strata of subinterval A₁ have been discovered is in Orange County.

The Caseyville Formation of Illinois and western Kentucky is almost connected with the Lee Formation of eastern Kentucky by a series of small outliers on a drainage divide between the Jessamine and Nashville domes (fig. 13) of the Cincinnati arch (pl. 3A).

Strata referred to interval A in the Illinois basin are characterized by massive coarse-grained to pebbly crossbedded sandstones that include small rounded quartz pebbles. With few exceptions, no other strata of Pennsylvanian age in the Illinois basin contain quartz pebbles. Sandstones of interval A resemble corresponding sandstones of the Appalachian basin except that maximum pebble size is a little smaller in the Illinois basin. Sandstones are more difficult to correlate in the deep parts of the Illinois basin, than along the outcrops, but many of them can be identified because drillers' logs commonly note "hole full of water" at the positions of sandstones in interval A, which are coarser and more permeable than those higher in the Pennsylvanian.

Studies of two coals at the base of the Pennsylvanian along the northwest margin of the Illinois basin in Rock Island County, Ill., and in Muscatine County, Iowa, showed the presence of spores equivalent to those in the Caseyville along the southeast margin of the basin (R. M. Kosanke, in O'Brien, 1963, p. 73). The two coals are widely separated from other known strata of interval A, but they are referred to this interval on the basis of their spore floras. They occur in outcrops for about 15 miles and underlie the Babylon Sandstone Member of the Abbott Formation (interval B) (table 12), which in most of western Illinois is the basal unit of the Pennsylvanian.

The Lower Pennsylvanian of southeastern Iowa has not been studied much but is known to contain coal locally and perhaps should in part be referred to interval A.

**UPPER BOUNDARY OF INTERVAL A**

Strata yielding plants of floral zone 6 are generally a few feet below the top of the Caseyville Formation. Above the Caseyville in the southern part of the basin is a series of massive sandstone beds named, in ascending order, the Grindstaff, Finnie, and Murray Bluff Sandstone Members of the Abbott Formation. Similar rock units occur in the lower part of the Tradewater Formation of western Kentucky and in the upper part of the Mansfield Formation and lower part of the Brazil Formation of Indiana (pl. 13); they are assigned together with the intervening rocks to interval B of this paper.

Sandstone above the Caseyville generally has no quartz pebbles and includes progressively more clay minerals, feldspar, and rock fragments from the top of the Caseyville upward (Siever, 1957). It is also finer grained and less permeable than the sandstone of the Caseyville.

In many places the uppermost unit of the Caseyville is the massive crossbedded Pounds Sandstone Member, the top of which is considered as the upper boundary of interval A. In Indiana, where no formalational boundary in this position has been recognized, similar lithologic criteria are used to separate interval A from the overlying interval B. In the northern part of the east side of the Illinois basin, in Indiana and eastern Illinois, strata of interval A are restricted to valleys incised in pre-Pennsylvanian rocks, and interval A is overlapped in intervalley areas by interval B.

---

**Table 11.—Stratigraphic units of less than formation rank mentioned in text for interval A and part of interval B but not shown on plate 13: Illinois basin region**

[Modified from Kosanke and others (1960)]

<table>
<thead>
<tr>
<th>Interval</th>
<th>Southern Illinois</th>
<th>Indiana</th>
<th>Western Kentucky</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Abbott Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Murray Bluff Sandstone Member</td>
<td>Finnie Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Grindstaff Sandstone Member</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A₂</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Caseyville Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pounds Sandstone Member</td>
<td>Gentry coal Setlers Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Battery Rock Sandstone Member</td>
<td>Lusk Shale Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>A₁</td>
<td>Hindostan Whetstone Beds</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

¹ Named units not in text are omitted.
THICKNESS TRENDS

Interval A in the Illinois basin is largely limited to a subtriangular area in the southeastern third and is present throughout the Kentucky part (pl. 3A). Its north and northwest margins have been interpreted from characteristics of sandstones penetrated by oil exploration wells; thick highly permeable sandstones with large self potentials on electric logs are referred to this interval. The interval is not recognized in the Pennsylvanian outcrop area in western Illinois north of Randolph County. An abrupt thinning of Lower Pennsylvanian strata occurs along the trend of the Duquoin monocline (fig. 13), which separates the Fairfield basin from the shelf in western Illinois. Interval A seems to extend generally westward about to the monocline. The northeast border of this interval is marked by linear bands of rock that extend northeastward within nine pre-Pennsylvanian valleys; interval A crops out in some of these bands, and elsewhere it has been recognized only through subsurface studies. The valley farthest northeastward, near Remington, Jasper County, Ind., is known only by a small outlier in which pebbly sandstone rests unconformably on black mudstone of the New Albany Shale (Mississippian and Devonian).

Because of the very uneven topography below the base of the Pennsylvanian (Siever, 1951, p. 550-551, 554-555; Wanless, 1962, p. 18-19; Bristol and Howard, 1971, p. 9-13), thicknesses of interval A are extremely variable, commonly differing as much as 100 feet in less than a mile. To map the thickness variation accurately without using a substantial part of the more than 100,000 records of oil test bores in this area would be impossible; thus, the isopach map (pl. 3A) showing differences in thickness of 250 and 500 feet must be regarded as highly generalized. A maximum thickness of 677 feet is in Union County, Ky. Differential downwarping of the Fairfield basin during Caseyville time and movement on the Duquoin and La Salle monoclines folds on the margins of the basin are factors that also influenced thickness variation; however, comparative influences of downwarping and of the filling of pre-Pennsylvanian valleys have not been wholly evaluated.

LITHOFACIES TRENDS

Rocks of interval A in the Illinois basin include three massive pebbly sandstones in the Caseyville Formation: the Pounds Sandstone Member (upper), the Battery Rock Sandstone Member, and a sandstone in the basal Lusk Shale Member. Of these, the Battery Rock is the most conspicuous and the most pebbly, and at some places, such as in Crittenden County, Ky., it is more than 250 feet thick where it fills valleys. The other two sandstones follow somewhat different trends of maximum thickness (Potter and Desborough, 1965, p. 11-13). Pebble distribution in the sandstones is highly irregular.

At some places the rock is a conglomerate of rounded quartz pebbles in a matrix of interstitial sand, whereas at other places, only a few miles away, the rock may contain few or no pebbles.

Major sandstones of the Lusk, Battery Rock, and Pounds cyclothem in interval A resemble those of higher cyclic successions in the Pennsylvanian, but coals, clays, and mudstones are much less widespread than in the later sequences. As a result, correlation of small units from place to place within interval A of the Illinois basin is difficult. One minor marine transgression may have come from the direction of eastern Kentucky to the southeast.

In some areas much of interval A is mudstone, commonly dark gray (pl. 3B). Two or three coals are known, but none forms an extensive blanket as do numerous coals higher in the Pennsylvanian of this basin. Also included in interval A are local undercut, clay ironstone in thin bands or nodules, and one very local marine limestone, the Sellers Limestone Member of the Caseyville Formation in Hardin County, Ill.

A laminated siltstone, the Hindostan Whetstone Beds of the Mansfield Formation in Orange County, in southern Indiana, belongs to subinterval A, and is represented on the lithofacies map of that unit (pl. 4B).

Positions of known or interpreted pre-Pennsylvanian valleys have been sketched on a lithofacies map (pl. 3B) on the basis of data from Siever (1951), Wanless (1962), and Potter and Desborough (1965). Nearly all places where interval A contains 80 percent or more sandstone are in or close to these valleys. The boundaries between lithofacies patterns on plate 3B suggest that pre-Pennsylvanian valleys probably were the principal routes of sediment transport and, therefore, that in most places lithofacies were controlled by the valleys. Along the trends of some major drainages, however, the maximum proportion of sandstone is somewhat less than 80 percent. Presumably such areas of relatively low sand content, distributed along known alluvial valleys, reflect the contribution of nearby sources consisting largely of mudstone and limestone of Chester age.

The lithofacies distribution as mapped probably presents a more orderly pattern than actually existed. Drainage channels cut into pre-Pennsylvanian strata are the only ones that have been mapped, because the patterns of sandstone channels of the Pounds and Lusk Members are unknown. For this reason the map probably shows expectable relations between a system of southwestward-draining valleys and the sediments accumulated in and between them. This degree of order cannot be attained in mapping intervals above A, for intervals B-E each includes a composite of five or more separate sets of lenticular sandstones of river or distributary delta channels.
SOURCES AND ENVIRONMENTS OF DEPOSITION

SOURCES

Caseyville strata in Illinois commonly display, in spectacular fashion, crossbedding and other structures indicating current direction. Abundant measurements that have been accumulated (Potter and Olson, 1954; Potter and Siever, 1956; Potter, 1962, 1963; Potter and Desborough, 1965) indicate a predominant direction of sediment transport from northeast to southwest (Potter and Siever, 1956, p. 242). Where measurements were made in the valley of a meandering stream, the dominant crossbedding direction follows successive meander curves within the stream valley (Potter and Siever, 1956, p. 240).

In the Caseyville Formation, although most sand was derived from older sandstones, there is generally greater angularity of grains, especially of tourmaline, and larger amounts of clay minerals and feldspar than in the underlying Mississippian sandstone. Principal sources of these materials apparently were older sediments and crystalline rocks to the northeast, especially in the Canadian Shield and northern Appalachian highlands. Quartz pebbles in the Caseyville resemble those in the Lee and New River Formations of eastern Kentucky, Virginia, and West Virginia and probably had a common source. Those that reached the Illinois basin were carried by a river that crossed Kentucky. Certain blue quartz pebbles are similar to quartz that is common in veins of crystalline schists in the Virginia Piedmont. Petrologic studies bearing on Caseyville sediment sources were summarized by Siever and Potter (1956) and by Potter and Pryor (1961).

ENVIRONMENTS OF DEPOSITION

The only sedimentation known to occur in the Illinois basin during subinterval A, (late Springer) time was in a lake in Orange County, Ind.

The predominant environment of deposition during subinterval A, as represented by the Caseyville Formation of the Illinois basin, was fluviatile, as shown by primary structures such as large-scale crossbedding and by the geometry of the sandstone bodies. Other structures that reflect the type of environment include current ripple marks, flute marks, microrcrossbedding, and mudstone-pebble conglomerates (Potter, 1963, p. 27-49).

Shifting channel positions during sedimentation account for some of the complexities of sandstone distribution. Sandstone bodies of interval A in the Illinois basin generally lack the sheetlike form that is fairly common for sandstone in the Appalachian region and that is exemplified by the widespread Sewanee Conglomerate of Tennessee, a possible correlative of the Battery Rock Sandstone Member. The probable explanation is that pre-Pennsylvanian landforms in the southern Appalachian area had already been smoothed by burial at a time when Illinois basin sedimentation was still restricted mainly to valley floors. Siltstone and mudstone of interval A probably were deposited on flood plains adjacent to channels, but fine-grained sediments also accumulated in lakes, marshes, and deltas. Near the present Mississippi River in Jackson County, Ill., a broad outcrop belt, in which interval A contains a high percentage of sandstone, lies normal to the paleoslope and suggests a series of delta distributaries.

An original southwestward slope, demonstrated in Illinois, evidently extended farther to the southwest, reaching the sea either in Arkansas or between Arkansas and southern Illinois. Pennsylvanian strata have been removed from intervening areas. In Arkansas the Hale Formation has been described (for example, Giles and Brewster, 1930; Frezon and Glick, 1959) as a coarse-grained sandstone containing abundant casts and molds of marine fossils. It is approximately the same age as the Battery Rock of Illinois, and each unit is the first major sandstone below strata containing floral zone 6.

Near French Lick, in Orange County, Ind., the Hindostan Whetstone Beds of the Mansfield Formation contain fine regular laminations interpreted as lacustrine varves; more than 300 laminations can be counted in individual whetstone quarries. Strata associated with a coal below the varved whetstone contain a rich flora. The varves seem to indicate a seasonal climate for the area, but no markings indicative of freezing temperatures have been observed.

PALEOTECTONIC IMPLICATIONS

After Late Mississippian (Chester) sedimentation ceased in the Illinois basin, extensive uplift took place along the Duquoin and La Salle fold axes; upwarp of the La Salle structure in north-central Illinois amounted to perhaps 1,000 feet. This uplift was followed by a period of active valley cutting prior to the beginning of interval A sedimentation. Meanwhile, downwarping took place sufficient for several thousand feet of strata to accumulate farther south in Arkansas and Oklahoma and in the Black Warrior basin (fig. 7) of Mississippi.

Abnormally thick strata of the Caseyville in bands parallel to hinge lines of the Duquoin and La Salle folds were deposited in synclinal troughs formed during the time of interval A. By the end of interval A time the pre-Pennsylvanian erosional features had generally been smoothed and buried, and no further tectonism had occurred in the southeastern part of the basin. In the north and west parts of the basin, however, land areas that had been raised before Pennsylvanian time continued to influence sedimentation during the time of interval B.
INTERVAL B FORMATIONS INCLUDED

Interval B, as defined in Illinois, includes all of the Abbott Formation of the McCormick Group and the basal few feet of the Spoon Formation (including the Seville Limestone Member) of the Kewanee Group (Kosanke and others, 1960; Wanless, 1962). The classification of these units and their correlation with units recognized in other States of the Illinois basin are shown on plate 13.

In Indiana the upper part of the Mansfield Formation and the Brazil Formation are referred to interval B. A limestone overlies the Brazil Formation in Parke, Fountain, and Clay Counties, Ind. Although the limestone has been referred to the Staunton Formation by some geologists, it includes Fusulinella iowensis and is Atoka in age and may be older than the Staunton (Thompson and Shaver, 1964, p. 19) in other parts of Indiana.

In western Kentucky the lower part of the Tradewater Formation (table 12) is referred to interval B.

SURFACE OF DEPOSITION

In the southeastern part of the Illinois basin, pre-Pennsylvanian topographic irregularities had been smoothed and buried by sediments of interval A by the time extensive thin layers of interval B were deposited. In the northern and western parts of the basin, however, sediments of interval B were deposited, in part, on an irregular eroded surface of pre-Pennsylvanian rocks still exposed at the close of interval A time. By the end of interval B time pre-Pennsylvanian rocks were still exposed only on the crest and on the east flank of the La Salle anticline in northeastern Illinois and on the Mississippi River arch from Monroe County north to Adams County, Ill., near the Mississippi River (fig. 13).

STRATIGRAPHIC RELATIONS

Strata referred to interval B are characterized by fossil plants of floral zones 7 and 8 at localities in southern and western Illinois (Read and Mamay, 1964). The same floras also occur in Indiana and western Ken-

---

Table 12.—Stratigraphic units of less than formation rank mentioned in text for interval B and parts of adjacent intervals but not shown on plate 13: Illinois basin region

[Modified from Kosanke and others (1960)]

<table>
<thead>
<tr>
<th>Interval</th>
<th>Southern Illinois</th>
<th>Western Illinois-Missouri</th>
<th>Indiana</th>
<th>Western Kentucky</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cereal Springs</td>
<td>Granger Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Limestone Member</td>
<td>Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Seville Limestone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rock Island (No. 1) coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Murray Bluff</td>
<td>Bernadotte Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sandstone Member</td>
<td>Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Delwood coal</td>
<td>Pope Creek coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Finnie Sandstone Member</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Willis coal</td>
<td>Tarter coal</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Grindstaff Sandstone Member</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Babylon Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A,</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Caseyville</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Formation (part)</td>
<td>Pounds Sandstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Member</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1 Named units not in text are omitted.
tucky, but collections from these areas have received little study.

Fusulinids have proved useful in identifying the limits of Atoka rocks assigned to interval B in the Illinois basin. *Fusulinella iovensis* and *F. iovensis* var. *stouti* occur in the Sevile Limestone Member of the Spoon of western Illinois and *Fusulinella* sp. in the Grindstaff Sandstone Member of the Abbott of southern Illinois (Boskydell sandstone of Dunbar and Henbest, 1948, p. 14-15). *Fusulinella iovensis* is in the type outcrop of the Curlew Limestone Member of the Tradewater of western Kentucky, and *Profusulinella*, which is characteristic of the lower part of interval B, occurs in a limestone near Morgantown, Ky. (Thompson and others, 1959). *Fusulinella* and *Profusulinella* are widely distributed in western Kentucky and along the east border of the Illinois basin in Indiana as far north as Warren County (Thompson and Shaver, 1964).

**UPPER BOUNDARY OF INTERVAL B**

*Fusulinella iovensis* is the fossil used to establish the upper limit of interval B. It occurs in the Curlew Limestone Member of the Tradewater in western Kentucky and, locally, in southern Illinois; it occurs also in the Perth Limestone Member of the Brazil in Indiana and the Sevile Limestone Member of the Spoon in western Illinois (Thompson and Shaver, 1964). The stratigraphic positions of the Curlew in western Kentucky and of the Perth in Indiana are known in a few places, but these limestones are probably not continuous over most of the basin.

The Sevile of western Illinois seems to be restricted to a series of estuarine embayments, beyond the sites of which neither the limestone nor the underlying Rock Island (No. 1) coal has been found.

In Saline County, southern Illinois, a limestone, formerly designated as Curlew, yields *Fusulina leei* and is probably correlative with the Cretaceous Limestone Member of the Spoon of Williamson County, which is assigned here to interval C. This species also occurs in the Putnam Hill Limestone Member of the Allegheny Formation of Ohio, which is in and very near the base of interval C in the Appalachian basin. In large parts of the Illinois basin, therefore, particularly in the deep central part, the boundary between intervals B and C is rather difficult to place consistently.

**THICKNESS TRENDS**

Thicknesses of interval B range from a few feet at the north and west margins of the basin to about 650 feet in Union County, Ky. In the northern part of the western Illinois outcrop area, interval B is somewhat less than 200 feet thick; it thins southward and wedges out against the Mississippi River arch in Adams and Brown Counties.

Interval B is absent along the La Salle anticline in northeastern Illinois and in other areas along this fold in parts of Ford, Champaign, Douglas, and Edgar Counties. Farther south, it is thinner on the crest of the anticline than on either flank as far as Crawford County, Ill. The interval is less than 200 feet thick along the Duquoin fold in Marion, Jefferson, and Washington Counties of southern Illinois.

Thicknesses of interval B exceed 400 feet west of the La Salle anticline from near the south margin of the interval north as far as Effingham and Cumberland Counties. A fairly regular thickening southward, virtually to the present southern outcrop, suggests that originally the basin extended farther south than it does now and perhaps joined the Black Warrior basin in Alabama south of the Nashville dome.

Local variations in mapped thickness result from a combination of: (1) folding during deposition, (2) overlap on hills of the pre-Pennsylvanian land surface, (3) presence of local detrital wedges, and (4) greater compactional thinning in mudstones than in adjacent sandstones. Difficulties in recognizing contacts consistently on electric logs may locally be a factor in the deep parts of the basin, where good key beds are lacking near both the upper and the lower boundaries of the interval.

**LITHOFAVIC TRENDS**

In most of the Illinois basin, sandstone and mudstone of interval B are nearly equal in amount. The amount of each constituent varies but is mostly within the range of 20-80 percent; in a few local areas the interval contains more than 80 percent sandstone or mudstone (pl. 6B). Most areas where sandstone content is high are near the borders of the basin; in such places the upper part of the interval commonly has been eroded, leaving the lower part, in which sandstone is disproportionately concentrated.

Three major sandstone beds, each locally attaining thicknesses of 50 feet, occur in the southern part of the basin. In ascending order, they are the Grindstaff, Finnie, and Murray Bluff Sandstone Members of the Abbott. The amount of mica, feldspar, interstitial clay, and minute rock fragments incorporated in these sands increased gradually during interval B time (Siever, 1957). The three sandstones of interval B can be readily distinguished from underlying and overlying sandstones. The Pounds Sandstone Member of the Caseyville, just below interval B, is highly quartzose, pebbly, and coarse grained. The Granger Sandstone Member of the Spoon, just above interval B, is very micaceous and argillaceous and resembles most of the other younger Pennsylvanian sandstones.

In northwestern Illinois, interval B includes only two conspicuous sandstones. The Babylon Sandstone
Member, near the base is highly quartzose, has secondarily enlarged grains, and locally contains chert fragments derived from underlying Mississippian rocks; the Bernadotte Sandstone Member, near the top, is very fine grained, well indurated, and quartzose. Conglomerate is rare in interval B.

Interval B includes gray silty mudstone, dark-gray carbonaceous mudstone (containing ironstone nodules), coal, and underclay. Dark mudstone typically occurs above the coal, especially in southern Illinois. The Cheltenham Clay of the region near St. Louis, Mo., and a nearly equivalent clay (not shown in table 12) of northeastern Illinois, locally called Goose Lake Clay, are valuable refractory clays of interval B that accumulated locally at or very near the base of the Pennsylvanian, at many places overlying strata belonging to interval A.

Economically important coals include the Rock Island (No. 1) coal of northwestern Illinois; the Upper and Lower Block coals and the Minshall coal of the Brazil region, Indiana; the Buffaloville coal of southern Indiana; and the Mannington and Bell coals of western Kentucky. Coals of interval B are highly variable in thickness and local in distribution. The most persistent coals of mineable thickness are probably the Block coals of the Brazil area, Indiana, equivalents of the usually thin Pope Creek and Tarter coals of western Illinois, and the Delwood and Willis coals of southern Illinois.

In western Illinois, interval B contains one marine limestone, the Seville, which is restricted in distribution to a series of estuarine embayments (Wanless and others, 1969). Its equivalents, the Perth in Indiana and the Curlew in Kentucky, are somewhat more continuous. These limestones are cherty in some outcrops. Additional limestones that occur lower in the interval in Indiana and western Kentucky do not seem to extend into Illinois. These include the Ferdinand and Fulda Limestone Members of the Mansfield Formation of Indiana and limestone in the Tradewater of western Kentucky.

Ironstone occurs as thin beds and concretions in the dark mudstones but, in this interval, is less conspicuous in the Illinois basin than nearby to the east in Ohio.

**SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES**

Sandstones are thicker and compose a larger part of interval B along the eastern and southern borders of the basin than in the western part, which suggests an eastern or northeastern source for the detrital sediments. A greater thickness of sandstone in southern Illinois and western Kentucky than elsewhere may, however, result more from subsidence of the basin than from nearness to sediment sources. The part of the basin just west of the La Salle anticline, from the north end of the anticline as far south as Douglas and Moultrie Counties, east-central Illinois, generally has less than 30 percent sandstone, indicating that the anticline acted as a barrier to westward transfer of sediment during deposition of interval B. Farther south the anticline was covered by interval B and probably did not block sediment movement.

The Babylon and Bernadotte Sandstone Members of the Abbott in western Illinois differ in mineral composition from the Grindstaff, Fannie, and Murray Bluff Members in the southern part of the basin. In western Illinois the sandstones are highly quartzose and apparently were derived from older sandstone. A suggested northwestern source for these sandstones is supported by cross-bedding dip directions (Siever and Potter, 1956).

An increase in the amount of feldspar, mica, clay minerals, and rock particles in sandstones of interval B suggests that northeastern source areas had been stripped of the mantle of sedimentary rock that had supplied much of the detritus for interval A, and that older crystalline rocks were beginning to furnish a significant part of the debris during interval B time.

**ENVIRONMENTS OF DEPOSITION**

Distribution of the principal sandstones has not been mapped in detail, but the sand bodies seem to have both linear (channel-fill) and sheetlike forms. The channel sandstones probably fill fluvial channels and distributary channels of deltas. Since crossbedding orientation suggests an eastern source for the major sandstones, fluvial channels probably are more prevalent in Indiana and, perhaps, western Kentucky, than in southern Illinois, where delta distributaries likely are more numerous. The Grindstaff Sandstone Member in parts of western Kentucky and southern Illinois would be an example of such transition westward from fluvialite to deltaic sandstone.

In southwestern Illinois the Boskydell Sandstone Member, which seems to be equivalent to part of the Grindstaff Sandstone Member in southern Illinois, contains many fragments of marine invertebrate fossils that probably accumulated near the seaward margin of a delta (Desborough, 1959). In the same region some sandstones show slump structures (Potter, 1963, p. 29-35) suggestive of accumulation on unstable slopes, such as distal margins of a delta.

The sheetlike sandstones may be flood-plain accumulations, interdistributary sands of a delta, or shallow marine sands. The Bernadotte Sandstone Member in western Illinois, a typical sheetlike sandstone, is the immediate substratum of the Rock Island (No. 1) coal in many places and contains abundant root (Stigmaria) impressions, which indicate that it served as a soil.

Some dark-gray mudstone in both northwestern and
southern Illinois contains abundant clay ironstone concretions and probably formed in brackish-water lagoons or bays. In Rock Island County, northwestern Illinois, similar mudstone contains inarticulate brachiopods but otherwise is generally unfossiliferous. Some kaolinitic underclays in interval B are very widespread and have been used as refractory clays in several parts of the basin. They probably were accumulated slowly, perhaps on or marginal to deltas in late stages of growth, and probably little or no sedimentation took place while they underwent diagenetic change from illitic to kaolinitic clay.

Coal swamps seem to have been fairly widespread during interval B in areas where pre-Pennsylvanian topographic relief had been subdued by the infilling of sediment during interval A. Exceptions to this generalization are the Pope Creek and Tarter coals, each generally less than 1 foot thick, which occur at many scattered places in western Illinois and probably accumulated in local topographic basins.

During the early part of interval B time, seas invaded only the eastern part of the Illinois basin. The Felda and and Ferdinand Limestone Members of the Mansfield Formation in Indiana were deposited in these seas. Limestones are conspicuous on electric logs of wells in southern Indiana but not on those in Illinois. Marine strata comparable in age to the Indiana limestones are present in eastern Kentucky; the seaways represented by the limestones may have entered from the east along a sag (Cumberland saddle) in the Cincinnati arch between the Jessamine and Nashville domes (fig. 13). Pro fusulinella, which characterizes these limestones, has not been found elsewhere east of the Mississippi River; therefore, an eastern route of inundation cannot be proved by fossil evidence.

Deposition of the Seville across western Illinois seems to have been largely in estuaries that widened westward toward a seaway in eastern Iowa. No continuity has been demonstrated between the Seville in the west and the Perth (Indiana) and Curlew (Kentucky) in the east, though all three contain Fusulinella iouensis. Seas in which the Perth and Curlew were deposited may have transgressed from the east, because Fusulinella iouensis is present in the lower Mercer Limestone Member of the Pottsville Formation in Ohio and in its correlative, the Magoffin Limestone Member of the Breathitt Formation, in eastern Kentucky.

Sedimentation in the Illinois basin was cyclic. The Tarter, Pope Creek, and Seville cyclothsms of western Illinois seem to be matched by sequences associated, respectively, with the Willis, Delwood, and Bidwell and New Burnside coals of southern Illinois; the Bell, Ice House, and Mannington coals of western Kentucky; and the Lower Block, Upper Block, and Minshall coals of Indiana. Interval B cyclothsms are like those in interval C for which environmental maps have been prepared (Wright, 1965; J. L. Weiner and J. B. Tubb, Jr., in Wanless and others, 1963). However, in interval B such positive elements as the La Salle anticline and Mississippi River arch probably constituted peninsulas of land separating depositional areas, features not represented in subsequent intervals.

PALEOTECTONIC IMPLICATIONS

During interval B time moderate differential downwarps in the Fairfield basin between the La Salle and Duquoin anticlines may have occurred, but the large thickness of interval B sediments in that basin probably resulted partly from the basin not having been full at the beginning of interval B time.

Influxes of detritus that formed the Grindstaff, Finnie, Murray, Bluff, Babylon, Bernadotte, and other sandstones and the associated mudstones may have been initiated by uplifts in source areas or by climatic changes that increased streamflow. Recurrent downwarps of the deepest part of the basin apparently caused considerable sedimentation there, but at no time during interval B deposition did the sea inundate all or even a major part of the basin.

INTERVAL C FORMATIONS INCLUDED

In Illinois, interval C includes all but the basal part of the Spoon Formation; all of the Carbondale Formation, Kewanee Group; and almost the entire lower half of the Modesto Formation, McLeansboro Group (pl. 13). The lower boundary of the interval is the base of the Granger Sandstone Member of the Spoon (table 13); the upper boundary is the base of the Trivoli Sandstone Member of the Modesto. The classification and nomenclature adopted by the Illinois Geological Survey in 1960 (Kosanke and others, 1960) succeeds an earlier classification (Wanless and Siever, 1956).

In Indiana the interval is composed of the Staunton, Linton, Petersburg, and Dugger Formations of the Allegheny Series and the Shelburn Formation and West Franklin Limestone of the Conemaugh Series. The base of interval C is a few feet above the Perth Limestone, the uppermost member of the Brazil Formation; the top of the interval is placed a few feet above the West Franklin Limestone, at the base of the rock unit equivalent to the Trivoli Sandstone Member of the Modesto in Illinois.

In western Kentucky the interval consists of the upper part of the Tradewater Formation, the Carbondale Formation, and the lower part of the Lisman Formation. The lower boundary is the base of the Granger Sandstone Member; and the upper boundary is placed at the base of a sandstone which lies a few feet above the Madisonville Limestone Member of the Lisman and is equivalent to the Trivoli Sandstone Member of the Modesto in Illinois.
### Table 13.—Stratigraphic units of less than formation rank mentioned in text for interval C and parts of adjacent intervals but not shown on plate 13: Illinois basin region

[Modified from Kosanke and others (1960); Missouri modified from Searight and Howe (1961)]

<table>
<thead>
<tr>
<th>Interval</th>
<th>Illinois</th>
<th>Indiana</th>
<th>Western Kentucky</th>
<th>Missouri</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Middletown Formation</td>
<td>Trivoli Sandstone Member</td>
<td>Ditney Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lonsdale Limestone Member</td>
<td>West Franklin Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Conant Limestone Member</td>
<td>Shelburn Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Jamestown coal</td>
<td>Dugger Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. 6 coal</td>
<td>Petersburg Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. 5 coal</td>
<td>Coal V</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Covel Conglomerate Member</td>
<td>Coal IVa</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. 4 coal</td>
<td>Mecca Quarry Shale Member</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Francis Creek Shale Member</td>
<td>Linton Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Colchester (No. 2) coal</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Isabel Limestone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sebree Sandstone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spoon Formation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seville Limestone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spoon Formation</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seville Limestone Member</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Upper Boundary of Interval C</td>
<td>Fusulinella iowensis is in strata a few feet below the interval boundary. This species occurs in the Seville Limestone Member of the Spoon Formation of Illinois, the Curlew Limestone Member of the Tradewater Formation of western Kentucky, and the Perth Limestone Member of the Brazil Formation of Indiana.</td>
<td>Fusulina occurs in marine limestones throughout interval C in the Illinois basin, but it is not known to occur in younger strata. Stratigraphic details of interval C have been given much attention by geologists working in the area, for nearly all the coal produced in the basin is mined from strata of that interval.</td>
<td>Fusulina occurs throughout this interval in the Illinois basin, but it is not known to occur in younger strata. Stratigraphic details of interval C have been given much attention by geologists working in the area.</td>
<td></td>
</tr>
<tr>
<td>krebs Subgroup</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

1 Named units not in text are omitted.

In St. Louis County, Mo., interval C is represented by the 
the Cherokee Group and the Fort Scott (three formations, not listed) and Appanoose (seven formations, not listed) Subgroups of the Marmaton Group (Searight and Howe, 1961, p. 85, 91). All these are in the Des Moines Series.

In Muscatine and Scott Counties, Iowa, interval C consists of the Des Moines Series, but in this area the Des Moines is thin and is not subdivided.

**Stratigraphic Relations**

Assignment of the rock units listed in table 13 to interval C is based chiefly on fusulinid zonation but in part on physical tracing of distinctive beds throughout large parts of the area and on dating by leaf-compression florals and small spores.

The lower boundary of the interval is believed to correspond closely to that used in the Appalachian coal field and in the Michigan basin, for in all three areas.
removed by erosion there and the lower part of C is the youngest Pennsylvanian present.

For purposes of mapping in the Illinois basin, the interval boundary has been placed at an erosional surface probably slightly above the time boundary separating the Des Moines from the Missouri Series. This slight departure from the time boundary is necessary because faunal control as found in the midcontinent region is not available here. Thus, the time boundary as recognized on the basis of fossils in the midcontinent region, when projected into the Illinois basin, seems to be between the Lonsdale Limestone and Eoline Limestone Members (table 13) of the Modesto Formation and below the assigned C-D boundary.

In the Illinois basin, evidence of an erosional break is less conspicuous below the Eoline than above it or its equivalents. The Trivoli Sandstone Member occupies several channels deeply cut in the Eoline Limestone Member, so its base is easy to recognize. The erosional surface seems to record a period of abnormally extensive emergence of the central United States and is believed to correlate with the Warrensburg and Moberly Sandstone Members of Missouri, which were at one time used as the base of the Missouri Series.

The genera Fusulina, Mesolobus, Prismopora, and Chaetetes are all believed to have died out at the end of the Des Moines. All these except Chaetetes are conspicuous in the Lonsdale Limestone Member and are known also at higher stratigraphic positions.

THICKNESS TRENDS

The thickness of interval C ranges from less than 200 feet in an outlier in St. Louis County, Mo., to more than 900 feet in southern Hamilton County, southeastern Illinois. The upper part of the interval has been eroded from all of Illinois west of the Illinois River valley except parts of Peoria County.

Interval C overlaps directly on rocks of pre-Pennsylvanian age in two areas in the Illinois basin: (1) northern Illinois east of the La Salle anticline, where locally the Coalchester (No. 2) coal directly overlies the St. Peter Sandstone of Ordovician age, and (2) Brown, Adams, and Pike Counties in western Illinois, where the same coal almost directly overlies Mississippian limestone and mudstone. Most of the Spoon Formation in the lower part of the interval is absent from these areas or is present only very locally.

The interval thickens generally southward across the basin and thins from east to west. The westward thinning reflects the presence of several units of sandstone and mudstone that form detrital wedges with thick margins. For example coals V and IVa are about 120 feet apart near Terre Haute, Ind., but their Illinois correlatives, coals No. 5 and No. 4, are within 5-10 feet of each other west of the Illinois River. Many other coals and limestones are closer together along the western border of the basin than in the eastern part.

Interval C seems to thicken westward rather abruptly locally at the La Salle anticline. In Cumberland County, Ill., it increases in thickness from 373 feet at locality 637 (pl. 1) to 534 feet at locality 636 in a distance of about 6 miles. Somewhat less marked thickening occurs eastward across the Duquoin monocline in southwestern Illinois. In southeastern Clinton County the interval thickens from 315 feet on the monocline to 461 feet 4 miles to the east. Within the interval, rock units between coals also show differential thickening across the La Salle anticline toward the Fairfield basin, the deep central part of the Illinois basin.

Many lenticular sandstone bodies occur in interval C in the eastern part of the Illinois basin. Where these sandstone bodies lie between two widespread coal beds, as in Jefferson County, southern Illinois, the vertical distance between coals may be as much as 30 percent greater than in places where mudstone lies between coals and constitutes most of the interval (Mueller and Wanless, 1957); the difference is probably due to differential compaction. At some places several lenticular sandstones are superposed in interval C; at other nearby localities the interval contains no lenticular sandstones. These differences may account for the large local variations in thicknesses of the interval as shown on the isopach map (pl. 7A).

LITHOFACIES TRENDS

Interval C in the Illinois basin contains mudstone of several sorts, sandstone, limestone, and coal. Many distinctive beds can be followed in outcrops and identified in well records, and these persistent beds constitute a stratigraphic framework for reconstructing the history in large parts of the basin. A tabulation shows that nearly 90 beds can be identified separately throughout large areas, including 16 sandstones, 15 underclays, 18 coals, 6 black fissile mudstones, 13 fossiliferous marine limestones, 3 freshwater limestones, and 16 gray mudstones, some containing marine fossils. All these beds except the sandstones and gray mudstones are thin, generally less than 5 feet thick. The aggregate thickness of these thin beds is rarely sufficient to appreciably affect the total lithologic composition of the interval. Sandstone and gray mudstone compose most of the interval, and the lithofacies variations on the map (pl. 7B) show almost wholly the various proportions of these two rock types.

At most places, the interval is less than 50 percent sandstone and commonly less than 20 percent. The greatest amount of sandstone is in the southeastern part of the basin, southeast of a line extending from Ver-
milion County southwestward to Randolph County, Ill. Variations in lithofacies pattern are numerous and are largely the result of differences in thickness of lenticular sandstones at various sampling localities.

The spacing of sample sections used in this study averages one control point per township. By way of comparison, an average of four control points per square mile was used by Staffeld (1954) in Richland County, southeastern Illinois, for the study of a stratigraphic unit nearly equivalent in thickness to interval C. He showed that within Richland County the proportions of sandstone to mudstone ranged from a ratio of 1:1.75 to 1:9. The lithofacies map accompanying the present report, therefore, is far more generalized than it would be if more records were used.

In the whole Illinois basin only 15 of approximately 1,-500 localities studied have more than 20 percent non-detrital rocks. These localities are near the Mississippi River in south-central Illinois. The higher proportion of non-detrital sediments in this area results from its position west of the margins of many of the detrital wedges that characterize the basin. Limestones are thicker and mudstones thinner than elsewhere in this basin.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

SOURCES

Paleoslopes in the Illinois basin during the deposition of interval C ranged from southward to westward, but a southwestward direction was predominant (Potter, 1962). At least 15 detrital wedges of sandstone and mudstone are known in interval C in the Illinois basin. These have fanlike or deltaic outlines, and the greatest thickness is near the point of entrance to the basin. Individual wedges are not directly superposed. The bulge of one wedge apparently stood as a slight rise when the next wedge formed. Thus, sediment accumulated in unfilled parts of the basin until sediments of several wedges were nearly comparable in thickness from place to place.

Components of a few wedges were transported into the basin from the north in northern Illinois, but most of the wedges are along the east margin of the basin—especially in south-central Indiana in an area from north of Terre Haute southward to the Kentucky border.

The grain size of sediment ranges from fairly fine grained sand to clay, suggesting derivation from a distant source area. The upslope trends of axial lines bisecting many of the detrital wedges indicate that important sources lay in the eastern part of the Canadian Shield or in the northern part of the Appalachian highlands in eastern Canada or New England. The sandstones contain angular quartz grains, mica, feldspar, interstitial clay, stable heavy minerals, and some moderately unstable minerals, such as garnet (Siever, 1957). Their composition contrasts with that of underlying sandstone, in the same area, which contains higher proportions of quartz and less mica, feldspar, interstitial clay, and unstable heavy minerals. Garnet was not found in sandstone below the Sechourne Limestone Member of the Spoon Formation near the base of interval C in western Illinois, but it forms about 20 percent of the heavy mineral content of the Isabel Sandstone Member of the Spoon Formation, which lies about 30 feet higher stratigraphically in the same area. If both sandstones came from the same source area, the earlier sandstone was formed by redeposition of the sands from earlier sediments, but the later probably is made up of first-generation sand derived directly from crystalline rocks of New England.

The Cincinnati arch may have been the source of lenticular highly micaceous sandstones in interval C of western Indiana (Friedman, 1960). Carefully mapped sandstone channels appear to narrow eastward, as though the head of the drainage system lay only a few miles distant from the present channels. Source materials probably were earlier Pennsylvanian and Mississippian sandstones, according to Friedman (1960).

Objections to the interpretation of Friedman include the following: (1) Widespread truncations of earlier Pennsylvanian sandstones by later ones are not known on the Cincinnati arch and (2) the Lower Pennsylvanian and Mississippian sandstones that outcrop east of the channel sandstones in Indiana are very quartzose and contain little feldspar, mica, or interstitial clay, which characterize sandstone in the channels. Lenticular channel-type sandstone has not yet been projected back to its source highlands, so the sources can be postulated only in a general way on the basis of lithology and paleogeography.

**ENVIRONMENTS OF DEPOSITION**

The large number of rock units composing interval C in the Illinois basin represents a diversity of environmental conditions which, as shown by multiple repetitions of nearly similar lithologic types, must have recurred episodically in the same localities during interval C time. Records of marine and nonmarine environments alternate at least 14 times in interval C and thus indicate 14 marine transgressions followed by regressions.

Marine transgressions of interval C generally came from the west. The sea passed around the northern border of the Ozark highlands and, when most extensive, deposited marine sediments as far east as Indiana and western Kentucky. During some transgressions, such as the one during which the Conant Limestone Member of the Carbondale was deposited on the Jamestown coal, the sea extended only a short distance into the basin.
Cyclic sedimentation was particularly well developed in the Illinois basin during the time of interval C (Weller, 1930; Wanless and Weller, 1932). During this time the sea transgressed more times in the Illinois basin than in the Appalachian region, and more non-marine sediments were deposited in Illinois than in the northern midcontinent region. Thus, the strandline must have migrated much more frequently across parts of the Illinois basin than across areas farther east or west.

A generalized account of the succession of environments during a single cycle of sedimentation, beginning at the time of maximum transgression, follows.

Limestone formed in a sea that had entered the Illinois basin from the west through Missouri and Iowa and spread eastward until it met an advancing front of prodelta mud. The eastward termination of the main area of limestone was an arc concave to the east or northeast, and narrow tongues of limestone extended eastward beyond the arc. These tongues were probably in sheltered bays between distributaries.

The prodelta front advanced westward, and as it did so it buried the recently formed limestone, maintained a habitable sea floor for a short time, but then crowded back the sea. At the distal margins of the prodelta, almost no detrital sediment was deposited in the marine waters, but phosphatic nodules and occasional fossil detritus accumulated. An example of this environment of deposition is the Covel Conglomerate Member of the Carbondale in northern Illinois (Willman, 1939). The sea regressed in some areas in which limestone had recently formed. The limestone emerged and was partly dissolved to form a knobby surface or, in some places, was entirely dissolved.

Mud of the advancing prodelta was eroded by river distributaries, especially where the prodelta built a platform sufficiently high to allow river discharge to cross it in channels. Streams later filled some of the eroded channels with sand. Contacts between the mud and the sand were generally abrupt, as shown by the basal beds of sandstone that commonly contain mudstone pebbles or ironstone nodules derived from underlying prodelta materials. Similar sharp erosional contacts may also have resulted from erosion by rivers on the deltas (Weller, 1930, p. 116-117), for both fluvial valleys and delta distributaries probably existed in the Illinois basin during the time of interval C.

When the supply of sediment diminished, either because of climatic change, diversion of stream courses, lowering of the highland by erosion, or subsidence of the source areas, sediment accumulated less rapidly and was composed of smaller sized particles than before. The land surface probably stood near or slightly above sea level; small ponds or lagoons in which freshwater algal deposits formed were scattered across the delta or alluvial plain. Vegetation probably grew on exposed clays or sands, or even in shallow water, much as mangroves and cypress trees grow today. Evidence is furnished by root impressions that are abundant in underclays that underlie coals or thin carbonaceous films and in sooty streaks where coals are absent.

High on the paleoslopes clays may have been exposed frequently enough to allow some weathering, but across much of the area the presence of vegetation resulted only in some loss of stratification through penetration by many roots. Where clays lack root impressions, waters did not drain sufficiently to permit the growth of vegetation. Such underclays commonly are calcareous and contain pelletlike algal limestone nodules and, in a few places, freshwater ostracodes or ostracitids.

Vegetation doubtless accumulated where water-table conditions were favorable. Such areas probably were on platforms built above the sea bottom by prodelta and delta sedimentation or in stream valleys, in estuaries, or in lagoons. Meanwhile if a temporary marine transgression occurred, a water table sufficiently high to permit peat accumulation probably resulted. Plant detritus that started to accumulate as peat would later have been compressed and altered to coal.

In parts of the Illinois basin, wedges of gray mudstone, generally without marine invertebrates, were deposited immediately following peat accumulation. A bed of this type is the Francis Creek Shale Member of the Carbondale, overlying the Colchester (No. 2) coal in northeastern Illinois. This unit is well known for its clay-ironstone nodules containing fossil plants, arachnids, insects, crustaceans, fish, and amphibia at Mazon Creek (Nee, 1925; Richardson, 1956). This deposit represents the period of maximum regression of the sea. A few miles from Mazon Creek large numbers of soft-bodied marine organisms have been found preserved in similar concretions (Johnson and Richardson, 1970), showing that the prodelta of the Francis Creek Shale Member was advancing into the sea nearby. Special environmental conditions that favored precipitation of iron carbonates or sulfides around both freshwater and marine decaying organisms permitted the preservation of this remarkable record of soft-bodied organisms.

In other areas the first sediment deposited in a re-advancing sea above the peat was a black mud that became a highly fissile, nonplastic mudstone. Such mudstones include fish remains, inarticulate brachiopods, condonoids, and pectenoid pelecypods. Two of these mudstones in interval C of western Indiana were studied intensively by Zangerl and Richardson (1963), who reported a normal marine bottom fauna in the basal inch or so of the mudstone, just above the coal. Above this bottom layer, however, all fossils were nektonic or
planktonic in habit, and nearly all larger fossils showed bite marks. Because of these features a flocculent of seaweed protecting the deeper water from agitation and permitting preservation of paper-thin laminae was postulated by Zangerl and Richardson. Shortly after marine waters had first reached the area, the bottom water apparently became toxic as a result of the accumulation of products of organic decay. The water may have been about 20 feet deep, for the Mecca Quarry Shale Member, which is the black fissile mudstone over the Colchester (No. 2) coal, is not present where the Francis Creek detrital wedge is thicker than about 25 feet. The spread of mudstones that have remarkable similarity in physical, chemical, and organic characteristics and in thickness, across tens of thousands of square miles, is difficult to explain. It indicates clearly, however, that inundation of coal swamps by marine waters took place swiftly and that a vast area was soon covered.

Following black mud deposition, marine waters gradually deepened, the bottoms became habitable, sediments were lighter colored and more calcareous, and marine limestone was deposited. Between the very fissile black mudstone and the limestone a gradational zone, a foot or so thick, is present in most areas. The seas in which the limestone was deposited probably were soon inhabited by vast numbers of shallow-water marine invertebrates. The transgressions were short lived, however, as indicated by the thinness of the limestones in most parts of the Illinois basin. Generally they are less than 1 foot thick, and few exceed 5 feet in thickness.

An advancing front of prodelta mud commonly halted transgression and initiated a new but similar cycle of deposition. No two records of these cyclic transgressions and regressions are alike as regards area inundated, lithology of the marine unit, or magnitude and location of the prodelta-delta complex. Individual differences between these records permit the positive identification of individual strata in interval C of the Illinois basin across thousands of square miles.

**PALEOTECTONIC IMPLICATIONS**

By the early part of interval C time all remnants of pre-Pennsylvanian erosional topography had been buried, and extensive sheets of sediments were forming. Differential upward or downward movement apparently warped this level surface, inducing erosion of newly elevated areas, and accelerating sediment accumulation in areas of subsidence. Furthermore, wedges of sand and mud from newly elevated areas outside the basin advanced into it and produced new depositional forms.

The La Salle and Duquoin flexures that separated bordering shelves from the Fairfield basin evidently were differentially uplifted, and the basin was downwarped, during interval C time (Clegg, 1965; Brownfield, 1954).

Many distant and separate places in the source areas to the west probably were uplifted and yielded detritus to this basin. Marine transgressions and regressions were obviously independent of local tectonic events within the basin.

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D occupies a large area in the deep part of the Illinois basin in central and southern Illinois and along a narrow belt in Indiana, just east of the Illinois line. It also occurs in the Moorman syncline of western Kentucky, south of the Shawneetown-Rough Creek fault zone. Interval D extends in the subsurface nearly 100 miles northward from the main part of the basin west of the La Salle anticline, but in that area the record of interval D is poor. Very small outliers of interval D occur west of the Illinois River in Peoria and Bureau Counties, Ill., and in St. Louis County, Mo., about 15 miles from outcrops in Illinois, and nearly 100 miles from the nearest occurrences in the midcontinent region.

Interval D of the Illinois basin includes the upper part of the Modesto Formation above the base of the Trivoli Sandstone Member, the Bond Formation, and the lower part of the Mattoon Formation to the base of the Omega Limestone Member (Kosanke and others, 1960) (table 14). All of these formations are in the McLeansboro Group (pl. 13). The lower boundary of the interval at the base of the Trivoli Sandstone Member is a little higher stratigraphically than the base of the Hepler Member of an unnamed formation in the Pleasanton Group, which forms the lower boundary of the Missouri Series in the Missouri-Iowa region. In the Illinois basin the equivalent of the Hepler Member is the sandstone below the Ecline Limestone Member of the Modesto Formation, but because this sandstone is very discontinuous and the Trivoli Sandstone Member is widespread and generally conspicuous, the Trivoli is arbitrarily selected for use. Furthermore, its distribution has already been mapped in the southern part of the basin (Andresen, 1961).

In general, interval D corresponds to the Missouri Series of the midcontinent region and is characterized by early fusulinids of the genus Triticiticites that are referred to the subgenus Kansanelia (Thompson, 1957, p. 299-301). Fusulinids of this type have been found in several of the marine limestones of interval D in the Illinois basin (Dunbar and Henbest, 1943) but not in the Cramer, Carlisle, and Macoupin Limestone Members of the Modesto Formation, which form marine zones in the lower part of the formation.

In Indiana interval D consists of the Ditney and
Table 14.—Stratigraphic units of less than formation rank mentioned in text for interval D and parts of adjacent intervals but not shown on plate 13: Illinois basin region

[Modified from Kosanke and others (1960)]

<table>
<thead>
<tr>
<th>Interval</th>
<th>Illinois</th>
<th>Indiana</th>
<th>Kentucky</th>
</tr>
</thead>
<tbody>
<tr>
<td>E</td>
<td>Omega Limestone Member</td>
<td>Merom Sandstone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mattson Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td>D</td>
<td>Bond Formation</td>
<td>Wabash Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bond Formation</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Millersville Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bunje Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mount Carmel Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Shoal Creek Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Clintonville Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Carlinville Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Cramer Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Trivoli Sandstone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Exline Limestone Member</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C</td>
<td></td>
<td>West Franklin Limestone</td>
<td>Madisonville Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Member</td>
</tr>
</tbody>
</table>

1 Named units not in text are omitted.

Wabash Formations of the Conemaugh Series. The original top of interval D is not present, unless the Merom Sandstone, a channel sandstone here included in interval D, corresponds to the basal sandstone of the Virgil Series of the midcontinent region.

In western Kentucky interval D corresponds to the part of the Limestone Formation that lies above the top of the Madisonville Limestone Member.

In St. Louis County, Mo., most of the thin sequence of strata referred to interval D is assigned to the Pleasonton Group, but the sequence may also include basal units of the Kansas City Group.

Upper boundary of interval D

In the Illinois basin the equivalent of the Missouri-Virgil boundary in the midcontinent region should correspond to (1) the base of the predominantly detrital Douglas Group in Missouri and (2) the near disappearance of Kansanella, subgenus of Triticites, among the fusulinids. The position of the upper boundary of interval D in relation to the series boundary, based on these criteria, is not at present positively determined.

The base of the Omega Limestone Member of the Mattson Formation is probably a little above the Missouri-Virgil boundary but is used for the top of interval D because this limestone constitutes a key bed that can easily be traced throughout most parts of the basin.

Thickness trends

Interval D is completely preserved only in the deeper parts of the Illinois basin, west of the La Salle anticline, from Douglas County in the north to Wayne County in the south. The total thickness in this area ranges from about 400 feet in the north to more than 800 feet in Clay County farther south.

In the small and nearly detached area of interval D in the Moorman syncline area of Webster and Union Counties in western Kentucky, and eastern Gallatin County in Illinois, the interval has a maximum total thickness of 895 feet. This thickness is attained in Union County, Ky.

Where interval D is present in the remainder of the basin, its upper part and interval E have been removed by erosion. The amount of interval D remaining in these areas has been determined both by structural position and by local depth of pre-Pleistocene erosion. The interval D isopach map (pl. 8A) shows abrupt thickening westward, west of the La Salle anticline from Champaign County, Ill., south to Lawrence County, and also thickening eastward, east of the Duquoin monocline in Jefferson and Franklin Counties of southern Illinois. Thin sequences east of the La Salle anticline and west of the Duquoin monocline represent less complete parts of interval D than are present in the basin between structures; thus, the apparent thickening in the basin shows more the effect of post-Pennsylvanian erosion than of differential downwarping during the time of interval D.

Lithofacies trends

Interval D in the Illinois basin includes sandstone, mudstone, limestone, coal, and underclay. At a few localities in northern and central Illinois, limestone makes up more than 20 percent of the total thickness of the interval (pl. 8B). This high percentage results from the presence of several limestone units—including the Shoal Creek Limestone Member of the Bond Formation of northern Illinois, two limestones near Pontiac, Livingston County, and the Millersville Limestone Member of central Illinois—that locally are 25-50 feet thick and are the thickest Pennsylvanian limestones in the basin.

Except in northern and central Illinois, terrigenous sediments compose more than 80 percent of the total
thickness of interval D. The thickness of the mudstone—including gray and black marine mudstone, gray nonmarine mudstone, a little red mudstone, and underclay—exceeds in most places the thickness of sandstone, and the ratio of mudstone to sandstone ranges from 1:1 to 4:1 and locally may exceed 4:1. Sandstone is thicker than mudstone only in limited, elongate areas where one or more discontinuous sandstone beds are exceptionally thick because they were deposited in a channel. Examples of such areas are three northeast-trending bands in southeastern Illinois (pl. 8B).

A generalized section of interval D (part of McLeansboro Group) in Illinois (Kosanke and others, 1960, pl. 1) shows sandstone at 10 stratigraphic positions, limestone at 10, mudstone at 18, and coal at 12. Thus, as in other Pennsylvanian intervals of the Illinois basin, the total lithologic makeup of interval D varies greatly from place to place, according to the thicknesses of a large number of different rock units, each of which has appreciable local variations. Furthermore, because at most places the upper part of interval D has been removed by erosion, the lithologic proportions of the remainder may not be representative of the original sequence.

The most conspicuous sandstones, which are the Tricoli Sandstone Member and the Inglefield Sandstone near the base of the interval, the Mount Carmel Sandstone Member and the sandstone below the Buje Limestone Member near the middle, and the Merom Sandstone near the top, vary greatly in thickness, largely because they fill integrated systems of stream channels. The principal channels, except for the one filled by the Merom Sandstone, have been mapped (Andersen, 1961; Glover, 1964; Horne, 1965).

The most conspicuous limestone units of interval D are the Shoal Creek Limestone Member in the lower one-fourth of the interval and the Millersville near the middle. These are probably equivalent, respectively, to the Cambridge and Ames Limestone Members of the Conemaugh Formation of the Appalachian region (table 7) and to the Winterset Limestone Member of the Den- nis Limestone and the Wyandotte Limestone of the Missouri-Iowa region. (See chap. F.)

Mudstone of interval D is mostly gray, but black fissile mudstone also occurs at seven or more stratigraphic positions. A few units of red or mottled red and gray mudstone also are known, but their total thickness is very small and none extends across the basin as do the red beds in the Pittsburgh Formation of Missouri age in the northern Appalachians.

The coals are generally thin and moderately discontinuous and have been mined only to a limited extent. Several coals have fairly thin but widespread nonlaminated underclays that contain root impressions.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Distribution patterns of the sandstones show that detritus of interval D was introduced into the Illinois basin largely from the northeast or north (Andersen, 1961; Glover, 1964; Palominario-Cardenas, 1963; and Horne, 1965). This conclusion is supported by a limited number of measurements of dip directions in crossbedding (Potter, 1962; Andersen, 1961). Sand sources presumably were distant from the basin, because the sandstones contain much mica, feldspar, and argillaceous material probably derived from crystalline source areas in the Canadian Shield or in the northern Appalachian highlands of New England or eastern Canada. Probably because of the distance between source and depositional basin, the sandstones include no particles larger than medium sand grains, except for mica flakes and large fragments of mudstone and ironstone apparently derived nearby.

ENVIRONMENTS OF DEPOSITION

At the beginning of interval D time the Illinois basin was largely emergent, according to Andersen (1961) and Glover (1964), and a system of subaerial river channels provided drainage. These channels subsequently were filled with sand. They are believed to have been contemporary with channels in Missouri that were filled with sand to form the Moberly and Warrensburg Sandstone Members in an unnamed formation of the Pleasanton Group. The east-central part of the United States appears to have been more emergent at this time than at any other time during the Middle Pennsylvanian.

Above the channel sandstones a succession of eight deltas occurs in the lower half of interval D. In each delta the initial sediment is laminated gray mudstone that contains beds or nodules of ironstone and that thickens to the east or northeast as a prodelta mud. Channels in the mudstone contain distributary sandstone lenses that form a bird-foot pattern and that provide, by their geometry, convincing evidence of drainage from the northeast or north. Each deltaic deposit represents the initial phase of a cycle of sedimentation.

Sandstones decrease in coarseness upward and grade into fine silty mudstones, which in turn grade upward into nonlaminated underclays containing numerous root impressions. The underclays are generally overlain by thin coals or by carbonaceous films that indicate coal horizons. The coals and underclays are nearly coincident areally with the delta platforms on which they rest, and the coals locally grade laterally into marine mudstones or into thin limestones in the western part of the basin away from the deltaic areas.

For the Illinois basin, 46 environmental maps for the
lower half of interval D up through the Millersville Limestone Member of the Bond Formation have been prepared by students at the University of Illinois (Palomino-Cardenas, 1963; Glover, 1964; and Horne, 1965). Environmental studies of a small succession in the upper part of interval D have been made by Horne (1968). These maps give a representative picture of the environmental history of the whole interval but show that limestone deposition was probably more widespread during the time represented by the lower part of the interval than later.

Commonly a coal bed or associated limestone is overlain by a black fissile marine mudstone containing inarticulate brachiopods, conodonts, fish remains, and pectenoid pelecypods. This combination of animals suggests that life existed in upper zones of the shallow water, where the animals swam or floated, but not on the bottom, where conditions apparently were toxic.

The records of poorly ventilated seas generally are limited to a foot or so of sediment, which is succeeded by calcareous mudstones or limestones that contain normal marine benthonic forms, such as articulate brachiopods, crinoids, corals, mollusks, and fusulinids. These limestones record widespread marine transgressions from north of the Ozark uplands, which lie to the west. Individually they seem to correlate with thicker and purer limestones of the Forest City basin (fig. 15) in the northern midcontinent region. The more conspicuous limestones, such as the Shoal Creek and Millersville, may also correlate with somewhat thinner limestones of the northern Appalachian basin. Thus, during the most extensive transgressions the sea may have spread eastward at least 1,000 miles, deposited sediments 5-40 feet thick, and then withdrawn.

The terminal position of each marine transgression appears to have been controlled by a wedge of prodelta mud over which the sea could not easily spread. Maps of the lower half of interval D show a total of 14 marine limestones representing transgressions during which the sea reached the Illinois basin; the upper part of the interval, which is unmapped, probably contains the record of six or seven additional transgressions.

**PALEOTECTONIC IMPLICATIONS**

Deformation of the Illinois basin during interval D time is suggested by the abrupt thickening shown on the isopach map (pl. 84) near the La Salle anticline and the Duquoin monocline. Thicknesses preserved on the borders of the deep basin do not represent the original thickness of the interval, so structural movements can be determined only by preparing detailed isopach maps of small subdivisions of interval D in the basin and on the flanks. Such studies (Clegg, 1965) seem to prove that progressive downwarping of the deep basin was taking place during most of interval D time. A study of

thickness variations at localities where the entire interval is present suggests that downwarping was greater in the south than in the north; this relationship is illustrated by comparing thicknesses of more than 800 feet of sediment deposited in Clay County, Ill., and Union County, Ky., with thicknesses of 400-500 feet farther north. Some differences in thickness must also be accounted for by the greater compactability of mudstones than of sandstones.

The successive development of 8 to 15 deltas in the same part of the basin might be explained by repeated orogeny in areas to the east or north, by intermittent arrival of sediment in the basin as a result of climatic cycles, or by the lateral shift of the outlets of major rivers. A shift of river outlets was proposed by Swann (1964) to account for the somewhat similar Mississippian deltaic sediments of the Illinois basin.

The alternate transgressions and regressions of the sea eastward and westward, respectively, for hundreds of miles could hardly result from local tectonism. Possible mechanisms producing this effect by causing eustatic changes in sea level include extensive deformation of the ocean floor and alternate growth and waning of Gondwana glaciers in the southern hemisphere. According to the second theory, the general emergence of the central United States at the beginning of interval D time may indicate a very extensive episode of glaciation in the southern hemisphere and a consequent lowering of sea level. Formation of the Millersville Limestone Member may correspond to a maximum deglaciation.

**INTERVAL E**

**FORMATIONS INCLUDED**

Interval E, which consists of the youngest Paleozoic strata in the Illinois basin, occurs in two principal small areas and in two outliers. The largest of the areas is in east-central Illinois and extends from northern Douglas County southward to northern Richland County, a distance of about 75 miles. On the east this area is bounded by a relatively abrupt upturn of strata along the west flank of the La Salle anticline; the west border, where strata dip gently eastward, is more irregular and extends to eastern Shelby and western Effingham Counties.

The second area of interval E in the Illinois basin is in the lowest part of the Moorman syncline immediately south of the Shawneetown-Rough Creek fault zone, in Union, Webster, and Hopkins Counties, western Kentucky, and Gallatin County, southeastern Illinois. Because the syncline is asymmetrical, strata on the south side dip moderately northward, but those on the north are upturned sharply in the fault zone.

The two outlying areas of outcrop E are in Clay and Wayne Counties and in Edwards County, Ill.

All strata in the Illinois basin referred to interval E
belong to the upper part of the Mattoon Formation and McLeansboro Group, of Illinois, and the Henshaw Formation of western Kentucky (pl. 13). Where the Omega Limestone Member is recognized, the lower boundary of interval E is placed at its base. In Kentucky the bottom of the Dixon Sandstone Member is the base of the Henshaw Formation and is considered the base of interval E (table 15).

Strata in the Illinois basin are referred to interval E because the Omega, Shumway, and Greenup Limestone Members of the Mattoon contain fusulinids that belong to the genus Triticites and are correlated with Virgil forms of this genus in the midcontinent region (Dunbar and Henbest, 1943, p. 27-28). Equivalents of both the Shawnee and Wabaunsee Groups of the Virgil Series probably are present in the Illinois basin. The Omega, Shumway, and Greenup Limestone Members have not been recognized in the western Kentucky outlier in the Moorman syncline, but equivalent strata are certainly present, as indicated by the considerable thickness of strata above known key beds.

UPPER BOUNDARY OF INTERVAL E

In Illinois, Pleistocene glacial drift directly overlies interval E (pl. 12), and in western Kentucky, Pleistocene loess and Quaternary alluvium (not shown on pl. 12) overlie interval E. Permian rocks were reported long ago from along the Salt Fork River, Vermilion County, eastern Illinois, where amphibian and reptilian bones were found in a red clay zone (Case, 1900). This clay is now referred to the lower part of interval D, however, so no rocks of established Permian age are known in the Illinois basin.

THICKNESS TRENDS

All measurements represent only thicknesses preserved in eroded remnants of interval E, so their variation is more indicative of the amount of structural downwarp in post-Pennsylvanian time than of differences in the original or depositional thickness. Most of the thicknesses are determined from electric logs of oil exploration wells. The top of the log coincides with the base of the surface pipe, which is driven from a few feet to more than 100 feet into Pennsylvanian bedrock; therefore, most measurements are too low, because no allowance is made for strata of the interval that were penetrated by the surface pipe. Thus, some strata belonging to interval E possibly are present in the area between the two larger belts of interval E in southeastern Illinois.

Maximum thickness in the northern area in Illinois is a questionable 512 feet at locality 698 (pl. 1). Maximum thicknesses of 214 feet at locality 796 and 20 feet at locality 879 indicate much thinner sections of interval E in the small Illinois outliers. In the Moorman syncline, at locality 141, Union County, Ky., a maximum thickness of 800 feet was reported on the basis of a diamond drill core which was carefully logged by geologists (Smith and Smith, 1967). The log shows the Millersville Limestone Member at a depth of about 550 feet and includes more of the upper part of the Pennsylvanian System and of interval E than does any other record in the basin.

Well records in Illinois show an eastward increase in the thickness of interval E that results entirely from the preservation of strata to the east which have been removed by erosion at localities farther west.

LITHOFACIES TRENDS

In all records of the partial sections that compose interval E in the Illinois basin, limestone totals less than 20 percent. Lithologic variations range generally from a sandstone-mudstone ratio of 1:1 to less than 1:4. In the southern part of the main remnant in Illinois and in the Moorman syncline of Kentucky, sandstone exceeds mudstone; but in nearly all localities where this relation exists, less than 200 feet of interval E remains. Thus, sandstone probably is more prevalent in the lower part of the interval than in the higher part.

The detailed succession of units composing interval E in the Illinois basin is not yet well known, and some limestones that are sufficiently conspicuous to be good key beds in the northern part of the outcrop area appear to either wedge out or change facies farther south. Information is poor largely because interval E is covered in most places with glacial drift, and more than about 25 feet of strata is rarely exposed by dissecting streams. The piecing together of partial sections into a generalized section has not been done. The marine Omega, Shumway, Effingham, Bogota, Greenup, Woodbury, and Reissner Limestone Members of the Mattoon Formation (Kosanek and others, 1960) are the best known units, but relations between some of them are

### Table 15.—Stratigraphic units of less than formation rank mentioned in text for interval E but not shown on plate 13; Illinois basin region

<table>
<thead>
<tr>
<th>Interval</th>
<th>Illinois</th>
<th>Western Kentucky</th>
</tr>
</thead>
<tbody>
<tr>
<td>E</td>
<td>Reissner Limestone Member</td>
<td>Dixon Sandstone Member</td>
</tr>
<tr>
<td></td>
<td>Woodbury Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Gila Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Greenup Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Bogota Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Effingham Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Shumway Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Omega Limestone Member</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Henshaw Formation</td>
<td></td>
</tr>
</tbody>
</table>

*Named units not in text are omitted.*
uncertain and they may not occur from the base up in the order listed (table 15).

Representative thicknesses and frequency of occurrence of the principal lithologic types are illustrated by the Union County, Ky., diamond drill core (locality 141) as follows:

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Number of units</th>
<th>Total thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandstone</td>
<td>15</td>
<td>325</td>
</tr>
<tr>
<td>Mudstone</td>
<td>25</td>
<td>444</td>
</tr>
<tr>
<td>Limestone</td>
<td>8</td>
<td>25</td>
</tr>
<tr>
<td>Coal</td>
<td>13</td>
<td>6</td>
</tr>
</tbody>
</table>

**SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES**

No information is on record regarding directions of dip in sandstone foreset beds of this interval, nor have environmental maps been prepared for any of the sandstones or mudstones. Until such studies have been made, neither directions of sediment transport nor possible sources can be determined. The detritus that formed various discontinuous massive sandstone units probably entered the basin from the north, northeast, or east, if slope trends of earlier intervals still prevailed.

**ENVIRONMENTS OF DEPOSITION**

During the time of interval E the Illinois basin seems to have been inundated by the sea at least seven times, as indicated by the marine faunas of the Omega, Shumway, Effingham, Bogota, Greenup, Woodbury, and Reiser Limestone Members. The Gila Member, between the Greenup and Woodbury, has a brackish-water fauna. Coals occur beneath some of the limestones, notably the Omega and Woodbury, and a conspicuous black fissile mudstone lies below the Shumway. Little is known about the lateral distribution of mudstone and sandstone, but because these rocks resemble in lithology the corresponding types of interval D, they probably also represent fluvialite or deltaic channel deposits overlying prodelta mudstone. Black fissile mudstone apparently indicates toxic bottom conditions similar to those that prevailed during the times of several earlier intervals.

The sea probably transgressed from the west, and the Omega, Shumway, and Effingham Limestone Members may record the same transgressions, respectively, as do the Oread, Lecompton, and Deer Creek Limestones of the Shawnee Group in the midcontinent region. The Bogota, Gila, Greenup, and Reiser Members, if each represents a different marine transgression, probably correlate with limestones deposited during transgressions in late Shawnee (Topeka Limestone) and early Wabamunsee time in the midcontinent region.

**PALEOTECHTONIC IMPLICATIONS**

The scarcity of knowledge about interval E in the Illinois basin prevents drawing confident conclusions regarding earth movements in the basin or in surrounding source areas. The general similarity of the lithologic types and the sequences to those of interval D suggests that detrital sediments continued to be introduced periodically through stream channels in response to periodic elevation of distant sources. Furthermore, the basin must have continued to subside as a total of at least 800 feet of sediment accumulated in western Kentucky.

**TOTAL THICKNESS OF PENNSYLVANIAN ROCKS**

The thickness of Pennsylvanian rocks in various parts of the Illinois basin has been determined partly by structural features of that region, partly by relief of the pre-Pennsylvanian surface on which deposition occurred, and partly by post-Pennsylvanian erosion that has removed some of the system in many places.

Two of the principal structural features in the Illinois basin are the La Salle anticlinal belt (Cady, 1920; Green, 1957) and the Shawneetown-Rough Creek fault zone. The La Salle anticlinal belt extends for a distance of about 220 miles from central and northern Illinois about S. 15° E. to Wabash County, Ill., near the Indiana line. Many elements of this belt show steep west flanks and very gently dipping east flanks. The belt is a system of en echelon folds perhaps related to movement along a basement fault.

The Shawneetown-Rough Creek fault zone is an east-west structure which probably connects across central Kentucky with the Paint Creek-Warfield fault zone (fig. 7) of eastern Kentucky. It affects the Pennsylvanian strata from the east border of the basin westward for about 100 miles into southeastern Illinois. Directly south of the fault zone is the Moorman syncline, where, in Union County, Ky., the Pennsylvanian strata are at least 3,055 feet thick and form the thickest section known in the basin.

In southwestern Illinois the Du Quoin monocline trends nearly northward from approximately the west end of the Rough Creek fault system. The east flank has a moderately steep dip, but on the west side the rocks dip gently away from the axis of the monocline. This structure becomes less marked northward. A deeper segment of the Illinois basin, between the La Salle anticline and the Du Quoin monocline, is called the Fairfield basin.

Pennsylvanian strata in most parts of the Illinois basin represent an incomplete record of that system. They are erosional remnants of an originally thicker and more widespread sequence of deposits. All intervals of the system are represented, but interval A is limited to approximately the southeastern third of the basin, and interval E is preserved only in two relatively small areas in east-central Illinois and western Kentucky. A series of outliers that extends northwestward from St. Louis,
Mo., and another series in southeastern Iowa span much of the area between the Illinois basin and the Missouri-Iowa region. Several key strata can be traced through these outliers, which proves the former continuity of sediments between the two basins.

Nearly all of the basin in Illinois and its north half in Indiana are covered with Pleistocene glacial drift, which ranges in thickness from a few feet to more than 400 feet. The Pennsylvanian rocks are separated from underlying and overlying strata by conspicuous erosional unconformities. The erosional surface at the base of the Pennsylvanian in the southern part of the basin includes a system of generally southwest-trending steep-sided valleys, some as deep as 450 feet (Siever, 1951; Bristol and Howard, 1971). The present-day bedrock topography of Illinois shows a system of preglacial valleys that are now generally filled with thick glacial drift (Horberg, 1950). Both surfaces are illustrated on maps based on very large numbers of oil, water, and coal test records.

The two drainage systems—pre-Pennsylvanian and pre-Pleistocene—are not superposed, so the patterns of thickness that result from using the erosional surfaces as bounding surfaces for an isopach map (pl. 11) are very complicated. A much denser control net than the one used in this publication would be necessary to show clearly by isopachs the effects of these two sets of unrelated erosional valleys. Figure 14 shows the outlines of the principal known pre-Pennsylvanian and pre-Pleistocene drainage courses. The total thickness of Pennsylvanian strata may increase locally as much as 450 feet where basal channel-fill deposits are present, or it may thin locally by as much as 350 feet where Pennsylvanian strata were eroded by a major pre-Pleistocene stream.

On the basis of differences in thickness, the Illinois basin can be divided into four parts: the two shelflike areas east of the La Salle anticline and west of the Duquoin monocline, the Fairfield basin between them, and the Moorman syncline south of the Shawneetown-Rough Creek fault zone.

In the shelf area east of the La Salle anticline and north of the Shawneetown-Rough Creek fault zone, rocks dip gently westward from outcrops in Indiana. Pennsylvanian rocks thicken at the rate of 20-25 feet per mile to a maximum of about 1,700 feet near the southwest corner of Indiana. The north part of the shelf area is divided into two parts by the north-trending Oakland anticline, which lies 5-25 miles east of the La Salle anticlinal belt and has Pennsylvanian rocks less than 400 feet thick.

The Fairfield basin is roughly triangular in outline because the La Salle and Duquoin structures tend to converge northward. In the steepest part of the east flank of the basin near Tuscola, Ill., the Pennsylvanian strata thicken from a few feet to 1,400 feet in about 4 miles. Along the steepest part of the Duquoin monocline, the Pennsylvanian thickens from 500 to 1,400 feet in about 8 miles. The thickness of Pennsylvanian strata along the axis of the Fairfield basin increases from about 1,000 feet in De Witt County in the north to about 2,400 feet in Hamilton County in southeastern Illinois. Several minor anticlinal trends are shown by belts of thinner Pennsylvanian rocks.

In the shelf area west of the Duquoin monocline, Pennsylvanian strata are eroded in areas along the Illinois River, including the valleys that carried the preglacial Mississippi drainage and the preglacial Ohio (Mahomet or Teays) drainage. Locally these valleys are filled with nearly 400 feet of glacial drift. West of the Illinois River the maximum thickness of the Pennsylvanian is 500-600 feet. East of the Illinois River valley Pennsylvanian strata steadily increase in thickness eastward at an average of about 15 feet per mile. This rate of thickening increases to 50 and locally to 100 feet per mile at the edge of the Fairfield basin.

The Moorman syncline is a narrow asymmetrical belt
along which Pennsylvanian rocks on the steeper north flank thicken toward the axis locally at the rate of 400-500 feet per mile, and rocks on the more gently dipping south flank thicken toward the axis at the rate of 50-80 feet per mile. Thicknesses along the axis reach a maximum of more than 3,000 feet in Union and Webster Counties, western Kentucky.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

**UNITS OVERLYING PENNSYLVANIAN SYSTEM**

In the Illinois basin, igneous rocks consisting of mica peridotite dikes, sills, and plugs were intruded prior to the Late Cretaceous. The time during which intrusion could have taken place in southern Illinois extends from late Pennsylvanian to Late Cretaceous. The igneous intrusions are in contact with Mississippian rocks in the faulted mineralized districts of southern Illinois and western Kentucky, and several dikes have been found in coal mines in southern Illinois. One such dike, cutting a coal vein of the Pennsylvanian interval C, in Saline County, was described by Clegg (1955). In the Omaha oil field, in Gallatin County of southeastern Illinois, a laccolithic dome of mica peridotite was intruded into rocks of Middle and Late Pennsylvanian age (English and Grogan, 1948).

Cretaceous rocks overlap Pennsylvanian and Mississippian rocks at the northern end of the Mississippian embayment a few miles south of the southern border of the Illinois and western Kentucky coal fields. At some places Pennsylvanian outliers probably occur beneath the Upper Cretaceous McNaury Formation. Cretaceous strata have recently been recognized in a small area in western Illinois northwest of St. Louis (Frye and others, 1964). Most of these Cretaceous rocks rest upon the Mississippian, but they are mapped (Willman and others, 1967) as overlapping a short distance onto the Pennsylvanian rocks in Pike County, Ill.

Glacial drift covers most of the Illinois coal field, but the western Kentucky part of the Illinois basin and the southern counties of Indiana lack glacial drift. Pleistocene loess, lake beds, and outwash deposits cover substantial areas beyond the borders of the glacial drift and effectively conceal Pennsylvanian rocks across large areas. These periglacial Pleistocene deposits are not shown on the geologic map of overlying units (pl. 12).

**PALEOTECTONIC IMPLICATIONS**

In the Illinois basin, sedimentation probably continued from latest Pennsylvanian into Permian time, but all strata of Permian age have subsequently been removed. Later, in the Triassic or Jurassic, renewed differential downwarping caused the basin to deepen between the La Salle anticline and the Duquoin monocline. At this time the southern margin of the basin was upturned along the Shawneetown-Rough Creek fault zone, on which displacements were several hundred feet to more than 1,000 feet. Until this faulting occurred, the southern margin of the basin probably had a southward slope; after faulting, downwarping took place within the basin and south of it. Elevated areas were eroded and beveled. The surface of the region was worn to a fairly smooth plain, which was then buried beneath Upper Cretaceous and lower Tertiary sediments at the head of the Mississippi embayment. Still later, these deposits apparently were removed by erosion within the area of the Illinois basin, except for possible outliers in southern Illinois. A system of deep valleys then was cut into the Pennsylvanian and other Paleozoic rocks of the basin prior to Pleistocene glaciation (Horberg, 1950).

**REFERENCES**


Miller, A. M., 1910a, Coals of the lower measures along the western border of the eastern coal field: Kentucky Geol. Survey Bull. 12, 83 p.

1910b, Evidence that the Appalachian and central coal fields were once connected across central Kentucky: Geol. Soc. America Bull., v. 20, p. 621-624.


Richardson, E. S., Jr., 1956, Pennsylvanian invertebrates of the Mazama Creek area, Illinois—Introduction; Insects; Marine fauna: Fieldiana Geology, v. 12, nos. 1-3, p. 3-67.


 Wanless, H. R., Baroffio, J. R., and Trescott, P. C., 1969, Conditions of


Worthen, A. H., 1888, Geology of Alexander County, Union County, Jackson County, Perry County, Jersey County, Greene County, Scott County [Ill.]: Illinois Geol. Survey, v. 3, p. 20-144.


Missouri and Iowa

By HAROLD R. WANLESS

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-F
CONTENTS

Abstract ......................................................... 97
Region defined .................................................. 98
Paleogeology ..................................................... 98
Units underlying Pennsylvanian ........................................ 98
Lower boundary of the Pennsylvanian .................................... 99
Paleotectonic implications ........................................ 99
Interval A .......................................................... 99
Interval B .......................................................... 99
  Formations included ........................................... 99
  Stratigraphic relations ........................................ 99
  Upper boundary of interval B .................................... 101
  Thickness trends ............................................... 102
  Lithofacies trends ............................................ 102
  Sources and environments of deposition .......................... 103
  Paleotectonic implications .................................... 103
Interval C .......................................................... 104
  Formations included ........................................... 104
  Upper boundary of interval C ................................... 104
  Thickness trends ............................................... 104
  Lithofacies trends ............................................ 105
  Sources and environments of deposition .......................... 105
  Paleotectonic implications .................................... 106
Interval D .......................................................... 106
  Formations included ........................................... 106
  Stratigraphic relations ........................................ 107
  Upper boundary of interval D ................................... 107
  Thickness trends ............................................... 107
  Lithofacies trends ............................................ 107
  Sources and environments of deposition .......................... 108
  Paleotectonic implications .................................... 109
Interval E .......................................................... 109
  Formations included ........................................... 109
  Upper boundary of interval E ................................... 109
  Thickness trends ............................................... 109
  Lithofacies trends ............................................ 110
  Sources and environments of deposition .......................... 110
  Paleotectonic implications .................................... 111
Total thickness of Pennsylvanian rocks .............................. 111
Geologic units directly above Pennsylvanian System ................. 112
Units overlying Pennsylvanian ...................................... 112
Paleotectonic implications ....................................... 112
References .......................................................... 113

ILLUSTRATIONS

[For listing of plates (in separate case) see volume "Contents"]

Figure 15. Map showing tectonic elements in the Missouri and Iowa region in Pennsylvanian time .......................... 98
16. Map showing geographic features and localities in the Missouri and Iowa region mentioned in text ......................... 100
Pennsylvanian rocks of the midcontinent region extend from western North Dakota to Texas and from Colorado to the Mississippi River. The part of the midcontinent region treated here embraces Missouri and Iowa, except for small areas near St. Louis, Mo., and Muscatine, Iowa, that belong in the Illinois basin. The border of Pennsylvanian rocks extends southeast from northwestern Iowa to about the southeast corner of Iowa, then south to a position northwest of St. Louis, and from there it curves west and southwest around the borders of the Ozark uplift. There are also many outcrops of the Pennsylvanian in the Ozarks, but most are fillings of solution cavities. Northwestern Missouri and southwestern Iowa constitute the eastern part of the Forest City basin. This basin is bordered on the east and southeast by shelf areas west of the Mississippi River arch. A series of outliers in southeastern Iowa and northeastern Missouri are mostly thin remnants on the Mississippi River arch, although others are sink fillings or remnants of channel sandstones. Southwestern Missouri includes the northeast border of the Cherokee basin, separated from the Forest City basin by the weak Bourbon arch.

Before the Pennsylvanian or during the time of Pennsylvanian interval A, the Nemaha anticline, from southeastern Nebraska southwest to Oklahoma City, was elevated and bordered on the east by a fault scarps locally more than 1,000 feet high. The area to the east was downwarped, forming the asymmetrical Forest City basin with a long gentle east flank and a precipitous western slope. About 1,700-2,400 feet of Pennsylvanian strata remains in the deeper part of the basin, indicating downwarp of about that amount. The Mississippi River arch had high been most of the time since the Ordovician and was mantled by thin sheets of middle Paleozoic rocks. The Ozark uplift also had stood above surrounding basins during most of the time since the Early Ordovician, and its highest parts were probably more than 1,000 feet above surrounding plains. Shelf areas with low relief and thin sediment sequences bordered the Forest City basin on the east and the Ozark uplift on the north.

The Missouri-Iowa region was a part of the central stable craton in which no sediments seem to have been deposited in A or A1, except along the Arkansas border south of the Ozark uplift; there, thin remnants of subinterval A1 sandstones remain on a few hilltops.

During this period of subaerial exposure, stream channels probably developed on the pre-Pennsylvanian surface, but sink and cave features mask most irregularities that might be attributed to channeling. In the Ozark uplift, rocks of post-Early Ordovician age have generally been removed, and on the northern flank, cave and sink fillings to depths of 100 feet or more are common. Some are still roofed by 50-100 feet of Lower Ordovician or Upper Cambrian dolomites.

During interval B a sequence, locally more than 400 feet thick, of dark-gray mudstones, thin sandstones, a few coals, and at least one limestone accumulated in the Forest City basin, and fossiliferous limestone or chert containing Atoka (interval B) fossils formed at several places on the shelf area of southwestern Missouri. The thick fill of the Forest City basin does not crop out, and no cores were available for its study. It was probably an undrained basin that was poorly ventilated and had occasional influxes of marine water from the south across the narrow Bourbon arch.

Solution-cavity fillings and sheetlike basal sediments on the Mississippi River arch, referred to as the Cheltenham Clay of interval B, are composed of kaolinitic clays resulting from weathering under subaerial conditions in subtropical climates. South of the Missouri River in the northern and especially the northeastern Ozarks, many solution cavities are filled with clay, including diaspore (hydrated aluminum oxide) and kaolinite. A little farther west many sinkholes and caves are filled with pocket coals locally 80 feet or more thick. Still farther south and higher on the slopes of the Ozark uplift, numerous solution cavities were filled with deposits of hematite, pyrite, and sulfides of copper, lead, and zinc. Most cavity fillings show evidence of having settled or slumped farther into the cavity after initial deposition, indicating continued solution contemporary with cavity filling. This accounts for the extraordinary thickness of deposits of coal, hematite, and pyrite. A few sinks show coals with roofs of fossiliferous marine limestones or cherts, some with fossils of Atoka age (interval B), and one or two with fossils of Des Moines age (interval C).

Interval C blankets the region discussed in this chapter excluding most of the Ozarks. Except in the Forest City and eastern Cherokee basins, the initial sediments were deposited during middle or, on the Mississippi River arch, even late interval C time. Aggregate thickness of Interval C is 50-100 feet on the Mississippi River arch but reaches a maximum of more than 600 feet in the deeper part of the Forest City basin in Missouri. On the Mississippi River arch the interval consists largely of marine limestones alternating with underclays and generally sooty films which mark the positions of coal beds that, in the Illinois basin and farther west in Missouri, are well developed. The limestones indicate that the area was flooded numerous times by the sea. Wedges of sandstone are less abundant than in the Illinois basin. Many persistent lithologic units in interval C of the eastern midcontinent are easily traced and matched with correlatives in Kansas and Oklahoma to the west and in the Illinois basin to the east. The interval contains 22 separately recognized limestone beds and 22 coal beds.

Interval D has a maximum thickness of 600 feet in the Forest City basin. Except for a detrital unit at its base, the interval contains little sandstone. Early in deposition of the interval, sandstone was deposited by one west-flowing and one south-flowing stream in channels cut so
deeply as to reach the base of the Pennsylvanian at some places. Probably the two streams joined and drained southward. The interval contains limestones and four thin and local coals. Seven of the limestones underlie black fissile shales such as overlie numerous coals in the Illinois basin, and these probably were formed offshore while coal was accumulating on the coastal plain farther east. In a few locations in northwestern Missouri, limestone composes more than 50 percent of interval D.

The maximum thickness of interval E recorded in Missouri is 665 feet. All Missouri and Iowa sections are incomplete, but complete sections remain a short distance west in Kansas and Nebraska; by extrapolation from these sections the original maximum thickness for Missouri and Iowa is estimated at about 725 feet. Like interval D, interval E (Virgil) begins with a detrital unit, the Douglas Group, in which channel sandstones are prominent in northwestern Missouri. In the middle of interval E are four repetitions of the complex rhythm which has been named a megacycle. These contain the greatest quantity of limestones, of which there are 34 beds in interval E.

Sources of detrital sediments of Pennsylvanian age were distant, and sandstones are much less common in the Missouri-Iowa region than farther east. Principal sources, like sources for other intervals, probably were in the Precambrian shield of Canada to the north. Some sand, however, apparently reached the area from the northeast by way of the Illinois basin. A little sand may also have been derived from the Nemaha anticline to the west. During the time of intervals D and E some sand deposited in southwest Missouri was derived from the ancestral Ouachita Mountains to the south and the region of the Arbuckle Mountains to the southwest. Chert gravels derived from dissolved cherty carbonate rocks accumulated in the Ozark area and on the Mississippi River arch.

Sedimentation was cyclic in intervals C, D, and E, but marine limestones and mudstones predominate over nonmarine sediments. Twenty-three cycles are recognized in interval C, 13 in interval D, and 16 in interval E.

In post-Pennsylvanian time, subsidence of the Forest City basin probably continued for some time. Sheet deposits of Pennsylvanian sediments were stripped from the lower Ozark hills, exposing solution-cavity fillings of Pennsylvanian age. Jurassic gypsum was deposited near Fort Dodge, Iowa. Northwestern Iowa and probably other parts of the eastern midcontinent were buried with Cretaceous sediments, some of which were later stripped away. The area north of the Missouri River was buried under glacial deposits during the Pleistocene.

**REGION DEFINED**

The region discussed in this chapter encompasses the Pennsylvanian outcrops of Missouri and Iowa, excepting outliers in St. Louis County, Mo., and Scott and Muscatine Counties, Iowa, which are detached remnants of the Illinois basin and are described in chapter E. The two principal structural features of the Missouri and Iowa region are the Ozark uplift, which occupies nearly half of Missouri, and the eastern part of the Forest City basin in northwest Missouri and southwest Iowa (fig. 15). The Forest City basin extends westward a considerable distance into Kansas and into the southeast corner of Nebraska. This basin is flanked on the west by the Nemaha anticline of eastern Nebraska, Kansas, and Oklahoma. The Mississippi River arch extends north from the Ozark uplift, and in much of northeast Missouri, it separates the Pennsylvanian rocks of the eastern midcontinent region from those of the Illinois basin.

The Ozark uplands occupy a large area in central and southern Missouri and adjoining Arkansas. Pennsylvanian rocks are limited almost wholly to sink and cave fillings in the three tiers of counties in Missouri south of the Missouri River (Bretz, 1950, p. 796). The southernmost mapped areas of sink fillings are in Dent (loc. 560), Iron (loc. 533), and Shannon (loc. 589) Counties (pl. 1). The number of filled sinks known in the region is far larger than the number of locations shown on the map. Bretz's map shows as many as 97 known sinks in a single township in Gassonade County. Farther west, in Dallas (loc. 561) and Webster (loc. 569) Counties, small Pennsylvanian outliers are remnants of basal Pennsylvanian channel sandstones rather than sink fillings.

**PALEOGEOLoGY**

**UNITS UNDERLYING PENNSYLVANIAN**

Much of the Missouri-Iowa region (fig. 15) is underlain by Upper Mississippian carbonate strata of Meramec age, but the paleogeologic map (pl. 2) shows flanking areas of Osage strata adjoining the Ozark uplift. Older strata are also exposed in central and eastern Iowa northeast of the Forest City basin. These include some of Kinderhook age (Early Mississippian) and some of Devonian and Silurian age. Nearly all the Devonian and Silurian rocks are preserved beneath Pennsylvanian outliers, many of which are sinkhole or cave fillings.

![Figure 15.—Tectonic elements in the Missouri and Iowa region in Pennsylvanian time.](image-url)
LOWER BOUNDARY OF THE PENNSYLVANIAN

An Ozark upland had existed during much of the time following the Early Ordovician, although major transgressions of several ages had lapped onto one side or another. Rocks of post-Early Ordovician age form narrow bands on the flanks of that uplift. Numerous outliers of Pennsylvanian fluvial sandstones occupy channels incised in Mississippian limestone at the contact on the Springfield Plateau (fig. 16), west of the principal Ozark uplift.

Cambrian and Ordovician rocks of the uplands are largely dolomites that were exposed above sea level so long that complex cave and sinkhole systems developed. During the Early Pennsylvanian, these cavities became largely filled with a diversity of Pennsylvanian sediments. As many as 97 filled sinks in a single township and other sinks distributed across at least 16 counties have been recorded (Bretz, 1950, p. 796). In other parts of the Ozark uplands there doubtless are many that have not been discovered. The pre-Pennsylvanian surface in the Ozark area probably affords the finest examples of paleokarst topography in the United States.

PALEOTECTONIC IMPLICATIONS

At some time after the Late Mississippian retreat of the sea, the floor of the Forest City area was downwarped several hundred feet to form a partly enclosed basin. This sinking may have begun as early as Late Mississippian (Chester) time, but more sinking probably occurred either late in the time of Pennsylvanian subinterval A2 (late Morrow) or early in interval B time. Mississippian and other carbonate rocks that did not at once become covered with sediment were subject to both solution and stream erosion, and a pavement was formed of insoluble residue including local chert gravels.

West of the Forest City basin, in Kansas and Nebraska, the eastern escarpment of the Nemaha anticline persisted from Mississippian Pennsylvanian time. The Ozark area may also have persisted as a landmass. A hill of pre-Pennsylvanian rocks in Dallas County, Iowa, was not buried until the time of interval D. Farther northwest, Precambrian sedimentary rocks have been encountered directly below the Pennsylvanian in one well, whereas in surrounding wells the Mississippian St. Louis Limestone forms the substratum. This relationship may represent a very local pre-Pennsylvanian domal uplift or an erosional monadnock of the Precambrian which was not buried until Early Pennsylvanian.

INTERVAL A

Interval A is probably absent or, if present, is extremely restricted in extent in the Missouri-Iowa region. The Forest City basin of southwestern Iowa, northwestern Missouri, and adjacent parts of Nebraska and Kansas contains in its deeper areas several hundred feet of Pennsylvanian strata older than those that crop out to the south along the margins of the Ozark uplift. These strata have been called Cherokee Group and have been considered Des Moines in age, but recently most of them were assigned an Atoka age (interval B) (Searight and Howe, 1961, p. 80). No search for spores from the coal beds in this sequence had been made in 1969, and the sequence was not dated, but possibly a part of it belongs to interval A. Similar strata may crop out in southeastern Iowa, in Henry and Van Buren Counties, or northwestern along the northeast border of the Forest City basin. No recent stratigraphic studies have been made of these Iowa strata, so none of them is here referred to interval A, except in the small area near Muscatine, Iowa, which is considered part of the Illinois basin region.

INTERVAL B

FORMATIONS INCLUDED

Interval B of the Missouri-Iowa region is mainly restricted to western Missouri and southwestern Iowa. It includes unexposed strata in the deeper part of the Forest City basin of Missouri and limestone in cave fillings in southwestern Missouri; deposits in both areas were assigned to the Burgner Formation of Atoka age by Searight and Howe (1961). Other basal Pennsylvanian strata assigned to interval B in Missouri include the Riverton Formation, the McLouth Formation, and the Cheltenham Clay (pl. 13), as well as deposits in caves and sinkholes in the Missouri Ozarks south of the Missouri River. The name Graydon Channel Sandstone has been applied to channel sandstones of interval B at the base of the Pennsylvanian in southwestern Missouri.

In Iowa no formal names, other than Des Moines Series and Cherokee Shale or Group, have been applied to Pennsylvanian strata that are here referred to interval B, or Atoka Series.

Parts of eastern Missouri presumably have very little or no sediment representing interval B. These areas are on the crest and western slope of the Mississippi River arch, where initial Pennsylvanian sediments are probably of Des Moines age.

STRATIGRAPHIC RELATIONS

The name “Des Moines beds” was originally applied to all Pennsylvanian strata below the “Missouri formation” in outcrops extending for many miles along the Des Moines River valley, Iowa (Keyes, 1893). In the present discussion marine strata yielding fusulinid genera Fusulinella or Protofusulinella are referred to interval B. Compression floras belonging to floral zones 7 and 8 (Read and Mamy, 1964) are also used as criteria for identifying interval B. In Arkansas and Oklahoma, strata yielding these diagnostic fossils are placed in the Atoka
Figure 16.—Geographic features and localities in the Missouri and Iowa region mentioned in text.
Rocks of this age were not known in Missouri and Iowa with certainty until the discovery of a primitive form of *Fusulinella* in a Pennsylvanian limestone in a sinkhole in Jasper County, southwestern Missouri (fig. 16) (Searight and Palmer, 1957; Thompson, 1953). This indicated that rocks belonging to interval B (Atoka) are present in Missouri in a sequence which previously had been referred to the Des Moines.

After a thorough review of the earliest Pennsylvanian rocks in various parts of Missouri, the following units were removed from the Des Moines Series and were referred to the Atoka (Searight and Howe, 1961, p. 80):

- **Burgher Formation**, named for the cave filling in Jasper County; **Cheltenham Clay** of eastern Missouri;
- widespread clay filling caves and sinkholes in the Ozark area of Missouri; **Riverton Formation** of southwestern Missouri; and a thick succession of dark mudstones with minor amounts of sandstone, limestone, and coal from the deeper part of the Forest City basin, to which the name **McLouth Formation** had been applied. The geological surveys of Iowa, Kansas, and Nebraska did not consider the extension of rocks of Atoka age into their States, although in the deeper parts of the Forest City basin, rocks referred to the **McLouth Formation** (Atoka) in Missouri can be traced into adjoining parts of Iowa, Kansas, and Nebraska.

The author of this chapter has projected the Des Moines-Atoka boundary in deep well records in the three States by using the widespread highly permeable sandstone known as the **Warner Formation** of the **Krebs Subgroup** as the base of interval C; all underlying Pennsylvanian strata—units named in the preceding paragraph and local rocks that may be part of the **Hartshorne Formation** (Des Moines)—have been referred to interval B. Where the Warner is absent, the preponderance of very dark gray to nearly black mudstone in the lowest Pennsylvanian has been used to define interval B. Unfortunately, no cores penetrating the **McLouth Formation** and correlatives are available for paleontological study in any of the four States.

The Pennsylvanian rocks of the Ozark area have been correlated principally on the basis of marine faunas in sparse limestones and chert beds preserved in sink fillings. A fauna in the roof of a coal bed that was formerly mined at locality 480, Miller County (Hoare, 1961, p. 2), has been referred to the **Burgher Formation** of Atoka age. At a clay pit in Maries County (loc. 475) the **Blackjack Creek Formation** (basal Marmaton Group, middle Des Moines) overlies the **Cheltenham Clay** (McQueen, 1943, p. 179). Although coal beds are present in the sink fillings at some places, diagnostic evidence of their age as determined from small spores has not been published.

In addition to occurrences of interval B shown on plate 64 in southeastern Iowa, Pennsylvanian outcrops in Van Buren and Henry Counties, north of the Mississippi River arch, may belong to interval B, but no sequence of these outcropping strata has been described, and no study of fossil plant or animal remains has been made.

**UPPER BOUNDARY OF INTERVAL B**

The lower part of the Des Moines Series in Oklahoma has been named the **Krebs Group** (pl. 13). This term has been extended into Missouri as a subgroup within the **Cherokee Group** (Searight and Howe, 1961). The basal unit of the Krebs is the **Warner Formation** in most parts of southwestern Missouri, but the Missouri Geological Survey recognizes a thin discontinuous unit as the **Hartshorne(f) Formation** below the Warner (Searight and Howe, 1961, p. 81-83). An effort has been made to use the base of the Warner as the top of interval B, and to project this boundary, using lithologic criteria, into southwestern Iowa, northeastern Kansas, and a small area in extreme southeastern Nebraska. Strata of interval B end abruptly west of Missouri and Iowa at the eastward-facing escarpment of the Nemaha anticline.

The upper boundary of interval B in the eastern mid-continent region probably is at a lower stratigraphic horizon than the B-C boundary to the east in the Illinois basin and in Ohio. In the latter areas the upper limit of B is arbitrarily placed at the top of the **Seville Limestone Member** of the Spoon Formation or of correlative limestones, all containing **Fusulinella iowensis** of early Des Moines age. The Seville probably is younger than the oldest rocks of the Des Moines Series and, on the basis of brachiopod faunas, may correlate with a limestone in the upper part of the **Krebs Subgroup** of Missouri to which the name **Seville** has been extended (Searight and Howe, 1961, p. 84); the Seville in Missouri, however, has not yielded **Fusulinella iowensis**. A horizon equivalent to the base of the Krebs has not been recognized in the Illinois basin.

The **Cheltenham Clay**, whose type locality is in St. Louis, Mo., extends as a stratigraphic unit through large parts of Missouri east of the Forest City basin and is present in one sink and cave filling in the Ozarks in Missouri. The application of the name *Cheltenham* to clays beyond the type locality was supported by the finding of clay similar in mineralogy and in ceramic properties in the Ozark area beneath limestone or chert containing fossils of Atoka age. At most localities there is no paleontologic evidence for the age of the clay referred to the Cheltenham. Because this clay is the oldest Pennsylvanian deposit over a wide area, including the Mississippi River arch as well as the Ozarks, and because the typical *Cheltenham Clay* is a sheetlike deposit of fairly wide extent, mineralogic resemblance...
between the type *Cheltenham* and the clay in the cave fillings may be doubtful proof of equivalence in age. Locally, strata overlying these clays range from chert or limestone of probable Atoka age in some of the cave fillings to the *Tawak Member of the Scammon Formation* of the *Cabaniss Subgroup*, the *Crowebury Formation*, and the *Blackjack Creek Formation* of the lower and middle parts of interval C (Des Moines age). Interval B time is unquestionably represented in the *Cheltenham Clay*. Thus, a thickness of 10-20 feet of strata across much of the Mississippi River arch in eastern Missouri and western Illinois that was assigned to interval C on the accompanying isopach map (pl. 7A) should be included in interval B instead (pl. 6A). There is reasonable doubt, however, that the entire *Cheltenham* is of interval B age; such an interpretation would require either an erosional unconformity or a sizable hiatus at some places between the deposits of clay and the deposits of superjacent strata.

**THICKNESS TRENDS**

In northeastern Kansas interval B reaches a maximum thickness for the Forest City basin of about 460 feet. In northwestern Missouri, southwestern Iowa, and extreme southeastern Nebraska, however, the thickness is less than 400 feet. Strata of this interval wedge out in the subsurface toward the north, east, and south from the areas of maximum thickness. The basin in which they occur deepens westward nearly to the Nemaha anticycle, where interval B is abruptly truncated.

In central Missouri the *Cheltenham Clay* has been included in interval B except in those areas in which superjacent strata belong well up in the Des Moines Series and are placed in interval C. Thicknesses of those strata referred to interval B in the *Cheltenham Clay* areas rarely if ever exceed 30 feet in thickness.

Because most Pennsylvanian strata in the Ozark area are fillings of isolated caves and sinkholes, the thicknesses vary greatly and prohibit isopach mapping. In some places as much as 50 feet of Jefferson City Dolomite of Early Ordovician age is underlain by 100 feet or more of Pennsylvanian cave fillings.

**LITHOFACIES TRENDS**

Interval B is mostly soft dark-gray to black mudstone where the interval is thickest in Iowa and in the deep part of the Forest City basin and is mostly kaolinitic clay where the interval is thin in parts of east-central Missouri. In southwestern Missouri the mudstones are predominantly gray and rather sandy, but some are dark gray.

Sandstone is present in subordinate amounts in Missouri and Iowa. At several places in Missouri the Pennsylvanian rocks include a basal conglomerate that is composed largely of chert pebbles derived from rocks of Mississippian or of Cambrian and Ordovician age. Locally this conglomerate underlies the *Cheltenham Clay* and probably belongs to interval B.

The lithofacies map (pl. 6B) shows principally a difference between areas where interval B contains more than 80 percent mudstone and those where it contains 50-80 percent mudstone. A few areas with a higher percentage of sandstone may mark the courses of stream channels. In southwestern Missouri, in the area of the *Riverton Formation*, some wells show more than 20 percent of sandstone.

Coal beds referred to the *Riverton Formation* are present and have been mined locally in southwestern Missouri. A few coals are mentioned in records of deep borings in the Forest City basin. They may be equivalent to Riverton coals of the outcrop area.

The limestone of the *Burkner Formation* is known to be fairly widespread in southwestern Missouri (Searight and Howe, 1961, p. 80-84). Limestone has been reported at comparable stratigraphic position in some wells in the Forest City basin and has been referred to the *Burkner*. It is also present in sink fillings, where a maximum thickness of 16 feet has been reported.

The sink and cave fillings of the Ozark area include a wide range of lithologic types. Refractory fire clay, including diaspore, and other clays of various colors are most commonly reported and mined. A sandstone lining the base and walls of caves and sinks is also common. Some sink deposits contain coal beds as much as 30-50 feet thick. Locally these coal beds have limestone or cherty limestone caprocks containing marine faunas. Deposits of hematite (Crane, 1912) and pyrite and other sulfide ores (Grawe, 1945) also occur. The diversity of deposits known as fillings of solution cavities in the Ozarks was described by Bretz (1950).

One notable characteristic of many of these deposits is the extreme deformation of stratified sediments filling the cavities, even where the carbonate rocks in which the fillings occur are largely undeformed. The sandstone rimrock commonly dips steeply around the margins of a sink or cave fill. Bedded mudstone may be nearly vertical, and at one locality coal nearly 180 feet thick occurs in a cylindrical opening less than 180 feet in diameter. In many places coal beds and other strata show slickensides. An iron ore deposit near Rolla contains about 40 feet of hematite overlying 120 feet of pyrite, which in turn overlies 80 feet of chert conglomerate in a filled sink 250 feet wide (Bretz, 1950, p. 815). Other cavity fillings include chalcopyrite, galena, and sphalerite.

The pocket coal deposits are largely in the northwestern part of the Ozark area in Moniteau, Morgan, and Miller Counties. The refractory clays are largely in the northeastern part of the area in Franklin, Gasconade, Osage, Crawford, and Maries Counties, and the ore deposits are principally farther south in Phelps, Dent, and southern Crawford Counties (Bretz, 1950, p. 796).
SOURCES AND ENVIRONMENTS OF DEPOSITION

SOURCES

There has been little study of the form of sandstone bodies in interval B in Missouri or Iowa. In general, sandstone probably was derived from northern or northeastern sources.

The sandstones in the Ozark area were probably redeposited from Cambrian and Ordovician sandstones from sources within the Ozark uplands. The channel deposits of the southern Ozarks may have had a similar source. Chert conglomerates were probably derived from the insoluble residues of cherty dolomites in the same region, accumulated as a regolith which was later washed into nearby sinks. The clays probably represent an accumulation of the fine fraction of the insoluble residue of several hundred feet of dolomite and limestone and were further altered by leaching after accumulation (McQueen, 1943, p. 209).

ENVIRONMENTS OF DEPOSITION

Cyclic sedimentation is well displayed in the section of interval B in southwestern Missouri. Sandstone, underclay, coal, and mudstone compose each cycle. A marine incursion is indicated by the Burgner Formation. The sea floor appears to have been well ventilated, as is indicated by the generally light gray color of the mudstones.

A substantial thickness of dark-gray to black mudstones in the McLouth Formation of the Forest City basin suggests that the basin was poorly ventilated. A greater thickness of section there than in all surrounding areas suggests differential downwarping of the basin and its partial isolation from the open sea by a barrier, possibly the Bourbon arch of southwestern Missouri and southeastern Kansas (fig. 15). Dark mudstones in the lower part of the interval were reported to contain linguloid brachiopods (McQueen and Greene, 1938, p. 174), as well as plant fragments, and may have been deposited in a brackish-water environment close to shore. The Bourbon arch probably isolated the Forest City basin during most of McLouth time, but toward the end of that time, accumulating sediments buried the barrier ridge, and as a result, normal circulation in the basin was restored. Precipitous cliffs of older Paleozoic rock and Precambrian granite fronted the western and deepest part of the basin.

The Cheltenham Clay, probably derived from a deeply weathered soil on carbonate rocks of the Ozark area, accumulated slowly and was exposed to subaerial weathering for long intervals of time. At several places in Missouri and western Illinois, coal and carbonaceous films can be traced through the clay in clay pits and natural exposures. Because the clay locally differs in color or structure on each side of these coal streaks, the clay may be a composite of sediments that accumulated while much thicker sediments were being deposited in Iowa, western Missouri, and western Illinois, both in interval B time and during the early part of interval C time. When coal was accumulating in surrounding basins, climatic conditions likewise favored growth of luxuriant vegetation on or near the Mississippi River arch. Because the arch stood slightly higher than the surrounding basins, the vegetation decayed, leaving in the Cheltenham Clay only carbonaceous films.

Coal beds and marine limestones that make up the cave and sink fillings represent deposits of marshes or shallow seas that were subsequently lowered into the sinks. The sandstones may have been washed into open cavities and subsequently deformed as further solution enlarged the cavities they had already filled. The refractory clays, including diaspore clays, were derived from the fine insoluble residues of a surface regolith and formed lateritic soils in the probably tropical climate of interval B time. After sediment washed into solution cavities, the silica in the clays may have been leached by waters charged with carbon dioxide and alkaline carbonates (McQueen, 1943, p. 203-215).

The sinkholes south of the Missouri River are larger and deeper than those in the Cheltenham Clay region of east-central Missouri and indicate that the Ozark land area stood somewhat higher than the adjoining Mississippi River arch. Solution enlargement of the caves and sinks after their filling with sediment occurred in the Ozark area but generally not in the Cheltenham Clay area.

Probably substantial thicknesses of Pennsylvanian strata have been removed from parts of the Ozark area. Where present, coals and carbonaceous mudstones probably yielded much sulfur to ground waters passing through them. The sulfuric acid thus formed would have leached the clays and reacted with country rock to form pyrite, marcasite, and other sulfides (Grawe, 1945; Bretz, 1950). Subsequently, pyrite in upper parts of the filled cavities would have oxidized to form hematite.

PALEOTECTONIC IMPLICATIONS

The Missouri-Iowa region was virtually stable and subsided only slightly during interval B time, as shown by the rather thin accumulation of sediment and by the predominance of fine detritus. In southwestern Iowa, northwestern Missouri, northeastern Kansas, and extreme southeastern Nebraska, differential downwarping of the Forest City basin permitted the accumulation of more than 400 feet of sediment in the deep parts. The basin was bounded on the west by the Nemaha anticline of Nebraska and Kansas, on the east and southeast by the Mississippi River arch, and on the south by the Ozarks. Along the southwest border, the Bourbon arch at least temporarily was weakly positive and formed a barrier.

The principal downwarping of the Forest City basin may have occurred before Pennsylvanian time or just
after the beginning of interval B time. During the earlier part of interval B time this area had some of the characteristics of a barred basin (Woolnough, 1937).

The sink and cave deposits of the Ozark area have been strongly deformed since deposition, and as a result, steeply dipping or vertical strata are common among materials filling cavities. The sediments cannot have accumulated with their present attitudes in solution depressions but must have subsided into cavities opened by solution after the filling of surface sinks and caves (Bretz, 1950, p. 830-832). This phreatic solution is carried on by moving ground water below the water table (Bretz, 1942).

The sinks containing fire clay and pocket coal deposits lie north of those containing pyrite and hematite (Bretz, 1950, p. 831), which are situated higher on the Ozark dome. Thus, the zonings of sinks containing iron ore, diaspore clay and pocket coal, and Cheltenham Clay (kaolinite) is a result of relative differences in height of the Ozark dome during the development, filling, and enlargement of the solution cavities.

**INTERVAL C**

**FORMATIONS INCLUDED**

The Des Moines Series was named Des Moines Group by Keyes (1883) for the Des Moines River valley, Iowa, and originally included all Pennsylvanian strata of the region between the base of the Missouri Formation and the top of the Mississippian. Subsequently the Des Moines Series was divided and the older strata in the southern midcontinent region were placed in the Morrow and Atoka Series. In the present publication, strata of the Atoka Series, and possibly the Morrow, are recognized in the Missouri-Iowa region and mapped as interval B. Interval C, therefore, includes most of the Des Moines Series; the unit listed below as the Hartshorne(!) Formation is mapped in interval B. The Missouri Geological Survey has adopted the following classification of units of formational or larger rank in the Des Moines Series (Searight and Howe, 1961, p. 85, 91); also given are names of some units of member rank, including all of those cited in this text.

**Rocks of the Des Moines Series—Continued**

**Marmaton Group**—Continued
Fort Scott Subgroup—Continued
Little Osage Formation
Flint Hill Member
Blackjack Creek Formation
Cherokee Group
Cabaniss Subgroup
Eccello Formation
Mulky Formation
Lagonda Formation
Sandstone member
Berrier Formation
Verdigris Formation
Crowebug Formation
Fleming Formation
Robinson Branch Formation
Mineral Formation
Scammon Formation
Chelsea Member
Taseah Member
Tebo Formation
Weir Formation
Krebs Subgroup
Bluejacket Formation
Seville Member
Drywood Formation
Rose Formation
Warner Formation
Hartshorne(!) Formation

The classification of the Des Moines Series used by the Iowa Geological Survey is like that of the Missouri Geological Survey.

**UPPER BOUNDARY OF INTERVAL C**

By present usage (Searight and Howe, 1961, p. 90) the base of the Hepler Member of the lower unnamed formation of the Pleasanton Group is the top of the Des Moines Series (interval C). The Hepler is reported to rest unconformably on upper limestones and mudstones of the Marmaton. Earlier the top of the Des Moines Series was placed at the base of the Warrensburg Member of the upper unnamed formation of the Pleasanton Group. This member is represented in several areas by sandstones that fill deep-cut channels, some of which extend to the base of the Pennsylvanian. In Missouri the Warrensburg is a more conspicuous sandstone than the Hepler. The Trivoli Sandstone Member of the Modesto Formation, the base of which is considered the top of interval C in the Illinois basin, probably corresponds to the Warrensburg rather than the Hepler Member.

**THICKNESS TRENDS**

In Missouri and Iowa the total thickness of interval C in areas where the upper part has not been eroded ranges from about 250 feet in north-central Missouri to about 675 feet in the deeper part of the Forest City basin. Thicknesses exceeding 500 feet are almost entirely limited to southwestern Iowa and westernmost
MISSOURI AND IOWA

Missouri in five areas, extending from Pottawattamie County, Iowa, south to Bates County, Mo.

On the crest of the Mississippi River arch, northwest of St. Louis, the total thickness of the interval would probably be less than 250 feet if complete sections of interval C were present. In this part of Missouri and on the eastern extension of the Mississippi River arch in western Illinois, the Krebs Subgroup and the formations in the Cabaniss Subgroup older than the Fleming Formation are generally absent, and thicknesses of the other formations are greatly reduced. The most nearly complete section of interval C on the crest of the Mississippi River arch is at locality 379 in St. Louis County, Mo., where the interval is more than 100 feet thick. In parts of the Mississippi River arch in Lincoln, Montgomery, Audrain, and Calloway Counties, interval C, if it were present as originally deposited, would probably be less than 100 feet thick. The lower part of interval C (the Krebs Subgroup and the lower part of the Cabaniss Subgroup) is generally absent from a belt east of a line extending southwestward from Lewis County to southern Howard County. Interval C thickens moderately into the Forest City basin.

LITHOFACIES TRENDS

In Missouri interval C consists of 16 sandstones, 23 gray mudstones (several containing marine invertebrates), 20 fossiliferous limestones, 22 coal beds, 26 underclays, and 6 black fissile mudstones. In a large part of the area gray mudstones, black shaly mudstones, and underclays make up more than 80 percent of the interval. In some other areas, particularly in linear belts adjacent to the northeastern border of the region, interval C consists of 50-80 percent sandstone.

In the western part of the region several well records show more than 20 percent nonclastics, mostly limestone; at most of these localities mudstones compose more than 80 percent of the detrital rocks present, but at some localities 50-80 percent of the detrital rocks are mudstones. Larger areas have 50-80 percent mudstones and 20 percent nonclastic components. In a few small areas lithofacies different from those mentioned above were mapped, but in these areas only a basal part of the interval is present because the remainder was removed by erosion.

Coal beds in Missouri and Iowa, though numerous, are generally thinner than their equivalent beds in the Illinois basin. Particular beds of limestone or coal are traceable with relative ease except for areas on the Mississippi River arch, where the lower five or six formations of nearby areas are absent. The 16 sandstones in interval C are generally thin sheet sands. In Iowa, limestone rarely makes up as much as 20 percent of the interval, and detrital sequences are somewhat thicker than in Missouri.

SOURCES AND ENVIRONMENTS OF DEPOSITION

In Missouri and Iowa, sandstones of interval C are thinner and finer grained than those in the Illinois basin, and lenticular channel-type sandstones are less common. The Warner Formation, at the base of interval C in Missouri, is very quartzose and probably was derived from older sandstones. Most of the higher sandstones contain some mica, feldspar, and interstitial clay and may have been derived directly from crystalline source rocks which probably were very distant. A few sandstones, such as the sandstone in the Lagonda Formation and the Flint Hill Member in the Little Osage Formation, are extensions of western Illinois units, whose sources lay to the east or northeast. The Engleveale Member of the Labette Formation seems to have been derived from the direction of the Canadian Shield to the northeast. The same direction is indicated for the source of the Walter Johnson Member of the Novata Formation.

The Bandera Quarry Member of the Bandera Formation, chiefly limited to western Missouri, seems to have an east-west trend and may have been derived either from the Nemaha anticline to the west or from the Ozarks to the east. Other sandstones, except the Chelsea Member of the Scoxmon Formation, are generally thin sheetlike bodies. The Chelsea is much thicker to the south or southwest and may have come from the direction of the Ouachita Mountains in Oklahoma or Arkansas.

The Nemaha antline extended nearly across Kansas from Nebraska to Oklahoma at the beginning of interval C time, but only a small part which formed an island in northeastern Kansas and southeastern Nebraska remained by the end of this time. It was probably a local source for channel sands in parts of northwestern Missouri.

The Ozarks are believed to have been a landmass of low relief and low elevation during much of interval C time. Because carbonate rocks formed the surface, the area was honeycombed with sinkholes and caves. Most of the cave and sink fillings are thought to be older than interval C, but in a sinkhole at locality 475 in Osage County, Mo., limestone of the Black Jack Creek Formation of the Fort Scott Subgroup overlies the Cheltenham Clay. Formerly, strata of interval C may have been widespread in Ozark sink deposits, but if so, their record has been largely lost through weathering and erosion. The Ozarks at times were probably high enough to shed fine clay from insoluble residues of the carbonate rocks.

ENVIRONMENTS OF DEPOSITION

For the upper part of interval C in Missouri and Iowa, from the Croweburg up through the Holdenville, 47 environmental maps of separate sedimentary units have
been prepared (Wright, 1965; Wanless and others, 1963; Gamble, 1967; Manos, 1963; Orloff, 1964). Shown on these maps are the major environments and the number of times each recurs:

- Prodelta
- Delta
- Prograding streams, channel sands
- Exposed delta, mudflats
- Coal swamps
- Restricted sea, black mud bottom
- Open sea, mud bottom
- Open sea, lime bottom
- Nondeposition

The number of repetitions of particular environments would be increased considerably if maps had been prepared for the lower part of interval C.

Interval C in Missouri and Iowa comprises about 20 cycles of sedimentation that include at least 23 separate marine transgressions from the south and west, of which 13 extended across Missouri to Illinois. The characteristics of this rock sequence indicate three principal differences between environments of deposition in Missouri and Iowa and those in Illinois. Environments in Missouri and Iowa had (1) less extensive deltas and alluvial fans, (2) more restricted coal swamps, and (3) seas in which a greater average thickness of fossiliferous marine limestone was deposited. The five units of black mud representing restricted seas are widespread not only in Missouri and Iowa but also in the Illinois basin to the east and in the Nebraska-Kansas-Oklahoma area to the west; each unit is generally less than 5 feet thick.

The Mississippi River arch may have been a lowland during the early part of interval C time. The refractory kaolinitic Cheltenham Clay underlies the earliest recognizable unit assigned to interval C and is referred to the Atoka by the Missouri Geological Survey. If the clay is of Atoka age, it may have been exposed after deposition and leached during the early part of interval C time. An alternative explanation is that it was exposed intermittently to weathering but that its accumulation continued into the early part of interval C time. This interpretation is supported by the presence in the clay of several carbonaceous films that can be matched to coal beds elsewhere and by the local occurrence above the Cheltenham of the Tiwahe Member of the Scammon Formation, a marine limestone well up in the Des Moines in some parts of Missouri and Iowa.

**PALEOTECTONIC IMPLICATIONS**

During interval C time the principal structural features of the Missouri-Iowa region were the Ozark uplift, the Mississippi River arch, and the Forest City basin. The Ozark uplift probably remained near sea level and underwent little structural change during this interval, but minor changes in elevation apparently took place in the other structural elements.

During the early part of interval C time the Mississippi River arch evidently served as a low barrier to marine transgression across east-central Missouri. The stratigraphic intervals between distinctive key beds in the upper part of interval C are abnormally small on and near the Mississippi River arch. This may be a result of slight upward movement during the later part of interval C (Marmaton) time, or it may be because detrital wedges derived from other sources generally pinched out before reaching the arch. The latter suggestion seems more likely, for the persistent units across the arch are the widespread marine limestones, which would not have formed if the area had stood topographically above surrounding basins.

The Forest City basin downwarped greatly during much of the Paleozoic, but its deep part appears to have been mostly filled before the deposition of interval C, which does not thicken markedly into the basin.

Because nonmarine detrital wedges of sandstone and mudstone are less numerous and cover smaller areas than in the Illinois basin, the seas were able to spread with fairly uniform depth over nearly all of Missouri and Iowa during the several marine transgressions. The frequency of these transgressions and the uniform thickness of the resulting deposits show clearly that the region was generally free from differential movements during interval C time.

**INTERVAL D FORMATIONS INCLUDED**

Interval D in Missouri and Iowa is equivalent to the Missouri Series, whose type locality is in this region. In Missouri the series is divided as in the following list (Searight and Howe, 1961, p. 97, 101, 107).

**Rocks of the Missouri Series (interval D)**

**Pedee Group**
- Iola Formation
- Weston Formation

**Lansing Group**
- Stanton Formation (5 named members not listed)
- Vila Formation
- Plattsburg Formation (3 named members not listed)

**Kansas City Group**
- Zarah Subgroup
  - Bonner Springs Formation
  - Wyandotte Formation (5 named members not listed)
- Lane Formation
- Linn Subgroup
  - Iola Formation (3 named members not listed)
  - Chanute Formation
  - Drum Formation (1 named member not listed)
  - Cherryvale Formation (5 named members not listed)
- Bronson Subgroup
  - Dennis Formation (3 named members not listed)
  - Galesburg Formation
  - Swope Formation (3 named members not listed)
- Lodore Formation
  - Hertha Formation (4 named members not listed)
In Missouri and Iowa, interval D is largely restricted to the eastern part of the Forest City basin; its area of occurrence extends into the western part of the basin in Kansas and Nebraska. Interval D has been eroded, however, from the alluvial valley of the Missouri River, which east from Kansas City divides the basin into northern and southern parts. An east-trending outlier in Chariton, Randolph, and Monroe Counties, Mo., is formed by the outcrop of the Moberly Member in the upper unnamed formation in the Pleasanton Group, which locally fills a channel cut through interval C down to pre-Pennsylvanian strata. A narrow north-trending band south of the Missouri River in Lafayette, Johnson, and Henry Counties is the area of outcrop of the Warrensburg Member of the upper unnamed formation in the Pleasanton Group, which is similar in age and character to the Moberly. A small outlier in St. Louis County, Mo., is a detached part of the Illinois basin and is described with it. Several other small outliers also occur in Missouri and Iowa.

STRATIGRAPHIC RELATIONS

The Missouri Series of the midcontinent region is characterized by the fusulinid genus Kansanella (Thompson, 1957; 1960, p. 111), which is not found below the Missouri but does occur in basal Virgil strata.

UPPER BOUNDARY OF INTERVAL D

A prominent unconformity below the Tonganoxie Sandstone Member of the Stranger Formation in the Douglas Group was recognized along the Missouri River near Leavenworth, Kans., and St. Joseph, Mo., and was chosen as the base of the Virgil Series by Moore (1952, p. 88), as it was believed to indicate tectonism in the southern midcontinent region. The strata between this sandstone and the Lansing Group below were designated the Pedee Group and were assigned to the top of the Missouri Series. A more recent regional study has shown that this erosional break is not widely traceable through the midcontinent region and is better interpreted as local channeling. The Kansas Geological Survey includes rocks of the Pedee Group of Missouri in the Douglas Group. They place the Missouri-Virgil boundary at the base of the Weston Shale Member of the Stranger Formation, which corresponds to the base of the Pedee Group of the Missouri Geological Survey. The Missouri Geological Survey has retained the Pedee as a group and continues to refer it to the Missouri Series, which is the assignment used here for Missouri and Iowa. No recent reclassification has been made in Iowa.

THICKNESS TRENDS

The entire uneroded thickness of interval D is present only in the deep part of the Forest City basin in an area extending from Platte County, Mo., north to Adair, Cass, and Pottawattamie Counties, Iowa (pl. 84). In this area measured thicknesses range from about 300 feet at the northern limit to possibly 600 feet or more in Buchanan County, Mo. Complete thicknesses are generally 300-500 feet.

The most obvious trend is a southward thickening of interval D from central western Iowa to northwestern Missouri.

LITHOFACIES TRENDS

In Missouri and Iowa, interval D consists of beds of sandstone, mudstone, limestone, and coal. A composite section for Missouri shows the following lithologic units: 8 sandstones; 35 mudstones, including shaly claystones, silty and sandy mudstones, black fissile mudstone, and underlains; 22 limestones; and 4 coal beds that are generally thin and local (Searight and Howe, 1961, p. 97, 101, 107). The four most prominent sandstone units are in the Pleasanton Group near the base of the interval. Interval D commonly contains 20-50 percent limestone and has a mudstone-to-sandstone ratio of more than 4:1. At some localities, especially in Iowa and in the northern tier of counties in Missouri, limestone makes up more than 50 percent of the interval. An eastward decrease in thickness of limestone is suggested in central Missouri, where limestone forms less than 20 percent of the succession at several localities. However, in this area most of the remaining part of the interval belongs to the Pleasanton Group, which contains only one very thin limestone. In the outlying areas of the Warrensburg Member of the upper unnamed formation of the Pleasanton Group, nearly the whole uneroded interval is sandstone, which in many places was deposited in channels that were cut into strata of interval C.

The limestones in interval D in Missouri and Iowa are equivalent to similar units in the Illinois basin, but individual limestones are thicker. Limestones of the Dennis and Swope Formations may each reach 30 feet or more in thickness, and the limestone in the Wyandotte Formation exceeds 50 feet in thickness at several places. The following formations are entirely or largely composed of limestone: Iatan, Stanton, Plattsburg, Wyandotte, Iola, Drum, Dennis, Swope, and Hertha.

Lenticular sandstones are less common in interval D in this region than in the Illinois or Appalachian basins,
and most of the interval except for the basal Pleasanton Group consists of widespread sheetlike limestones that are separated by equally widespread mudstones in such regular, layer-cake order that sample study logs and even driller's logs can be correlated with ease.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

The only sandstones in interval D that clearly show directions of flow of sediment are the Hepler Member at the base and the Moberly and Warrensburg Members higher in the Pleasanton Group. Channels of the Hepler extending southward or southwestward from Iowa into Missouri suggest that the source of sediment was in the Canadian Shield (Orlropp, 1964).

The Moberly Member forms an east-west outcrop in east-central Missouri, and directional studies show that this sandstone was deposited by a stream draining from east to west (Doty and Hubert, 1962). Although its course west of Chariton County, Mo., has been destroyed by later erosion, the Moberly stream is believed to have joined in Lafayette County with a southward-flowing stream from Iowa, and the combined river drained southward in the Warrensburg channel. This sandstone seems to have been derived from the northeast through the Moberly channel and from the north through a northward extension of the Warrensburg channel. Both may have come from different parts of the Canadian Shield, or the Moberly channel may have had a distant source in the New England or eastern Canadian highlands.

During interval D time the Forest City basin received detrital sediment from the Nemaha anticline, which had been reduced to small size and low elevation by progressive burial during the times of intervals B and C. An increasing volume of sediment seems to have flowed northward into the basin from the recently elevated Ouachita highlands to the south. Most of the sediments that are believed to have come from the Ouachitas and the Nemaha anticline, however, are in Kansas, and, if sediments from these sources were once deposited in Missouri, they have been almost entirely removed by subsequent erosion.

**ENVIRONMENTS OF DEPOSITION**

A series of 45 environmental maps of the Pleasanton, Kansas City, and Lansing Groups in the Missouri-Iowa region has been prepared (Orlropp, 1964; Glover, 1964; Horne, 1965; Kenny, 1968). These maps include all strata in interval D except the Pedee Group at the top, and they show environments in southeastern Nebraska, eastern Kansas, and northeastern Oklahoma for the same units.

The map representing the time of the Hepler Member, the basal unit of interval D, shows deltaic sandstones from Iowa extending into Missouri; similar sandstones derived from the Ouachitas to the south extend into southern Kansas near the Missouri line. The map of the Warrensburg and Moberly Members suggests that most of Missouri and Iowa was emergent and that the region was dissected by valleys 100-200 feet deep, in a few of which are remnants of these sandstones. Two maps show coal swamps in parts of the region. Ten maps show widespread thin black fissile mudstones that formed in shallow toxic seas in which there was no bottom life. The black mudstones contain inarticulate brachiopods, conodonts, and fish remains, but all fossils represent floating or swimming forms that sank to the bottom after death. Most of these mudstones are only a foot or so thick, and they lie between marine limestones or coal beds below and thicker marine limestones above.

Marine fossiliferous limestones occur at 21 stratigraphic positions. Seven of these are widespread thin limestones generally 1-2 feet thick that are directly overlain by black fissile mudstones. Because some of the black mudstones extend eastward and become the roof shales of coal beds, the seven thin limestones are believed to have formed in shallow seas offshore from coal swamps, and the extensive vegetative cover on land prevented inwash of much mud or sand. For some reason conditions in these limestone-depositing seas changed to toxic conditions that were responsible for black mudstone deposition at about the same time that the coal swamps to the east were inundated.

Presumably as the transgression continued and the depth of sea water increased, circulation improved, and the bottom became habitable for a rich and diverse fauna of brachiopods, mollusks, corals, fusulinids, crinoids, bryozoans, and other forms. As the great transgressions spread from the midcontinent region eastward to the Illinois basin and occasionally to the Appalachian region, the Missouri-Iowa region was inundated earlier and drained later than areas farther east. Therefore, the records of the transgressions in this region are much fuller than those farther east. In addition, the record of bottom life is very well preserved.

Overlying the thicker limestones in sharp contact are generally gray slightly calcareous mudstones. These have many of the characteristics of prodelta mudstones of intervals C and D in the Illinois basin, but they differ in that they maintain fairly uniform thicknesses of generally less than 20 feet over large areas and do not thicken markedly toward the sources of the sediment. Many of these mudstones also contain some marine microfossils. These sediments possibly were introduced into the seas as prodelta mud but were eventually redistributed by marine currents to more uniform thickness. The Ladora, Galesburg, Cherryvale, Chanute, Lane, Wyandotte (Island Creek Shale Member), Bonner Springs, Vitis, and Weston Formations are largely or wholly such mudstones. The Pleasanton Group includes
between the Warrenton and Hepler Members a thick mudstone formation, as yet unnamed, which is probably a prodelta deposit.

Marine waters may have covered the Missouri-Iowa region continuously during interval D time, except when exposure permitted the cutting of the Warrenton and Moberly channels and briefly when local coal beds formed, generally just a few feet above sea level.

**PALEOTECTONIC IMPLICATIONS**

The thickening of interval D toward the deep part of the Forest City basin suggests that there may have been some downwarping during the time of this interval.

The only significant sandstones appear to have been derived from far-distant sources to the north and northeast. This interval in Missouri and Iowa has so little sandstone that the absence of active tectonism is suggested, not only in the basin but also in surrounding uplands to a considerable distance. The nearest upland source at this time was the Ouachita uplift in Arkansas and Oklahoma, but most of the sediment from this source that was deposited during interval D time is now preserved farther west in Kansas.

The remarkable layered alternation of thin limestones, thin black fissile mudstones, thicker limestones, and thicker gray mudstones is repeated in 13 distinct cycles of sedimentation. The thick limestones seem to mark maximum transgressions, and the thick mudstones seem to have formed at times of active delta building and marine withdrawal from surrounding basins. Some of the mudstones may have formed in brackish or fresh water and may imply brief episodes of marine withdrawal from the areas.

As in other regions and intervals, the cycles may have resulted from periodic tectonism anywhere in the ocean basins of the earth or from eustatic shifts in sea level related to alternate glacial and interglacial episodes in Gondwana areas. Climatic variations in source areas also may have influenced the nature of sediments deposited.

### INTERVAL E FORMATIONS INCLUDED

Interval E is defined as equivalent to the Virgil Series, whose type area is in the northern midcontinent region. In Missouri, strata of interval E are classified as follows (Searight and Howe, 1961, p. 108-122).

**Rocks of the Virgil Series (interval E)**

**Wabaunsee Group**
- Stotler Formation (2 named members not listed)
- Pillsbury Formation
- Zeandale Formation (4 named members not listed)
- Willard Formation
- Emporia Formation (3 named members not listed)
- Auburn Formation
- Bern Formation (3 named members not listed)

**Shawnee Group**
- Topeka Formation (9 named members not listed)
- Calhoun Formation
- Deer Creek Formation (5 named members not listed)
- Tecumseh Formation
- Lecompton Formation (7 named members not listed)
- Kanawa Formation (3 named members not listed)
- Oread Formation (7 named members not listed)

**Douglas Group**
- Lawrence Formation (1 named and 2 unnamed members not listed)
- Stranger Formation (5 named members not listed)

This sequence of 3 groups, 19 formations, 61 named members, and several unnamed members reflects the complexity of the succession of strata composing interval E and indicates a far more complete knowledge of these strata in the midcontinent region than in the Illinois basin. It also indicates that the many individual units composing this interval in Missouri and Iowa have much greater lateral continuity than do units of comparable age in the Illinois basin, where lateral change in facies is more common.

The area from which interval E has not been eroded lies in the deep part of the Forest City basin in northwestern Missouri and southwestern Iowa. Interval E also occurs in a larger adjoining area in southeastern Nebraska and northeastern Kansas. In the latter area the entire interval to its upper limit at the Permian boundary is present, but in Missouri and Iowa the younger strata of interval E have been removed by erosion.

**UPPER BOUNDARY OF INTERVAL E**

At a few localities in Atchison County, in the northwestern corner of Missouri, a channel sandstone identified as the Indian Cave Sandstone, the basal unit of Wolfcamp age in the Permian System, has cut down into the Stotler Formation, the highest formation of interval E. Elsewhere the interval is unconformably overlain by Pleistocene glacial drift or alluvium of the Missouri River and its tributaries.

**THICKNESS TRENDS**

Interval E has a maximum recorded thickness in the Missouri-Iowa region of 665 feet. This maximum is in western Atchison County, Mo. (pl. 9A). However, farther west, across the Missouri River in Nebraska and also in a generalized section of Pennsylvanian rocks in Kansas (Moore, 1949, p. 178), about 60 feet of additional strata is referred to the Virgil Series. These strata overlie the Stotler Formation in the Wabaunsee Group and should be included in any restored thickness of in-
terval E for the Missouri-Iowa region. Thicknesses of less than 600 feet elsewhere in this area result in part from the varying degrees of erosion to which the strata have been subjected.

The very irregular outer border of the area of interval E results from removal of strata where the interval is thin, as in the intervalley areas. The determination of trends in thickness variation for interval E would require the preparation of isopachs of smaller units such as the Douglas or Shawnee Group for areas where these groups have not been eroded. This has not been done. The late L. A. Thomas (oral commun., 1961) indicated that the Shawnee Group thins northward in Iowa.

LITHOFA CIES TRENDS

Most of interval E in Missouri and Iowa is made up of 20-50 percent limestone and has a mudstone-sandstone ratio greater than 4:1 (pl. 9B). In eastern areas in Iowa and Missouri and in southern areas from central Andrew County, Mo., southwest, the limestone content is less than 20 percent and the mudstone-sandstone ratio is more than 4:1. At four localities in Iowa, limestone makes up more than 50 percent of the total. The Douglas Group, at the base of the interval, is largely detrital, and in areas where interval E consists mostly or entirely of the Douglas Group it contains less than 20 percent limestone. The largest proportion of limestone is in the Shawnee Group; the Wabaunsee Group has a somewhat smaller proportion.

The generalized section for Virgil strata in Missouri includes the following 106 units: 10 sandstones; 45 mudstones, including 6 underclays and 6 black fissile mudstones; 34 limestones; and 5 coals.

Of the 10 sandstones, only one, the Tonganoxie Sandstone Member, at the base of the Stranger Formation of the Douglas Group, has developed a marked channel facies and cuts down locally as far as 50 feet into underlying strata of interval D in outcrops along the east bluff of the Missouri River in Platte and Clay Counties, Mo. The aggregate thickness of the sandstones is small compared with the thickness of sandstones in other basins.

The mudstones and limestones alternate many times, but in the Shawnee Group they form four sequences containing four or five limestones and four interbedded mudstones, including one widely persistent black fissile mudstone in sharp contact with underlying marine limestone and grading upward into gray mudstone. These sequences record a complex rhythm which has been called a megacyclothem (Moore, 1949, p. 145-146). Most of the mudstones contain some marine invertebrates, but a few, such as the Lawrence Formation, include abundant fossil plants at some places and may be nonmarine. All limestones are marine and contain normal marine faunas.

The coal beds are thin, but two, the Nodaway and Elmo, have been mined in a few places. At some places the coal beds have underclays, but several are reported to rest directly on laminated mudstones.

SOURCES AND ENVIRONMENTS OF DEPOSITION SOURCES

Although the Tonganoxie Sandstone Member of the Stranger Formation and other sandstones of interval E in the Missouri-Iowa region are cross-stratified (Ball, 1964, p. 90-101), no published comprehensive study of directions of crossbedding is available for use in determining sources of sediments. Thickness trends provide some help. The total thickness of the Douglas Group, the most detrital component of interval E, increases southward and southeastward, and sandstones are more plentiful in the thicker section (Ball, 1964, p. 69). The Douglas Group ranges in thickness from 50 feet in southern Nebraska to 600 feet at the Kansas-Oklahoma line, but the remnant of the group in Missouri and Iowa is too small to display such conspicuous southward thickening.

On the basis of total thickness variation of the Douglas Group, the Arbuckle Mountains in southern Arkansas, the Ouachita Mountains in eastern Arkansas and Oklahoma, and the western flank of the Ozark uplift have been suggested as major sources of sediments of early Virgin age (Ball, 1964, p. 228-230). The Arbuckle orogeny occurred at about the end of Missouri (interval D) time, and the Arbuckle Mountains probably yielded much sediment to the northern midcontinent basins, principally to those west of Missouri and Iowa. The Ouachita area was more distant, but probably contributed some sediment. The Ozark area does not now contain the types of rocks that would be necessary to produce the micaceous sandstones in the lower part of interval E, but Ball (1964, p. 227) suggested that the source rocks may have been earlier Pennsylvanian sandstones that were deposited on the western flank of the Ozarks and that have now been entirely removed. The Shawnee Group also thickens southward and southeastward; the proportion of mudstone and sandstone increases southward, principally in Kansas and Oklahoma. Source areas during the time of deposition of the Shawnee Group, therefore, seem to have lain to the south or southeast, as they did during the time of deposition of the Douglas Group.

In a generalized section for Missouri the Wabaunsee Group contains about 30 feet of sandstone in a total thickness of 300 feet (Searight and Howe, 1961, p. 118, 121). No directional studies of these sandstones have been reported, and an isopach map of these strata would show little variation in the small area in Missouri and Iowa in which they remain. Probably a distant area to the south or southeast provided detritus for the
Wabansssee Group as it did for the earlier parts of interval E. The Nemaha anticline had been entirely covered with sediment during later interval D time and probably had little or no topographic expression during interval E time. Therefore it probably was not a source of sediment for interval E. The thin section in Iowa and Nebraska and the small proportion of detrital material indicate that no highlands lay anywhere near the northern midcontinent region during interval E time.

ENVIRONMENTS OF DEPOSITION

All the 34 limestone units in interval E in Missouri and Iowa contain marine faunas, and marine fossils are also found in many of the mudstones and, at a few localities, in parts of the sandstone units. Some of the coals may not have formed in situ, for the interbedding of coal beds and carbonaceous mudstones suggests that the plant detritus was reworked. The coal beds in this area do not have a typical underclay root zone, and at many places they lie between strata that contain marine fossils.

The environments of the Douglas Group at the base of the interval have been discussed at length by Ball (1964, p. 271-311). He suggested that during the time of deposition of the Douglas Group the southern shoreline of an inland sea shifted back and forth generally within the boundaries of the present State of Oklahoma, and that the Kansas, Missouri, and Iowa areas should be regarded as almost wholly marine. Nearness to shore in the southern part of the area is suggested by increasing abundance of Myalinid pelecypods, which are thought to be indicative of shallow or nearshore environments.

Ball (1964, p. 306-311) referred the sandstones of the Douglas Group to a mixed environment. He thought that the mechanism of their emplacement was similar to that described by Moore (1957) in the following quotation:

The clastics were spread mostly at near zero gradients throughout most of the area of sedimentation*** transportation being effected by generally slow back and forth shifting, induced by waves and currents*** in discontinuous fluctuating ponds, extremely shallow lakes, and lagoons of varying size, and semi-enclosed extensions of shallow seas, the plexus of water bodies ranging from fresh or slightly brackish to normally saline water.

Discontinuous sandstone bodies have been described as channel deposits, but Ball proposed that some may be offshore barriers. He did not believe that the shapes of the sandstone bodies are well enough known at present to assign them to a particular environment. An alternative explanation is that the sands and muds were introduced from the land south or southeast of the basin by streams, at the mouths of which detrital wedges of mud accumulated in prodeltas that were later cut by distributary channels as lengthening stream courses extended across the newly deposited muds. Further reworking at times of rising sea level may have formed frontal barrier islands to the deltas, as suggested by Ball. This explanation would not necessitate extensive redistribution of the sediment by currents after it had been introduced into the basin.

The megacyclothem of the Shawnee Group were mapped environmentally by Souter (1966) for the Oread and Lecompton Formations and by Dickson (1965) for the Deer Creek and Topeka Formations. The significance of megacyclothemic sequences has been discussed by Moore (1931, 1932, 1949, 1950), but no obvious mechanism to account for them has been discovered. The ecology of several of the limestone and mudstone units of the Shawnee megacyclothem also has been described by Moore (1964).

The Wabansssee Group commonly has a simple alternation of limestones with gray mudstones and locally with sandstones, but the Howard Formation, near the base of the group, has characteristics of an incomplete megacyclothem.

PALEOTECTONIC IMPLICATIONS

The channel sandstones at the base of the Douglas Group were formerly used as evidence of emergence of the eastern midcontinent region between deposition of the Missouri and Virgil Series (Moore, 1936a, p. 1802), and the emergence was believed to correspond to a major orogeny in the Arbuckle Mountains to the south. More recently, Ball (1964, p. 88) suggested that the relief of the unconformity at the base of the Tonganoxie Sandstone Member does not exceed 50 feet and that other disconformities in the Pennsylvanian succession are of similar magnitude. Ball concluded that complete emergence and subaerial erosion is not required to produce the observed relations. Nevertheless, the southward or southeastward thickening of interval E in Kansas, where the entire sequence is present, indicates a vast influx of sediment into the basin from that direction. This in turn probably indicates a highland produced by the Arbuckle orogeny and possible renewed uplift of the Ouachita folded belt and of the western flank of the Ozarks, as proposed by Ball. The disappearance of the last surface expression of the Nemaha anticline and the marine origin of nearly all the rocks of interval E in the northern midcontinent region indicate stability for the area in all directions except to the south.

The megacyclothemic rhythms of the Shawnee Group may have had a tectonic cause, but they can be explained equally well by a regular pattern of eustatic fluctuations of sea level.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

Pennsylvanian rocks form a westward-thickening wedge in parts of Missouri and Iowa. Within this region, the Mississippi River arch and the eastern slope of the
Forest City basin have a relatively thin section that includes only intervals B and C.

The deep part of the Forest City basin (Lee and others, 1946; McQueen and Greene, 1938) includes intervals B, C, D, and E. A substantial thickness of older strata of the Cherokee Group, probably representing interval B, is present in the deep part of the basin, but the strata wedge out in the subsurface beneath younger strata of intervals C, D, and E. In southwestern Iowa the Pennsylvanian rocks dip southward and westward at 10-15 feet per mile. The succession thickens at the rate of about 10 feet per mile westward across Missouri and Iowa to a thickness of about 1,700 feet in the deepest part of the basin in the northwest corner of Missouri. Pennsylvanian strata continue to thicken westward into eastern Kansas and Nebraska and reach a maximum of more than 2,400 feet in Jackson County, Kans. Because the complete Pennsylvanian column is not preserved anywhere in Missouri or Iowa, thicknesses shown on plate 11 are of erosional remnants.

The Ozark uplift in southeastern Missouri was a low landmass during most of Pennsylvanian time, and the strata exposed were primarily lower Paleozoic dolomites and limestones in which a complex of sink holes and caves was developed. As Pennsylvanian sediments encroached on the margins of this karst landscape, the cavities were filled with different kinds of sediments (Bretz, 1950). This area probably affords the best display of ancient cave and sink deposits in the United States. Thicknesses are given at a number of places (pl. 11), but no isopachs are drawn, because the bedrock of most of the area is Cambrian or Ordovician, and the Pennsylvanian sediments occur only as sink or cave fillings. Some well logs show more than 100 feet of Pennsylvanian strata in cave fillings beneath 50-150 feet of lower Paleozoic dolomites.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

Units overlying Pennsylvanian

Pennsylvanian sediments are beveled across an area from the middle part of the Forest City basin in northwestern Missouri and southwestern Iowa to the borders of the basin near the Mississippi River arch, along the northeastern borders of the basin in Iowa, and near the Ozark uplift to the southeast. Probably 1,000 feet or more of strata was removed from the upturned edges of the basin.

Anomalous gypsum deposits near Fort Dodge, Iowa (Keyes, 1895b; Wilder, 1903) overlie Lower Pennsylvanian and, locally, Middle Mississippian strata. Because they contain no visible fauna or flora, these overlying strata once were referred to the Permian or Upper Pennsylvanian, but the discovery of small spores of probable Late Jurassic age (Cross, 1966, p. 46) indicates that they are related to the Michigan Jurassic red beds, and may once have been continuous with them. The remnant *Fort Dodge Gypsum* may occupy a pre-Jurassic erosional depression in Lower Pennsylvanian and Middle Mississippian rocks.

The eastern border of the Upper Cretaceous in the Great Plains is in northwestern Iowa. At all localities the basal rock of this series has been mapped as the Dakota Sandstone. These Cretaceous rocks are mapped as virtually continuous across several northwestern Iowa counties, but in parts of these counties available well logs place the Pennsylvanian immediately below Pleistocene drift. Some early county reports of the Iowa Geological Survey (Lees, 1927) indicate that the Cretaceous has a spotty distribution. The area is drift covered and has few outcrops, and in some counties only two or three well records are available. The extent of the Cretaceous as shown on plate 12, therefore, is less than that shown on the Iowa State geological map, but the distribution on plate 12 seems more consistent with available information. The base of the Cretaceous bevels the Pennsylvanian; northward the Cretaceous rests on Lower Pennsylvanian and still farther north on Mississippian rocks.

Pleistocene glaciation covered the Iowa area of Pennsylvanian rocks and also the area of northern Missouri to the Missouri River. In central western Missouri the drift boundary is in the first tier of counties south of the river. Thus, in southwestern Missouri and in the Ozark uplands, Pennsylvanian rocks have no superjacent strata. Pennsylvanian rocks in the Ozarks, however, consist largely of sinkhole and cave fillings in Lower Ordovician carbonate rocks.

PALEOTECTONIC IMPLICATIONS

During and after the Pennsylvanian Period moderate downwarping occurred in the Forest City basin; therefore, younger Pennsylvanian strata have escaped erosion in the deepest part of the basin, which is near the northwest corner of Missouri. In general, about 1,000 feet of Pennsylvanian and, locally, Lower Permian strata has been destroyed as a result of broad regional upwarping. Comparable downwarping permitted Cretaceous sedimentary rocks to extend eastward to northwestern Iowa and probably as far as western Illinois, as indicated by an outlier there. The eastern margins of these strata were later eroded. Patchy distribution of the Cretaceous in Iowa suggests that the surface of these rocks had not yet been reduced to a level plain when Pleistocene glacial deposits buried the surface. Absence of any major tectonic activity has allowed weathering and slow erosion of Pennsylvanian rocks in southern Missouri to continue to the present.
REFERENCES


Lee, Wallace, and others, 1946, Structural development of the Forest City basin of Missouri, Kansas, Iowa, and Nebraska: U. S. Geol. Survey Oil and Gas Inv. Prelim. Map 48, 7 sheets.


1899, Geology of Scott County [Iowa]: Iowa Geol. Survey Ann. Rept. (1898), v. 9, p. 389-519.


Nebraska and Adjoining Parts of South Dakota and Wyoming

By GEORGE E. PRICHARD

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-G
CONTENTS

| Abstract ................................................................. | 115 |
| Region defined ......................................................... | 115 |
| Paleogeology ............................................................ | 115 |
| Units underlying Pennsylvanian ........................................ | 115 |
| Paleoecological implications .......................................... | 116 |
| Interval A ........................................................................ | 117 |
| Formations included ...................................................... | 117 |
| Upper boundary of interval A ............................................ | 117 |
| Thickness trends .......................................................... | 118 |
| Lithofacies trends ....................................................... | 118 |
| Sources and environments of deposition ................................ | 118 |
| Paleoecological implications ........................................... | 118 |
| Interval B ........................................................................ | 118 |
| Formations included ...................................................... | 118 |
| Upper boundary of interval B ............................................ | 118 |
| Thickness trends .......................................................... | 119 |
| Lithofacies trends ....................................................... | 119 |
| Sources and environments of deposition ................................ | 119 |
| Paleoecological implications ........................................... | 119 |
| Interval C ........................................................................ | 119 |
| Formations included ...................................................... | 119 |
| Upper boundary of interval C ............................................ | 120 |
| Thickness trends .......................................................... | 120 |
| Lithofacies trends ....................................................... | 120 |

Table C—Continued

| Sources and environments of deposition ................................ | 120 |
| Paleoecological implications ........................................... | 120 |

Interval D ........................................................................... | 121 |
| Formations included ...................................................... | 121 |
| Upper boundary of interval D ............................................ | 121 |
| Thickness trends .......................................................... | 121 |
| Lithofacies trends ....................................................... | 122 |
| Sources and environments of deposition ................................ | 122 |
| Paleoecological implications ........................................... | 122 |

Interval E ........................................................................... | 122 |
| Formations included ...................................................... | 122 |
| Upper boundary of interval E ............................................ | 122 |
| Thickness trends .......................................................... | 122 |
| Lithofacies trends ....................................................... | 122 |
| Sources and environments of deposition ................................ | 122 |
| Paleoecological implications ........................................... | 122 |

Total thickness of Pennsylvanian rocks ................................ | 124 |
| Thickness trends .......................................................... | 124 |
| Paleoecological implications ........................................... | 124 |
| Geologic units directly above Pennsylvanian system ................ | 125 |
| Units overlying Pennsylvanian .......................................... | 125 |
| Paleoecological implications ........................................... | 125 |

References ................................................................. | 125 |

ILLUSTRATION

[For listing of plates (in separate case) see volume "Contents"]

FIGURE 17. Map showing structural features of Pennsylvanian (solid lines) and Laramide (dotted lines) ages and geographic features in Nebraska and adjacent areas .................................................. 116

TABLES

| TABLE 16. Divisions of the Hartville Formation .................................................. | 117 |
| 17. Stratigraphic units of Nebraska assigned to interval D ................................ | 121 |
| 18. Stratigraphic units of southeastern Nebraska assigned to interval E ................ | 123 |
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES,  
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM  

NEBRASKA AND ADJOINING PARTS OF SOUTH DAKOTA AND WYOMING  

By George E. Prichard  

ABSTRACT  

Rocks of Pennsylvanian age crop out only on the flanks of the Black Hills (fig. 17) in South Dakota, along the Hartville uplift and in the Laramie Range of Wyoming, and in southeastern Nebraska. Although several thousand oil and gas test holes have been drilled in the region, only about 1,000 holes penetrate Pennsylvanian rocks. The test holes are irregularly distributed, and in large parts of the region they are widely scattered (pl. 1).  

PALEOGEOLOGY  

UNITS UNDERLYING PENNSYLVANIAN  

In much of the northern midcontinent region, Pennsylvania rocks are separated by an unconformity from underlying rocks which range in age from Precambrian to Mississippian. The Mississippian rocks have been studied in detail and subdivided by Carlson (1963a; unpub. data). Other pre-Pennsylvanian rocks have been differentiated and mapped by Reed, Svoboda, Prichard, and Fox (1958) and by Burchett and Carlson (1966, fig. 10). In the Nebraska part of the paleogeologic map (pl. 2), the classification by these authors is followed.  

On the east flank of the Laramie Range and in the Hartville uplift in southeastern Wyoming, Pennsylvania rocks rest unconformably on Mississippian limestone. Because of a lack of test holes of adequate depth near the southeast corner of Wyoming, the eastern limit of Mississippian rocks cannot be accurately located (pl. 2). Farther east toward the Cambridge arch in Nebraska (fig. 17), Pennsylvanian rocks overlie the Precambrian.  

On the south flank of the Black Hills in southwestern South Dakota and in northwestern Nebraska, strata of Pennsylvanian age rest on a widespread limestone of Mississippian age which probably extends continuously to the Hartville uplift in Wyoming (Carlson, 1963b). This unit is known as the Madison Limestone in northwestern Nebraska, the Guernsey Formation in the Hartville uplift, and the Pahasapa Limestone in the Black Hills.  

In southwestern Nebraska on the crest of the Cambridge arch, Pennsylvanian strata unconformably
overlie Precambrian rocks (pl. 2). Westward on the flank of the arch, truncated Cambrian and Lower Ordovician rocks undivided form a band around Mississippian rocks (Carlson, 1963a, pl. 3). Eastward from the arch into the Central Nebraska basin, Cambrian, Ordovician, Devonian, and Mississippian rocks form northward-trending bands beneath the Pennsylvanian strata. Elsewhere in Nebraska westward from the 100th meridian, except in the northwest and southwest corners, the Pennsylvanian overlies Precambrian rocks.

Pennsylvanian rocks in southeastern Nebraska rest unconformably on Precambrian rocks on the crest of the Nemaha anticline (pl. 2). Westward and northward from the crest, successively younger rocks, of Ordovician, Silurian, Devonian, and Mississippian ages, are found unconformably beneath the Pennsylvanian beds. The rocks of Ordovician through Mississippian ages are also present on the east flank of the anticline, where their areal distribution is partly controlled by pre-Pennsylvanian faulting.

The northern limit of the Pennsylvanian rocks in Nebraska is along a northwestward-trending arcuate line, convex southward, which extends from near the 42d parallel at the Missouri River on the east to the 99th meridian at the Nebraska-South Dakota boundary.

Precambrian and Ordovician rocks underlie the Pennsylvanian in southernmost South Dakota, and Mississippian rocks are north of the Ordovician. The upper and lower contacts of the Ordovician rocks are projected toward the Black Hills through an area of sparse well control.

**PALEOTECTONIC IMPLICATIONS**

Rejuvenation of the ancient continental-interior landmass, Siouixia, and subsequent erosion in Late Mississippian and Early Pennsylvanian time produced the widespread unconformity that separates Pennsylvanian strata from older rocks throughout the northern midcontinent region. Uplift was greatest in the west, where Pennsylvanian rocks rest on Precambrian rocks in most of western Nebraska.

The Cambridge arch, a positive element that trended southward in west-central Nebraska, and the north end of the Nemaha anticline in southeastern Nebraska also were uplifted and deeply eroded at this time. Exposure and erosion prior to deposition of Pennsylvanian rocks ended in Early Pennsylvanian time in western Nebraska and adjacent areas, in Middle Pennsylvanian time in southeastern Nebraska, and in Late Pennsylvanian time in small areas in the central and eastern parts of the State.
### Table 16.—Divisions of the Hartville Formation

<table>
<thead>
<tr>
<th>System</th>
<th>Smith (1903)</th>
<th>Condra and Reed (1935)</th>
<th>Approximate age of units according to Love, Herbst, and Denson (1958); Agassiz (1954); Foster (1958)</th>
<th>This paper (from Tranter and Petter, 1963; Hoyt, 1963)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Permian</td>
<td></td>
<td>Condra, Reed, and Scherer (1940)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hartville Formation</td>
<td>Division I</td>
<td>Cassa Group</td>
<td>Leonard</td>
<td>Cassa Member</td>
</tr>
<tr>
<td></td>
<td>Division II</td>
<td>Broom Creek Group</td>
<td>Wolfcamp</td>
<td>Broom Creek Member</td>
</tr>
<tr>
<td></td>
<td>Division III</td>
<td>Wendover Group</td>
<td>Virgil</td>
<td>Wendover Member</td>
</tr>
<tr>
<td></td>
<td>Division IV</td>
<td>Meek Group</td>
<td>Missouri</td>
<td>Meek Member</td>
</tr>
<tr>
<td></td>
<td>Division V</td>
<td>Hayden Group</td>
<td>Des Moines</td>
<td>Hayden Member</td>
</tr>
<tr>
<td></td>
<td>Division VI</td>
<td>Roundtop Group</td>
<td>Atoka</td>
<td>Roundtop Member</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Reclamation Group</td>
<td></td>
<td>Reclamation Member</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Fairbank Formation</td>
<td></td>
<td>Fairbank Member</td>
</tr>
</tbody>
</table>

## INTERVAL A

**FORMATIONS INCLUDED**

Interval A, which contains rocks of Morrow age, is restricted to the westernmost part of the Nebraska region. In the northwestern part of the State, the interval is a detrital unit that unconformably overlies mostly Mississippian strata but overlaps onto Precambrian rocks. Interval A includes rocks equivalent to the *Fairbank Member* of the Hartville Formation in the Hartville uplift (pl. 13; table 16). The Hartville Formation (Smith, 1903) was separated into *Divisions I* through *VI*, from youngest to oldest, by Condra and Reed (1935). These divisions were later reclassified, and the formation was divided into seven groups and one formation by Condra, Reed, and Scherer (1940). These divisions of the Hartville Formation were subsequently classified as members by Tranter and Petter (1963), as shown in table 16. These members are tentatively correlated with similar units of the Minnelusa Formation of the Black Hills uplift (pl. 13).

On the south flank of the Black Hills uplift and in central South Dakota, interval A consists of a detrital unit at the base of the Minnelusa Formation. This unit was described by Baker (1947, p. 5) as “the basal laterite of the Minnelusa,” which is dated as Early Pennsylvanian because it contains *Chaetetes milleporaceus* and *Mesolobus mesolobus*, and possibly it is equivalent to the “red part of the Amsden” in Wyoming. The basal laterite of the Minnelusa has been correlated with the *Fairbank Member* of the Hartville Formation in the Hartville uplift by Bates (1955, p. 991) and with the *Bell sand* of eastern Wyoming.

Eastward from the Laramie Range to western Nebraska, test holes do not reach interval A, and in the adjacent part of western Nebraska only a few holes penetrate these strata. Consequently, the thickness and eastern limit of the interval in the subsurface can only be estimated. The Casper Formation or its equivalent is probably present in part of the area. In southwestern Nebraska, Morrow rocks are not recognized in test holes, but the pinchout of these rocks probably lies close to the State line. Strata in a test hole in Colorado a few miles from the Nebraska line are assigned to interval A (Maher, 1948, p. 7).

### UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A is traceable across most of the northwestern part of the northern midcontinent region. Interval A conformably underlies interval B and, in the absence of paleontologic control, the boundary is selected on the basis of a marked lithologic change from mainly sandstone interbedded with some red mudstone in interval A to mostly limestone in inter-
val B. Locally, the basal bed of interval B consists of calcareous mudstone, siliceous limestone, limestone-chert breccia, or chert.

THICKNESS TRENDS

The thickness of interval A in western Nebraska and eastern Wyoming is variable, partly because the unit fills karst topography (Bates, 1955, p. 1997) on the underlying Mississippian limestone. The isopach map (pl. 3A) shows a maximum thickness of 115 feet in eastern Wyoming.

The rocks of interval A in South Dakota lie mainly north of the 44th parallel. The edge extends generally eastward from the Black Hills to the Missouri River, where it curves northward rather abruptly. Interval A is less than 50 feet thick near the 44th parallel.

LITHOFAcies TRENDS

Interval A consists dominantly of pink to gray sandstone which is mostly medium grained and locally quartzitic; it contains scattered red mudstone partings. It is mostly the residual detritus of an eroded limestone sequence but includes small amounts of transported detritus. It has been described as "the basal laterite of the Minnelusa" (Baker, 1947, p. 5), and as "the reworked residual mantle from the post-Mississippian erosion surface" (Mc Crae, 1956, p. 85). In the Hartville uplift area and to the south, the rock consists mostly of non-calcareous sandstone and includes very little mudstone (pl. 3B). Toward the Black Hills area the ratio of mudstone to sandstone increases, and locally in southwestern South Dakota mudstone is a major constituent. East of the Black Hills the ratio of mudstone to sandstone is greater than in the Hartville uplift, but sandstone increases in central South Dakota.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Interval A probably consists in part of detritus derived through erosion of a thick limestone sequence of Mississippian age. Precambrian rocks were exposed in much of western Nebraska during Late Mississippian and Early Pennsylvanian time, and erosion in this area produced a mantle of arkosic residue or granite wash. Probably some sediment was transported northwestward from Nebraska shorelines of the lowlying Siouxis landmass of central North America, and a small quantity of sediment may have been transported into the area from a western positive element.

The basal laterite in the Black Hills indicates warm moist climatic conditions. The residual mantle was reworked later in interval A time in a shallow sea.

PALEOTECTONIC IMPLICATIONS

Epeirogenic uplift in Late Mississippian or earliest Pennsylvanian time seems to have initiated deep weathering of Mississippian limestone in eastern Wyoming and near the Black Hills. Widespread karst topography with considerable relief formed on the exposed Mississippian rocks.

In western Nebraska the Mississippian and older rocks were gently tilted westward, and pre-Pennsylvanian erosion exposed Precambrian rocks in a large area (pl. 2). The period of erosion was followed by subsidence in westernmost Nebraska and adjacent States which allowed northward transgression of the Early Pennsylvanian seas and deposition and reworking of the strata of interval A.

INTERVAL B FORMATIONS INCLUDED

Interval B, which consists of rocks of Atoka age, is present in eastern Wyoming, westernmost Nebraska, southwestern and south-central South Dakota, and southeasternmost Nebraska. In the northwestern area interval B consists of the Reclamation Member and part of the Roundtop Member of the Hartville Formation (pl. 13) and, southwest of the Hartville uplift, part of the Casper Formation. On the south flank of the Black Hills, the Minnelusa Formation, which is of Pennsylvanian and Permian age, was divided into lithologic units by Jennings (1959, p. 990); fossil zones within these units have made possible a correlation with units of the Hartville Formation. Interval B farther east in south-central South Dakota also is a part of the Minnelusa Formation (pl. 13).

In the Forest City basin in southeastern Nebraska, 300 feet of Atoka(? ) rocks resting on Mississippian rocks has been assigned to interval B. The sequence was arbitrarily included in the Cherokee Group by Condra and Reed (1943, p. 56), although paleontological proof of its age is lacking for this area. Lithologically similar strata in adjoining parts of the Forest City basin are considered to be Atoka in age by H. R. Wanless (pl. 13; chap. F, this publication).

UPPER BOUNDARY OF INTERVAL B

The boundary between intervals B and C is approximately the Atoka-Des Moines time boundary and is conformable. Throughout much of this region, however, the position of the boundary is uncertain because of lateral lithologic changes that make correlations difficult in widely spaced well logs, and because fossils are commonly absent or not so distributed as to provide an exact boundary between rocks of Atoka and Des Moines age.

In southeastern Nebraska the upper boundary of interval B is the top of an apparently unfossiliferous sequence of black to dark-gray mudstone interbedded with sandstone and thin coal beds. The sequence is overlain by arkosic conglomeratic sandstone or red mudstone assigned to interval C. To the east in Missouri and Iowa
The maximum thickness of interval B in the northeastern part of the Nebraska region is approximately 250 feet along the axial trend of the seaway in eastern Wyoming (pl. 64). Locally, the sea in which interval B was deposited transgressed eastward beyond the limits of interval A, and relatively thin deposits lapped onto pre-Pennsylvanian strata.

In the Forest City basin in southeastern Nebraska the thickness of interval B exceeds 300 feet in several wells. The trend of maximum thickness appears to be north-northeast.

**THICKNESS TRENDS**

Dominantly carbonate marine rocks were deposited along a north-trending belt that marks the seaway in eastern Wyoming and western Nebraska (pl. 6B). Carbonate rocks extend southward from south-central South Dakota as far as the north flank of the Siouxiya landmass. Rocks on the east and west sides of the seaway have a high ratio of detrital to carbonate constituents. Minor amounts of evaporites are present near the common boundary area of South Dakota, Nebraska, and Wyoming.

In the Nebraska part of the Forest City basin, interval B consists of fine sandstone and mudstone in nearly equal quantities.

**LITHOFACIES TRENDS**

In western Nebraska and adjacent parts of Colorado and South Dakota, interval B was deposited in a dominantly marine environment on a stable shelf. The increase in ratio of detrital to carbonate rocks from the center to the margins of the seaway suggests principal sources of detritus in the low-lying Cambridge arch to the east and the Frontrange uplift (fig. 62, chap. N) to the west. A large area made up dominantly of sandstone in western Nebraska suggests that a delta encroached on the seaway from the east in this region.

The principal source of the detritus in interval B in southeastern Nebraska was the Nemaha anticline. Deposition kept pace with subsidence in an apparently shallow restricted basin; individual beds in interval B in this area are thin and widespread and locally contain mud cracks, indicating subaerial exposure. Thin coal beds suggest local swamps.

**PALEOTECTONIC IMPLICATIONS**

Slow subsidence in eastern Wyoming and adjacent areas was generally continuous from the time of interval A through deposition of interval B. The subsidence and the resultant slight widening of the seaway caused sinking and sedimentary burial of small positive elements and marginal areas of larger ones, both east and west. The abundance of shale and limestone in interval B indicates that land areas in central Nebraska were only slightly above sea level and that they contributed mostly clay and silt to the adjacent seas.

A low east-facing escarpment along the Nemaha anticline was the result of faulting and uplift that occurred intermittently from Late Mississippian into early Middle Pennsylvanian time.

**INTERVAL C FORMATIONS INCLUDED**

Interval C in Nebraska and adjoining areas consists of rocks of the Des Moines Series and equivalents and, in some areas, overlying rocks in the basal part of the Missouri Series. In eastern Wyoming part of the Casper Formation along the east side of the Laramie Range is assigned to interval C. Interval C consists of the Hayden Member and part of the underlying Roundtop Member of the Hartville Formation in the Hartville uplift. A dolomite and mudstone sequence, including some black mudstone, in the lower part of the middle member of the Minnelusa Formation in the southern Black Hills was correlated with the Hayden Member by Jennings (1959, p. 900) and is included in interval C (pl. 13).

In central and eastern Nebraska, rocks of the Des Moines Series are divided into the Cherokee Group and the overlying Marmaton Group (pl. 13). These groups are divided into formations where they crop out in eastern Kansas. These formations are thin and lithologically similar, and they are not easily recognized in the subsurface (Condra and Reed, 1943, p. 54). Only the groups are generally differentiated.

A relatively thin sequence of mudstone beds, the Pleasanton Group at the base of the Missouri Series, is included in interval C in central and eastern Nebraska.
The Pleasanton is assigned to interval C because the lower contact of this group generally lies within a mudstone sequence and is difficult to determine, whereas the upper contact lies at the base of a widespread easily identified limestone.

**UPPER BOUNDARY OF INTERVAL C**

The upper boundary of interval C is at the top of the Hayden Member of the Hartville Formation in the area of the Hartville uplift. In the Forest City basin in southeastern Nebraska, it is at the contact between the Pleasanton Group and an overlying limestone in the Kansas City Group. Between these areas the boundary is generally placed at the upward transition from dominantly detrital rocks to carbonate rocks. West of the Cambridge arch this transition commonly lies at, or not far above, the top of a widespread black or dark-gray radioactive mudstone. A similar mudstone that has been recognized in the Hayden Member of the Hartville Formation helps to define the boundary in southeastern Wyoming.

In much of the region interval C grades upward into interval D. Disconformity and unconformity have been recognized locally in the Hartville and Black Hills areas, respectively.

**THICKNESS TRENDS**

Interval C is absent on the Cambridge arch in an irregular area which extends from a point in the southern part of Nebraska northwestward nearly to the southeast flank of the Black Hills uplift in South Dakota (pl. 7A). The thickness on top of the Cambridge arch in southern Nebraska generally does not exceed 75 feet. West of the Cambridge arch the thickness of interval C gradually increases, and near the southwest corner of Nebraska, in the eastern part of the Denver-Julesburg basin, the thickness exceeds 400 feet.

East of the Cambridge arch the thickness of interval C is nearly 450 feet in the Central Nebraska basin near the Kansas line and is slightly more than 300 feet near the South Dakota line.

In southeastern Nebraska interval C is absent on much of the Nemaha anticline, which limits the Central Nebraska basin on the east. The east-facing escarpment of the anticline is the western limit of the Forest City basin, where the maximum thickness of interval C within Nebraska is nearly 800 feet.

**LITHOFACIES TRENDS**

Interval C in western Nebraska and adjoining areas is composed mainly of carbonate rock interbedded with shale in a broad northtrending band that corresponds to the central part of the Early and Middle Pennsylvanian seaway. On the west side of this carbonate belt, the grain size of the detrital interbeds increases, and the carbonate component decreases westward toward the Frontrange uplift. On the west flank of the Cambridge arch along the east side of the seaway, the facies is mostly calcareous terrigenous mudstone, grading to sandy calcareous mudstone at the north end of the arch in South Dakota. A thin but distinctive radioactive black mudstone is present southeast of the Black Hills (MacLachlan and Bieber, 1963, p. 87) and also farther south along the west flank of the Cambridge arch in western Nebraska (Maher, 1948, p. 6; Taylor, 1958, p. 65).

Anhydrite beds in eastern Wyoming, too thin to be represented on plate 7B, apparently are remnants of evaporite deposits that originally extended from eastern Wyoming northeastward through central South Dakota. An anomalous calcareous quartz sandstone facies in the southern Black Hills is an insoluble residue produced by extensive post-Pennsylvanian solution of sandy carbonates and evaporites in intervals C and D. The amount of thinning may have been about 200 feet (Baker, 1947, p. 4).

On the east flank of the Cambridge arch, interval C consists mainly of mudstone, some siltstone, lenticular beds of limestone, and, locally, thin beds of fine-grained sandstone. Near the north flank of the Nemaha anticline in eastern Nebraska, the interval locally is mostly argillaceous sandstone.

Alternating beds of mudstone, siltstone, and sandstone and a few thin limestone strata and thin coal beds constitute interval C in the Nebraska part of the Forest City basin. The limestone, as well as the mudstone and siltstone, is gray, dark gray, or black.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

A seaway in which the water was clear and warm extended northward across much of eastern Colorado through eastern Wyoming and parts of adjacent States. This seaway was the site of extensive carbonate deposition, and it also received sediment from nearby positive elements, including the Frontrange uplift, uplifts near the present Laramie Range, and the Cambridge arch. A maximum water depth of 200 feet in the seaway was postulated by Agatston (1954, p. 564) because of the presence of fusulinids and brachiopods.

Marine conditions also prevailed in the Central Nebraska basin between the Cambridge arch and the Nemaha anticline, although in this area the sea was probably very shallow and the bottom muddy. Detritus was derived from erosion of relatively low bordering positive areas, principally Siouxia to the northeast. The Nemaha anticline in southeastern Nebraska remained as a large island between the Forest City and Central Nebraska basins during interval C time.

**PALEOTECTONIC IMPLICATIONS**

A broad low positive element, part of Siouxia, persisted as land with little change during the time of inter-
vals A and B in most of Nebraska. Near the middle of interval C time, seas encroached on this positive element from the southwest. In southeastern Nebraska broad folding and faulting kept a part of the crest of the Nemaha anticline above sea level.

Relatively stable to moderately unstable shelf conditions are suggested by the lithology of interval C in the northwestern part of the region. Limestone interbedded with mudstone and sandstone is conformable between intervals B and D in much of the area. Local movement may be represented in the Hartville uplift, where strata of interval C are separated from interval D by a minor disconformity, and near the Black Hills, where interval C (less fossiliferous, more sandy, and more dolomitic than in the Hartville area) is locally unconformable at its upper boundary (Condra and others, 1940, p. 41). The lack of sandstone deposits on the west flank of the Cambridge arch indicates that uplift of the arch was slight.

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D, which consists of rocks of Missouri age, is the Meek Member of the Hartville Formation in the vicinity of the Hartville uplift. Southwest of the uplift, on the east flank of the Laramie Range near the Colorado line, the interval consists of coarse arkosic beds of the Casper Formation. Part of the Minnelusa Formation makes up interval D in the southern Black Hills. On the basis of more recent subsurface data, both lithologic and paleontologic correlations between the Hartville and Black Hills areas have undergone some modifications (Agatston, 1954; Bates, 1955; Jennings, 1959). East of these areas across Nebraska to the Nemaha anticline, the boundaries of interval D are readily determined, but the Kansas City and Lansing Groups are not separated.

In Nebraska, east of the Nemaha anticline, the Kansas City and Lansing Groups, with their component formations and members, are assigned to interval D. The nomenclature currently used by the Nebraska Geological Survey for this area is shown in table 17. Interval D west of the Nemaha anticline is presumed to contain rocks equivalent to the Kansas City and Lansing Groups, but neither group nor formation boundaries are projected west of the anticline.

**UPPER BOUNDARY OF INTERVAL D**

The upper boundary of interval D in the Hartville uplift is at the bottom of a thin lenticular sandstone at the base of the Wendover Member of the Hartville Formation. A widespread marker bed of gray and pink chert, about 2 feet thick, lies about 25 feet below the top of interval D (Love and others, 1953). The chert bed is overlain by a black petriferous limestone containing abundant fusulinids of Missouri age. The basal bed of interval E, locally disconformable on interval D, is referred to as the “Leo sand equivalent” by Bates (1955, p. 189).

Throughout most of the Nebraska region, interval D is overlain unconformably by a mudstone sequence of variable thickness and lithology at the base of interval E. At exposures in southeastern Nebraska, the upper boundary of the interval is a disconformity marked by a thin zone of weathered shale.

**THICKNESS TRENDS**

The thickness trends of interval D (pl. 8A) reflect widespread marine transgression from the south and the southeast. Interval D has a maximum thickness of

<table>
<thead>
<tr>
<th>Series</th>
<th>Group</th>
<th>Formation</th>
<th>Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lansing</td>
<td>Stanton Limestone</td>
<td>South Bend Limestone</td>
<td>Rock Lake Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Stoner Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Eudora Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Captain Creek Limestone</td>
</tr>
<tr>
<td></td>
<td>Vilas Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Plattsburg Limestone</td>
<td>Spring Hill Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Hickory Creek Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Merriam Limestone</td>
</tr>
<tr>
<td></td>
<td>Bonner Springs Shale</td>
<td>Farley Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Island Creek Shale</td>
</tr>
<tr>
<td></td>
<td>Wyandotte Limestone</td>
<td>Argentine Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Quindaro Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Frisco Limestone</td>
</tr>
<tr>
<td>Missouri</td>
<td>Lane Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Iola Limestone</td>
<td>Mustang Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Chanute Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kansas City</td>
<td>Drum Limestone</td>
<td>Corbus City Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Cenozoic City Limestone</td>
</tr>
<tr>
<td></td>
<td>Quivira Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sarpy Limestone</td>
<td>Westerville Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Wea Shale</td>
</tr>
<tr>
<td></td>
<td>Fontana Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Dennis Limestone</td>
<td>Winterset Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Stark Shale</td>
</tr>
<tr>
<td></td>
<td>Galesburg Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Swope Limestone</td>
<td>Bethany Falls Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Hoackney Shale</td>
</tr>
<tr>
<td></td>
<td>Lodore Shale</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Hertha Limestone</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
310 feet in the Forest City basin and is approximately 300 feet thick in the Central Nebraska basin and the southwest corner of Nebraska. The rocks of the interval gradually thin northward in Nebraska.

The Nemaha anticline in southeastern Nebraska and the Cambridge arch in western Nebraska were completely covered by sediments, apparently for the first time in the Pennsylvanian Period, during the deposition of interval D. The thickness of interval D is generally less than 200 feet in western Nebraska and parts of adjacent States.

LITHOFACIES TRENDS

Carbonate rock is dominant in interval D (pl. 8B) in about two-thirds of Nebraska and adjacent areas. A high percentage of mudstone near the Cambridge arch contrasts sharply with the marine carbonate rocks in surrounding areas.

Post-Pennsylvanian solution collapse accounts for an anomalous sandstone facies in the southern Black Hills. Anhydrite deposits in the area between the Hartville uplift and the Black Hills indicate the approximate southwestern limit of an extensive evaporite basin in west-central South Dakota.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Probable sources of detritus for interval D in the western part of the region include the Frontrange uplift and the more remote ancestral ranges of Colorado and Wyoming. White fine- to medium-grained well-sorted sandstone occurs as lenticular interbeds 5-40 feet thick in marine carbonate sequences. The sandstone probably was derived in part from older Pennsylvanian rocks such as the Tensleep Sandstone in western Wyoming (Agatston, 1954, p. 582). Oscillations of the strandline during a general marine advance from the south resulted in reworking of continental and marine detrital deposits in some areas.

In much of the midcontinent region, open-water marine carbonate deposition was dominant. A shallow sea is indicated by the faunal assemblage in the limestone beds. Locally in southwestern South Dakota, restricted marine environments resulted in deposition of evaporites.

During deposition of the lower part of interval D, a large part of the Cambridge arch was low, but positive. As the transgressing seas gradually covered the arch, the weathered surface material was eroded and deposited nearby as detritus of interval D. The accumulation of detritus was thickest on the south and southwest flanks of the arch.

PALEOTECTONIC IMPLICATIONS

Positive elements in Nebraska at the beginning of interval D time included two major structural features that had persisted throughout Early and Middle Pennsylvanian time. These features—the Nemaha anticline and Cambridge arch—subsided and were inundated before the middle of interval D time. Subsidence apparently was gradual and nearly continuous through most of Nebraska and adjacent States.

INTERVAL E
FORMATIONS INCLUDED

Interval E consists of rocks considered to be Virgil in age. In southeastern Wyoming interval E is the Wendover Member of the Hartville Formation and equivalent (pl. 13). In the subsurface northeast of the Hartville uplift and in the southern Black Hills, the upper part of the middle member of the Minnelusa Formation is placed in interval E on the basis of correlation of electric logs.

On the broad Cambridge arch in southwestern Nebraska, the large quantity of subsurface data available makes possible the identification of major subdivisions within interval E. The Virgil Series includes, in ascending order, the Douglas, Shawnee, and Wabunsee Groups. These groups are identified, on the basis of lithology, in the subsurface on both sides of the arch and also in the Central Nebraska basin. The Wabunsee Group consists of interbedded limestone, sandy mudstone, and some sandstone; the Shawnee Group consists mainly of thick beds of limestone separated by thin beds of calcareous sandstone and red, gray, and black mudstone; and the Douglas Group is chiefly dark-gray, red, or black mudstone with a few discontinuous beds of limestone.

In southeastern Nebraska the nomenclature for rock units included in interval E as shown in table 18 and on plate 13 is from Moore (1932), Condra and Reed (1943, revised 1959), Moore and Mudge (1956), and Burchett and Reed (1967, p. 38, 41). The groups defined in outcrop areas can be traced westward in the subsurface through the Central Nebraska basin and into the northwestern part of the State, but formations within the groups are difficult to identify nearly everywhere in the subsurface. Considerable uncertainty and confusion has resulted from attempts to correlate certain subsurface units with thin units recognized in exposures to the east.

UPPER BOUNDARY OF INTERVAL E

In the northwestern part of the region the upper boundary of interval E, which marks the top of the Pennsylvanian System, is placed at the base of a mudstone known as the red shale marker or the red marker. The value of this stratum as a marker bed was first recognized in the subsurface of the Lance Creek, Mule Creek, and Dewey Dome oil fields north of the Hartville uplift (Krampert, 1940a, b, c; fig. 17). The red marker is present in the subsurface throughout a large area, and it has been identified in outcrops both in the vicinity of the
<table>
<thead>
<tr>
<th>Series</th>
<th>Group</th>
<th>Formation</th>
<th>Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wabaunsee</td>
<td>Longdon Shale</td>
<td>Brownsville Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pong Creek Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Grayhorse Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Plumb Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Nebraska City Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tarkio Limestone</td>
<td>French Creek Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Jim Creek Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Friedrich Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Willard Shale</td>
<td>Paloum Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Morton Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Unnamed shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Dover Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Osborn Shale</td>
<td>Wabawus Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Soldier Creek Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Banglaspine Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Silver Lake Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rulo Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Cedar Vale Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Happy Hollow Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>White Cloud Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Howard Limestone</td>
<td>Elsco1 Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Harveyville Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Reading Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Severy Shale</td>
<td>Cool Creek Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Holt Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Do Bois Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Turner Creek Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sheldon Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Jones Point Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Council Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Iona Point Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Hartford Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Colbourn Shale</td>
<td>Empire Creec1 Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Burroak Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Lorr Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rock Bluffs Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Oskaloosa Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Oawanick Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Tecumseh Shale</td>
<td>Rouse Creek Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Otz Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Kewanah Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Leompton Limestone</td>
<td>Avoca Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>King Hill Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bell Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Queen Hill Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Big Springs Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Despoham Shale</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Spring Branch Limestone</td>
<td></td>
</tr>
</tbody>
</table>

**Table 18.—Stratigraphic units of southeastern Nebraska assigned to interval E—Continued**

<table>
<thead>
<tr>
<th>Series</th>
<th>Group</th>
<th>Formation</th>
<th>Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wabaunsee</td>
<td></td>
<td></td>
<td>Kanwaka Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Clay Creek Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Jackson Park Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Kenton Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Heumader Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Plattsmouth Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Heebner Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Leavenworth Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Snyderville Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Toronto Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Lawrence Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Haskell Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Little Pana Creek Shale</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Showmaker Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Plattford Formation</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Nebraska Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Unnamed shale</td>
</tr>
</tbody>
</table>

Black Hills (Jennings, 1959, p. 99) and in the Hartville uplift (Maugham, 1967, p. 132).

In northwestern Nebraska and southeastern Wyoming, the position of the red marker, or an equivalent bed, can be identified on mechanical logs (MacLachlan and Bieber, 1963, p. 88). The red marker overlies an unconformity (Foster, 1958, p. 39). Fusulinids from rocks below the red marker indicate a Pennsylvanian age (Love and others, 1953; Agatston, 1954; Jennings, 1959), and fusulinids from rocks above are Permian (Thomas, 1940; Agatston, 1954; Jennings, 1959; Momper, 1963).

The correlations of the American Stratigraphic Co. (unpub. data), Julifs (1953), Reed (1955), R. R. Burchett (Nebraska Geological Survey, unpub. cross section, 1959), and R. R. Burchett and M. P. Carlson (Nebraska Geological Survey unpub. cross section, 1959) are followed in determining the boundary in westernmost Nebraska.

The upper boundary of the interval in the northern part of the Central Nebraska basin is generally placed at the top of a thick sequence of interbedded limestone and gray mudstone. The overlying Admire Group of Permian age consists of red and green mudstone interbedded with sandstone, sandy mudstone, and thinbedded limestone. The position of the contact in this area is in accordance with that determined by Smith and Burchett (1967).

In the outcrop area of southeastern Nebraska, the top of the Brownville Limestone Member of the Wood Siding Formation (table 18) is generally recognized as the boundary between the Pennsylvanian and Permian Systems (Moore and others, 1951, p. 58; Moore and Mudge, 1956, fig. 1; Mudge and Yochelson, 1962). In the south the lowest limestone stratum in Kansas that contains a Permian fauna is 85 feet above the Brownville Limestone Member (Mudge, 1957, p. 88). The fauna of Virgil age and the lithology of the Brownville are distinctive, and the top of the limestone, therefore, is useful as an arbitrary systemic boundary.
THICKNESS TRENDS

The maximum known thickness of the interval in the Nebraska region is 828 feet in the Forest City basin. Except for abrupt thinning across the Nemaha anticline, the thickness of interval E decreases gradually northwestward. In western Nebraska and eastern Wyoming the interval is generally 100-200 feet thick. Subparallel isopachs in the northeastern part of Nebraska are the result of pre-Cretaceous erosion (pl. 9A). Minor changes in the thickness of strata of interval E in eastern Nebraska may reflect very slight tectonic instability of the northern midcontinent region (Reed, 1954, p. 114).

The north limit of interval E extends southeastward from northeastern Wyoming across northern Nebraska to the Missouri River (pl. 9A). Interval E is truncated along this line beneath Cretaceous rocks.

LITHOFACIES TRENDS

Interval E in the western part of Nebraska and adjacent areas farther west consists chiefly of limestone (pl. 9B). In some areas the limestone is sandy, and locally, near the western margin of the region, sandstone is dominant. The proportion of mudstone increases eastward in Nebraska except for an area of dominantly carbonate rock that extends from the Kansas border northward for about 40 miles into south-central Nebraska. A mudstone facies borders the west side of the Cambridge arch.

SOURCES AND ENVIRONMENTS OF DEPOSITION

The ancestral ranges in Colorado and southern Wyoming continued as highlands that furnished major amounts of sediment to eastern Wyoming. Except for these areas, the Nebraska region was inundated by a shallow sea. Brackish water and swampy conditions are indicated by a few thin coal beds in southeastern Nebraska in the upper part of the interval. The source of silt and clay in the eastern part of the region was low-lying land an unknown distance to the northeast.

PALEOTECTONIC IMPLICATIONS

The unconformity that separates interval E from the overlying Permian rocks in eastern Wyoming and adjacent areas is the result of broad uplift followed by shallow erosion during late Virgil or earliest Permian time. Uplift in this area probably coincided with a late stage in the Frontrange orogeny (Maughan, 1967, p. 132).

In southwestern and south-central Nebraska there is no evidence of interruption in deposition between interval E and the overlying rocks. The facies of the uppermost Pennsylvanian rocks in southern Nebraska is very similar to that of the overlying Permian. Local disconformities are known, but no evidence of regional truncation or beveling has been found (Reed, 1954, p. 114; Mudge and Yochelson, 1962; Mudge, 1967, p. 98).

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

THICKNESS TRENDS

The total thickness of rocks of Pennsylvanian age in Nebraska and adjacent areas generally tends to decrease from southeastern Nebraska northwestward (pl. 11). A conspicuous large area of thinning in northern Nebraska and adjacent South Dakota resulted from nondeposition in the early part of Pennsylvanian time and from erosion of intervals D and E in post-Pennsylvanian time.

The zero isopach delineates the eastern flank of the Laramie Range, the Hartville uplift, and the southern perimeter of the Black Hills. A zero isopach extending from central South Dakota through northeastern Nebraska marks the present subsurface limit of Pennsylvanian rocks in that area, but the original thickness there was greatly reduced by pre-Cretaceous erosion.

The maximum known thickness of Pennsylvanian rocks is 2,085 feet in the Forest City basin of southeastern Nebraska (pl. 11). Westward in the Central Nebraska basin, these rocks are at least 1,285 feet thick, and, near the Kansas-Nebraska State line in south-central Nebraska, extrapolated isopachs suggest a thickness of about 1,400 feet. The original thickness in the northern part of the Central Nebraska basin has been considerably reduced by pre-Cretaceous erosion. Pennsylvanian rocks are 1,144 feet thick near the southwest corner of Nebraska on the west side of the Cambridge arch. The thickness in eastern Wyoming is generally about 700 to 800 feet in a relatively narrow band that trends northward to northwestward. The thickness decreases westward and southwestward toward the Laramie Range.

PALEOTECTONIC IMPLICATIONS

An area consisting of a large part of Nebraska and southern South Dakota persisted as a positive area during the early part of the Pennsylvanian. It extended eastward from the Cambridge arch, to the Nemaha anticline in Nebraska, and northeastward into eastern South Dakota. This area was a part of the stable region of the continent, Siouxia (Schuchert, 1923, p. 165), and it was a segment of the Transcontinental arch (Eardley, 1962, p. 38). Tectonic activity was intermittent and generally minor during the deposition of the Pennsylvanian. The Cambridge arch apparently was nearly stable in Late Pennsylvanian time, but variations in thicknesses of intervals B through E indicate continuing intermittent movement of the Nemaha anticline, especially along the Humboldt fault on the east flank. Regional subsidence that allowed the sea to spread widely apparently began near the middle of the time of interval C. Except on the highest parts of the Cambridge arch and the Nemaha anticline, strata of interval C eventually covered the State and adjacent areas. Early in the time of interval D, the Cambridge arch and the Nemaha anticline were inundated, and sediment accumulated.
everywhere in the region to the end of the Pennsylvanian Period.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

**UNITS OVERLYING PENNSYLVANIAN**

In Nebraska and adjacent areas, Pennsylvanian rocks are directly overlain by rocks of Permian, Cretaceous, or Quaternary ages (pl. 12).

In most of Nebraska and adjacent areas, units of Permian age lie directly above the Pennsylvanian. Cretaceous rocks, assigned to the Dakota Sandstone, unconformably overlie Pennsylvanian strata in northeastern Nebraska and south-central South Dakota. Glacial and alluvial deposits of Quaternary age unconformably mantle the Pennsylvanian beds in southeastern Nebraska.

Tertiary rocks are widespread in Nebraska, but nowhere are they in contact with Pennsylvanian rocks. Other post-Pennsylvanian rocks present in the Nebraska region but not in contact with Pennsylvanian rocks are the Triassic in the western part of the region, the Jurassic in the western half, and the Tertiary over still wider areas.

**PALEOTECTONIC IMPLICATIONS**

Sedimentation throughout much of Nebraska continued through the end of Pennsylvanian time and into Early Permian time, indicating a continuation of the relatively stable tectonic environment in much of this region.

In eastern Wyoming and adjacent areas, an unconformity marks the top of Pennsylvanian rocks. The distinctive red marker unit, probably a lateritic soil or a reworked laterite (Foster, 1958, p. 39), is the basal Permian stratum above this unconformity, and regional uplift of the area in latest Pennsylvanian or earliest Permian time is inferred. It probably coincided with late stages of uplift of the Front Range (Maughan, 1967, p. 132).

Recurrent regional uplift, tilting, and beveling are reflected in numerous unconformities between or within rocks of Permian, Triassic, Jurassic, Cretaceous, and Tertiary age. The principal structures present in Pennsylvanian time persisted at depth; some modifications and new structures that resulted from Laramide activity are shown in figure 17. Because of the blanketlike deposition and minimal deformation of Tertiary rocks, the only positive elements evident on the present surface are the Black Hills and the partly exhumed Hartville uplift.

**REFERENCES**


Howard, W. V., 1941, Geology of western Nebraska demands careful prospecting [abs.]: Oil and Gas Journ., v. 39, no. 51, p. 11.


MacLachlan, James, and Bieber, Alan, 1963, Permian and Pennsylvanian geology of the Hartville uplift-Alliance basin-Chadron arch area, in Rocky Mountain Assoc. of Geologists Guidebook 14th Field Conf., northern Rocky Mountain basin and adjacent uplifts, 1963: p. 84-94.

Maher, J. C., 1948, Subsurface geologic cross section from Baca County to Yuma County, Colorado: Kansas Geol. Survey Oil and Gas Inv. 6, 11 p.


Oil and Gas Journal, 1941, Operating ideas: Oil and Gas Jour., v. 39, no. 37, p. 43.


1954, Central Nebraska has possibilities: World Oil, v. 139, no. 6, p. 113-116, illus.


Reed, E. C., Svoboda, R. F., Prichard, G. E., and Fox, Jeannette, 1958, Map of Nebraska showing areal distribution of pre-Pennsylvanian rocks, anticlines, and basins, oil and gas fields, pipelines, and unsuccessful test wells: U.S. Geol. Survey Oil and Gas Inv. Map Oil-198.


Kansas

By GARY F. STEWART

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES. PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-H
CONTENTS

Abstract .................................................. 127
Region defined ............................................... 127
Paleogeology ................................................ 127
- Units underlying Pennsylvanian ........................................ 127
- Lower boundary of Pennsylvanian ....................................... 127
- Pre-Pennsylvanian topography ......................................... 129
- Paleotectonic implications ............................................ 130

Interval A .................................................. 130
- Formations included ............................................... 130
- Upper boundary of interval A .......................................... 130
- Thickness trends .................................................... 131
- Lithofacies trends .................................................. 132
- Sources and environments of deposition ............................... 132
- Paleotectonic implications ............................................ 132

Interval B .................................................. 133
- Formations included ............................................... 133
- Upper boundary of interval B .......................................... 133
- Thickness trends .................................................... 134
- Lithofacies trends .................................................. 134
- Sources and environments of deposition ............................... 135
- Paleotectonic implications ............................................ 135

Interval C .................................................. 136
- Formations included ............................................... 136
- Upper boundary of interval C .......................................... 136
- Thickness trends .................................................... 137
- Lithofacies trends .................................................. 137
- Sources and environments of deposition ............................... 138

Interval C—Continued
- Paleotectonic implications ............................................ 140

Interval D .................................................. 140
- Formations included ............................................... 140
- Upper boundary of interval D ......... . ............................ 140
- Thickness trends .................................................... 140
- Lithofacies trends .................................................. 141
- Sources and environments of deposition ............................... 141
- Paleotectonic implications ............................................ 144

Interval E .................................................. 145
- Formations included ............................................... 145
- Upper boundary of interval E .......................................... 145
- Thickness trends .................................................... 146
- Lithofacies trends .................................................. 146
- Sources and environments of deposition ............................... 147
- Douglas Group ...................................................... 147
- Shawnee Group ...................................................... 149
- Wabaunsee Group ................................................... 150
- Paleotectonic implications ............................................ 152

Total thickness of Pennsylvanian rocks .................................. 153
- Thickness trends .................................................... 153
- Paleotectonic implications ............................................ 153
- Geologic units directly above Pennsylvanian System .................. 153
- Units overlying Pennsylvanian ........................................ 153
- Paleotectonic implications ............................................ 153

References .................................................. 154

ILLUSTRATIONS

[For listing of plates (in separate case) see volume "Contents"]

Figure 18. Map showing distribution of Pennsylvanian rocks by intervals in Kansas .................................................. 128
19. Map showing structural elements of Kansas in Late Mississippian and Early Pennsylvanian time ........................................ 128
20. Map showing counties in Kansas ...................................... 130
21. Type well of Kearny Formation, sec. 23, T. 22 S., R. 38 W., Kearny County, Kans .................................................. 131
22. Map showing distribution of types of lithology at upper boundary of interval A in Kansas .................................................. 132
23. Map showing distribution of types of lithology at upper boundary of interval B in Kansas .................................................. 133
24. Section showing stratigraphic relations of interval B and the lower part of interval C in eastern Kansas ........................................ 134
25. Map showing distribution of types of lithology at upper boundary of interval C in Kansas .................................................. 137
26. Maps showing limestone units in interval C in Kansas ................. 138
27. Generalized section of the Cherokee Group of southeastern Kansas .................................................. 139
28. Generalized section of interval D in eastern Kansas .................. 141
29. Generalized section of interval D rocks in eastern Kansas, showing a megacyclothem that contains three cycloths ............................. 142
30. Generalized section of uppermost Pennsylvanian (interval E) and lowermost Permian rocks in eastern Kansas ........................................ 143
31. Generalized section of interval D rocks in eastern Kansas, showing a megacyclothem that contains three cycloths ............................. 144
32. Map showing distribution of types of lithology at upper boundary of interval E in Kansas .................................................. 145
33. Generalized section of Douglas Group in outcrop area of eastern Kansas .................................................. 146
34. Section showing megacyclothemes and cycloths of the Shawnee Group .................................................. 149
35. Generalized section of Wabaunsee Group in outcrop area of eastern Kansas .................................................. 151

TABLES

Table 19. Major stratigraphic units underlying the Pennsylvanian in Kansas .................................................. 129
20. Major stratigraphic units of Kansas, as assigned to intervals A, B, C, D, E .................................................. 131
21. Rock types and fauna of cycloths in megacyclothem 1 of the Shawnee Group .................................................. 148
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART 1: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

KANSAS

By GARY F. STEWART

ABSTRACT

Pennsylvanian strata of Kansas lie on rocks ranging in age from Precambrian to Mississippian. Precambrian and lower Paleozoic rocks underlie the Pennsylvanian on the Central Kansas uplift and Nemaha anticline. Mississippian rocks underlie the Pennsylvanian in the Forest City, Cherokee, Salina, and Sedgwick basins and in the Hugoton embayment of the Anadarko basin. These structural elements, in existence in Mississippian time and rejuvenated during Pennsylvanian time, influenced both kinds and thicknesses of Pennsylvanian sediment.

Pennsylvanian seas encroached mainly from the south. Sedimentation began in the Hugoton embayment during time of interval A (Morrow). Most of the Hugoton embayment and the Cherokee and Forest City basins were submerged during time of interval B (Atoka and earliest Des Moines). All of Kansas except crests of the Nemaha anticline and Central Kansas uplift was covered by sediment of interval C (Des Moines and earliest Missouri). Sediment of intervals D (Missouri) and E (Virgil) extended throughout the State. Pennsylvanian rocks are slightly thicker than 3,000 feet in the deepest part of the Hugoton embayment; they are thinner than 700 feet on crests of the Nemaha anticline and Cambridge arch.

The Pennsylvanian consists of thin but extensive limestones, marine and nonmarine mudstones and sandstones, and terrestrially deposited coals. The strata compose many cyclothems, especially in intervals C, D, and E. Sediments were deposited almost entirely in shallow marine, paralic, and terrestrial environments. Seas were mainly clear, warm, and moderately low in energy. Coastal regions were locally emergent, swampy, and thickly vegetated and were crossed by sluggish streams. The source area of terrigenous sediment during time of intervals A and B was mainly the terrane of lower Paleozoic rocks in Kansas. Source areas of terrigenous sediment during times of intervals C, D, and E were mainly the emergent Arbuckle Mountains and Ouachita Mountains in Oklahoma and the Ozark dome in Missouri.

Pennsylvanian strata were overlapped by Permian beds that probably extended throughout Kansas. Permian beds were eroded from above the Pennsylvanian in eastern Kansas, perhaps mostly during the Triassic and Jurassic Periods. This terrane may have been overstepped by Cretaceous strata. Pennsylvanian strata of eastern Kansas were eroded to peneplainlike topography during the Tertiary Period. Rivers of eastern Kansas have been superimposed onto Pennsylvanian strata during post-Pliocene time.

REGION DEFINED

The region discussed in this chapter consists of the State of Kansas. Pennsylvanian rocks are exposed in the eastern part and occur in the subsurface throughout the State except in the extreme southeastern area (fig. 18). Beds dip gently westward, and rocks of the Pennsylvanian System thicken westward to a maximum of more than 3,000 feet in southwestern Kansas.

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

Pennsylvanian rocks overlie Mississippian rocks throughout most of Kansas (pl. 2). In some areas on the Central Kansas uplift and Nemaha anticline (fig. 19; pl. 2) rocks of Pennsylvanian intervals C and D lie unconformably on Paleozoic rocks older than Mississippian or on Precambrian rocks.

Pre-Mississippian rocks underlying the Pennsylvanian were not studied in detail during this investigation, and they are broadly grouped on the paleogeologic map (pl. 2; table 19). The Mississippian and Devonian Chattanooga Shale is mapped with the Hunton Group of Silurian and Devonian age. The Silurian and Devonian components of the Hunton Group have been differentiated in only a few places in Kansas. Similarly, the Arbuckle Group is a unit of Late Cambrian and Early Ordovician age; rocks of the two ages have been differentiated by other authors only in eastern Kansas. The Maquoketa Shale, Viola Limestone, and Simpson Group are combined as a map unit of Ordovician age. The Reagan Sandstone is Late Cambrian in age and is mapped with Precambrian rocks because of the limited areal extent of the Reagan where it directly underlies Pennsylvanian rocks on higher parts of the Central Kansas uplift and Cambridge arch.

LOWER BOUNDARY OF PENNSYLVANIAN

Throughout Kansas, lowermost Pennsylvanian rocks generally are coarsely clastic, and they lie unconformably upon deeply weathered and eroded rocks that are
dominantly limestone and dolomite. In general, the terrane of older rocks is covered by a veneer of residual material, and the reworked upper part of the residuum is accepted in Kansas as being the lowermost Pennsylvanian deposit. The boundary between reworked and undisturbed residual material commonly is difficult to recognize in oil-well bit cuttings. Accordingly, the lower boundary of the Pennsylvanian is not precisely known in some parts of Kansas.

Residuum on Mississippian carbonate rocks in southeastern Kansas commonly is composed of tripolitic chert that ranges in thickness from a few inches in the easternmost part of the area (Howe, 1956, p. 31) to more than 100 feet farther west in Sumner County (Lee and Girty, 1940, p. 76) (fig. 20). In central Kansas, residual material that developed on weathered dolomite of the Arbuckle Group consists mostly of chert, clay, silt, and sand (Walters, 1946, p. 695).

East of the Nemaha anticline (fig. 19) lowermost Pennsylvanian rocks dominantly are clayey mudstone, but sandstone that contains fragments of Mississippian chert is present in some places; in the Forest City basin (fig. 19) basal Pennsylvanian rocks contain a small amount of arkose (Lee, 1943, p. 82). In a few places in eastern Kansas, lowermost Pennsylvanian rocks occur in solution openings or caves, some tens of feet below the surface of the Mississippian or older Paleozoic carbonate rocks.
In much of western Kansas, rock identified as lowermost Pennsylvanian commonly is a conglomerate of coarse sandstone and is called the Pennsylvanian basal conglomerate. It comprises variegated and locally oxidized mudstone, sandstone, and chert and ranges in thickness from a featheredge to more than 150 feet (Ver Wiebe, 1941, p. 27). In southwestern Kansas, Pennsylvanian sandstone, limestone, and mudstone overlie similar rocks of Chester age (pl. 2), and in some places the unconformity at the base of the Pennsylvanian is obscure (McManus, 1959, p. 19).

The base of the Pennsylvanian is as old as Morrow in western Kansas and as young as Missouri on some parts of the Central Kansas uplift and Nemaha anticline (figs. 18, 19).

table 19.—Major stratigraphic units underlying the Pennsylvanian in Kansas

<table>
<thead>
<tr>
<th>System</th>
<th>Series</th>
<th>Group or formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mississippian</td>
<td>Chester</td>
<td>Unnamed</td>
</tr>
<tr>
<td></td>
<td>Meramec</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Osage</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Kinderhook</td>
<td></td>
</tr>
<tr>
<td>Devonian</td>
<td>Upper Devonian</td>
<td>Chattanooga Shale</td>
</tr>
<tr>
<td>Silurian</td>
<td>Upper Ordovician</td>
<td>Maquoketa Shale</td>
</tr>
<tr>
<td>Ordovician</td>
<td>Middle Ordovician</td>
<td>Viola Limestone</td>
</tr>
<tr>
<td></td>
<td>Lower Ordovician</td>
<td>Arbuckle Group</td>
</tr>
<tr>
<td>Cambrian</td>
<td>Upper Cambrian</td>
<td>Reagan Sandstone</td>
</tr>
<tr>
<td>Precambrian rocks</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

pre-pennsylvanian topography

The paleogeomorphology of the surface beneath Pennsylvanian rocks is well known in only a few places in Kansas. In Mississippian rocks in Jefferson and Leavenworth Counties in northeastern Kansas, features interpreted as caves are filled with black clayey mudstone of Pennsylvanian age (Lee, 1943, p. 78-79).

The Cherokee Group in much of southeastern Kansas has been described as overlying a weakly developed karst surface (Pierce and Courtier, 1937; Saueracker, 1965, p. 26-27). However, the surface beneath this group in the Joplin district of extreme southeastern Kansas and southwestern Missouri is described as "karst topography of great detail" by Smith and Siebenthal (1907, p. 11).

Probable sinkholes in Mississippian rocks have been discovered in Dickinson and Wabaunsee Counties (fig. 20) in east-central Kansas (Shenk, 1955, p. 1; Smith and Anders, 1951, p. 29) and in Sumner County in south-central Kansas (Lee and Girty, 1940, p. 76-77). Isolated sinkholes less than 20 acres in area and solution valleys 20-80 feet deep developed in rocks of the Arbuckle Group in Barton and Russell Counties (Walters, 1946; Ver Wiebe, 1941, p. 26-27, 82).

The widespread development and variable thickness of residual chert in the upper part of the Mississippian have been interpreted by Goebel (1966, p. 132) as evidence of karst topography throughout western Kansas. Thus, buried moderately developed karst topography may be present in several places in Kansas and perhaps in all areas in the State where carbonate rocks underlie the Pennsylvanian.

The amount of erosional relief on the surface beneath Pennsylvanian rocks is not well documented, but in some areas it may have been at most only a few tens of feet. In northeastern Kansas the pre-Pennsylvanian surface is described as having had peneplainlike topography with broad shallow valleys (Lee and Payne, 1944, p. 60-61). In southeastern Kansas, relief on the surface of Mississippian rocks locally may have been as little as 20 feet (Pierce and Courtier, 1937, p. 37-38), and in Greenwood County (fig. 20) it rarely exceeds 100 feet within a township (Bass, 1936, p. 21).

A pre-Pennsylvanian "mesa" in rocks of the Arbuckle Group was mapped by Ver Wiebe (1941, p. 82-84, pl. 1) in Barton and Russell Counties on the Central Kansas uplift. This feature includes about 12 square miles of karst terrain, and in some areas the slope of the surface is less than 10 feet per mile. Two pinnacles of Precambrian rocks that stand above the Arbuckle terrace are 30 and 60 feet high and less than 160 acres in area. Slope of the eroded surface of pre-Pennsylvanian rocks on the crest of the Cambridge arch (fig. 19) locally was only about 20 feet per mile (Merriam and Atkinson, 1955, p. 20-21).

Late or post-Mississippian erosional relief in parts of the Nemaha anticline and Central Kansas uplift presumably was more than that in other areas of Kansas; before deposition of Pennsylvanian sediments these two structural features may have had surface expression as ranges of low hills within broad plains of exceedingly low relief.
**PALEOTECTONIC IMPLICATIONS**

Most of the structural features that influenced deposition of Pennsylvanian sediments were present in part during Late Mississippian time. Initial movement of the Nemaha anticline may have taken place as early as Kinderhook time (Lee, 1943, p. 115), and minor anticlinal folding in the areas of the Central Kansas uplift and Cambridge arch took place before and during the Mississippian Period (Lee, 1953, p. 20; Goebel, 1966, p. 46). Areas now included in the Forest City basin, Cherokee basin, Salina basin, and the Hugoton embayment of the Anadarko basin (fig. 19) were synclinal structures in Mississippian time (Lee, 1943, p. 120; Goebel, 1966, p. 142). These structures were rejuvenated and strongly folded during Late Mississippian time, and many smaller anticlines and synclines were formed. The present structural framework of Kansas was well established before transgression of Pennsylvanian seas.

**INTERVAL A**

**FORMATIONS INCLUDED**

Interval A is present in western Kansas only, in the subsurface (pl. 3A). There the *Kearny Formation* lies between the Mississippian System and the Atoka Series (table 20; pl. 13). The *Kearny* is assigned a Morrow age on the basis of several species of *Millerella* (Thompson, 1944).

The type well of the *Kearny Formation* is in the Stanolind Oil and Gas Company No. 1 Patterson well, in Kearny County, Kans. (fig. 21). This formation consists chiefly of interbedded clayey, sandy, and calcareous mudstone in most places, but limestone and sandstone are predominant locally. The formation consists of two unnamed members; the lower is thinner than the upper and is overlapped toward the eastern margin of the Hugoton embayment (fig. 19).

**UPPER BOUNDARY OF INTERVAL A**

Throughout western Kansas the boundary between intervals A and B generally is placed at the contact of clayey mudstone and sandstone of the *Kearny Formation* with overlying limestone (fig. 22). In the deep parts of the Hugoton embayment, clayey mudstone of the uppermost parts of interval A is gray, greenish gray, black, and, locally, variegated. Coal beds are present at or a short distance below the boundary. Near the eastern margin of the interval the clayey mudstone commonly is variegated and contains a relatively large proportion of sandstone.

In most of western Kansas the lowermost beds of interval B are gray or brown dense finely crystalline chert-bearing limestone; locally the limestone is clayey or sandy. At some places the basal rocks of interval B consist of mudstone or of sandstone, as, for example,
where the Patterson sand lies above the Kearny Formation in the type well, Stanolind No. 1 Patterson, Kearny County (fig. 21). The occurrence of mudstone or sandstone above as well as below the upper boundary of interval A is more common in northwestern Kansas than in the southwest.

**THICKNESS TRENDS**

Interval A thickens irregularly westward and southwestward from a beveled edge in western Kansas (pl. 3A). It is less than 200 feet thick throughout most of its extent, but it thickens to at least 630 feet in the Hugoton embayment in southwestern Kansas. It thins to less than 500 feet farther southwestward on the flank of the Keyes dome (fig. 19). A reentrant in the eastern border of the interval in western Finney County and eastern Kearny County (fig. 20) probably is the result of both depositional thinning and erosion.

**TABLE 20.** Major stratigraphic units of Kansas, as assigned to intervals A, B, C, D, E.

<table>
<thead>
<tr>
<th>Series</th>
<th>Interval</th>
<th>Group</th>
<th>Formation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wood Siding Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Root Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stotler Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pillsbury Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Zeandale Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Willard Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Emporia Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Auburn Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ben Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Scranton Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Howard Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Severy Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Topka Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calhoun Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Deer Creek Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Terunesh Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lecompton Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kawwaka Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Oread Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lawrence Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stranger Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stanton Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Plattsburg Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vilas Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bonner Springs Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wyandotte Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lane Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Iola Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Charute Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Drum Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cherryvale Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dennis Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Galesburg Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Swope Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ladore Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hertha Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tuckett Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Checkerboard Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seminole Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Holdenville Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lenapah Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Nowata Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Altamont Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bandera Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pawnee Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Labette Shale</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fort Scott Limestone</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cabaniss Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Krebs Formation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Unnamed rocks</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Atoka B</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Morrow A</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Chairman Y.A. (East) 
Cherokee Group 

---

**FIGURE 21.** Type well of Kearny Formation, Stanolind Oil and Gas Co. No. 1 Patterson well, sec. 23, T. 22 S., R. 38 W., Kearny County, Kans. (From J. D. Davies, Kansas Sample Log Service Log 9219).
LITHOFAcies TRENDS

Interval A is composed mainly of sandy, calcareous, and clayey mudstone (pl. 3B). Sandstone and clayey sandstone are dominant in a few areas near the eastern limit of the interval and grade generally westward and southward into clayey and calcareous mudstone. Limestone is dominant in only a small area in Stevens and Seward Counties near the Oklahoma border.

Clayey mudstone of the lower member of the Kearny Formation is dark gray to black and thinly laminated or fissile. Sandstone, mostly lenticular, ranges from white to light gray and from fine to medium grained, and it contains granular glauconite in many places. Limestone is tan to light brown and arenaceous, and it contains crinoids, brachiopods, and bryozoans.

Clayey mudstone of the upper member typically is dark gray or black, locally variegated, and, in the upper part of the member, greenish to light gray in places. Sandstone of the upper member is mainly fine grained; limestone is crystalline and arenaceous in the lower part but dense and argillaceous in the uppermost parts. The upper member contains marine invertebrate fossils and, in its uppermost part, numerous thin coal beds.

Sources and Environments of Deposition

During much of the time of interval A, the Hugoton embayment in Kansas was a shallow sea that also covered the region of the adjacent Anadarko basin of Oklahoma and Texas. The shoreline at times oscillated across the area, as shown by interbedding of limestone, sandstone, and coal within the interval. Extensive units of silty and sandy rock and of clastic carbonate rocks (McManus, 1959, p. 60-77) indicate that from time to time the sea maintained moderately high levels of energy.

Variegated clayey mudstone along the east margin of the interval, a local preponderance of sandstone, and the abundance of frosted sand grains suggest that shorelines were not far east of the present east boundary of interval A. Large amounts of sandy, clayey mudstone in northwestern Kansas indicate that the embayment may have extended only a short distance northward into northwestern Kansas and northeastern Colorado. Sandy clayey mudstone in western Morton County (pl. 3B; fig. 18) and minor amounts of arkosic sandstone in Morton, Hamilton, Stanton, and Seward Counties imply that coarse detritus was brought into the Hugoton embayment of Kansas from areas of exposed basement rock in Colorado.

In western Kansas the sea was bounded on the east by a land area of low relief that included the Pratt anticline, the Central Kansas uplift, and the Cambridge arch (fig. 19). During early stages of transgression, detrital materials were supplied from the west flank of this land area and from land at the site of the Keys dome (McManus, 1959, p. 134); also, residual materials on the pre-Pennsylvanian surface were reworked. Depositional landforms during early stages of accumulation of interval A, according to McManus (1959, p. 134-135), included alluvial channels, deltas, offshore bars, and mud-floored lagoons. These environments were succeeded gradually and intermittently by open-marine conditions, as shown by an increased percentage of limestone (McManus, 1959, p. 132) in the upper part of the lower member of the Kearny.

The marine fauna of the interval, although not abundant, includes brachiopods, horn corals, bryozoans, gastropods, crinoids, pelecypods, foraminifers, and ostracodes (McManus, 1959). These forms are mainly indicative of littoral and shallow-water environments. In middle interval A time, the eastern shoreline may have reached its maximum extent on low parts of the west flanks of the Cambridge arch, the Central Kansas uplift, and the Pratt anticline. The western shoreline extended into Colorado and the Keys dome was buried by sediment. During later stages of deposition of interval A the sea regressed southward into the Anadarko basin. The terrane of the Hugoton embayment at that time probably consisted mostly of mudflats and extensive coal basins.

Paleotectonic Implications

During the early part of Pennsylvanian time, the Hugoton embayment in southwestern Kansas was a broad southward-plunging synclinal flexure that was bounded on the east by the slightly uplifted gentle anticlinorium comprising the Pratt anticline, Central Kan-
sas uplift, and Cambridge arch. The embayment extended northwestward into Colorado and, in early Morrow time, was bounded on the southwest by the flank of the Keyes dome, which was centered in the Oklahoma Panhandle.

Progressive development of all the major flexures (fig. 19) that were present in Kansas during the latter part of the Mississippian Period may have accompanied subsidence of the Anadarko basin in Oklahoma during latest Mississippian and Early Pennsylvanian time. The Hugoton embayment probably subsided slowly, and the regional anticlinal structures that bounded it presumably rose slightly. Maximum subsidence of the embayment probably did not exceed 750 feet.

The Kansas region became slightly emergent after deposition of interval A, during which time the easternmost boundary of the interval was eroded a short distance westward, and small amounts of material probably were eroded from the top of the interval even in central parts of the Hugoton embayment. Absence of the interval in western Finney County and eastern Kearny County is considered to be evidence of an active anticlinal fold in that area during and after accumulation of interval A. Sediments of interval A were removed from the crest of the fold during the post-interval A emergence of the Hugoton embayment.

**INTERVAL B FORMATIONS INCLUDED**

Interval B is present in the subsurface of both western and eastern Kansas (fig. 23). In western Kansas, interval B includes rocks of probable Atoka age between the Kearny Formation (interval A) and rocks of the Des Moines Series (interval C). Only a limited amount of research has been done on the paleontology of this interval in western Kansas, and therefore its age is not well established.

In eastern Kansas interval B lies upon the Mississippian and consists of rocks of probable Atoka age and rocks that have been classified in Kansas as earliest Des Moines. The rocks of interval B that are assigned a Des Moines age are older than the Warner Sandstone Member of the Krebs Formation, Cherokee Group (fig. 24). Most of interval B of eastern Kansas may be considerably older than the Krebs Formation of northeastern Oklahoma (H. R. Wanless, written commun., 1967) and probably is of Atoka age. The paleontology of Pennsylvanian rocks in deeper parts of the Forest City and Cherokee basins of Kansas (fig. 19) has not been studied in detail, but regional stratigraphic relations between this part of the section and Burgner, Riverton, and McLouth Formations (pl. 13) of Missouri (Searight and Howe, 1961, p. 79-81) support a tentative assignment of these rocks to interval B and to the Atoka and the lowermost part of the Des Moines.

**UPPER BOUNDARY OF INTERVAL B**

Near the east margin of interval B in western Kansas (fig. 23), the boundary between intervals B and C generally is placed at the upper surface of clayey mudstone that underlies limestone. This contact is believed to be near enough to the Atoka-Des Moines boundary to be dependable for the purposes of this publication. Within a band extending from Scott and Wichita Counties southeastward, clayey mudstone forming the uppermost part of interval B commonly is medium gray to dark gray or black but locally is brown or red brown. At some places the mudstone is limy; elsewhere it contains considerable amounts of glauconitic sandstone. North of Scott and Wichita Counties (figs. 18, 23) clayey mudstones of uppermost interval B are mostly variegated, or they are brown and sandy.

Limestones of lowermost interval C are chiefly gray to brown, aphanitic to finely crystalline, and sandy. They are interbedded with dark-gray clayey mudstone and are locally chert bearing. In parts of Thomas and Rawlins Counties in northwest Kansas and locally elsewhere along the margins of interval B, the contact of intervals B and C lies within the complex of detritus known as the Pennsylvanian basal conglomerate. This unit probably is post-Meramec to early Des Moines in age. In this terrane the boundary is difficult to determine because rocks both above and below generally include variegated and greenish-gray clayey mudstone and white to variegated fine-grained sandstone.
Conglomerate is present at some localities in uppermost interval B.

In the central part of the Hugoton embayment, the boundary of intervals B and C can be traced into a sequence of brown to gray cherty aphanitic to finely crystalline limestone interbedded with dark-gray clayey mudstone (fig. 23, area X). The contact seemingly is conformable, and beds above and below it are not markedly different in lithology.

Throughout most of eastern Kansas the boundary of intervals B and C is placed at the base of a sandstone unit in the general stratigraphic position of the Warner Sandstone Member of the Krebs Formation (fig. 24; table 20). Where sandstone is not present in the lower part of the Cherokee Group, as in parts of Elk, Wilson, Woodson, Greenwood, and Butler Counties and in Wabaunsee, northwestern Lyon, and eastern Morris Counties, the boundary is the contact of dark-gray clayey mudstone with overlying varicolored sandy, clayey mudstone or gray limestone below the Riverton coal bed.

Clayey mudstones near the upper boundary of interval B in eastern Kansas are gray to black or, locally, green, brown, or variegated. Sandstone near the base of interval C is predominantly white to gray, fine to medium grained, and, in a few places, conglomeratic.

**THICKNESS TRENDS**

Interval B in the Hugoton embayment thickens westward from an irregular beveled edge to at least 375 feet in northwestern Stanton County (pl. 64; fig. 20). It thickens southward into the Anadarko basin of northwestern Oklahoma and Texas, but marked thinning occurs in the deeper part of the Hugoton embayment in Morton, Stevens, and Grant Counties. Throughout most of the northwestern six counties of Kansas, the interval is less than 100 feet thick.

In eastern Kansas, interval B is thickest in the west-central part of the Forest City basin, where it is more than 450 feet thick in Jackson County. The interval thins abruptly westward to the zero line along the east flank of the Nemaha anticline. Thin lobes of interval B extend southward from the Forest City basin in Chase, Butler, Greenwood, and Elk Counties on the west and in Cherokee and Crawford Counties on the east.

The interval is uncommonly thick near the junction of Greenwood, Butler, and Chase Counties and in a small area in Anderson County. The eastern featheredge of interval B is within Kansas in eastern Johnson, Miami, and Linn Counties but extends southeastward into Missouri and northeastward through Missouri into southwestern Iowa.

**LITHOFACIES TRENDS**

Interval B of western Kansas consists mainly of thick beds of gray or brown cherty and locally sandy limestone, alternating with thinner beds of dark clayey mudstone. Near the east margin of the interval, variegated or greenish-gray clayey mudstone exceeds limestone, and locally the interval is comprised chiefly of white or gray glauconitic sandstone interbedded with red and green clayey mudstone.

Interval B as a whole in westernmost Kansas is mainly clayey limestone and limy, clayey mudstone (pl. 6B); a few small areas of clayey sandstone or of limestone with less than 20 percent detrital material are present in northwestern Kansas. A facies of clayey mudstone and sandy clayey mudstone borders the interval on the east from Thomas County southward to Finney County. From Haskell County southeastward through Clark County to Oklahoma, interval B is mainly clayey limestone.

Interval B of southeastern Kansas is composed mostly of dark-gray to black clayey mudstone that locally is sandy or limy. On the southern part of the east flank of the Nemaha anticline, variegated and green or gray sandy, clayey mudstone is included locally; also, siderite concretions and a seemingly greater amount of coal than
west of the anticline occur here. In northeastern Kansas the entire interval is mostly sandy, clayey mudstone. Rocks included in interval B of eastern Kansas have been studied closely only in a few places (see, for example, Lee, 1943, and Lee and Payne, 1944). The kinds of rocks present include the following: A basal sand and conglomeratic sandstone that is widespread on top of eroded Mississippian rocks; thick and extensive units of clayey mudstone; numerous lenticular sandstones; some marine limestone that contains crinoid and brachiopod remains; black shale that locally contains linguloid brachiopod shells (Lee and Payne, 1944, p. 120-121); fairly common coal beds and coaly mudstone interbedded with thin beds of sandstone and fossiliferous limestone (Lee, 1943, p. 84; Lee and Payne, 1944, p. 101); relatively abundant clay ironstone; and small amounts of sideritic limestone (Lee and Payne, 1944, p. 101).

The lowermost part of interval B lies on eroded Mississippian rocks and at many places contains large amounts of sandstone, cherty sandstone, or conglomerate in units a few feet to 30 feet thick or more. At a few places in the Forest City basin, near the Nemaha anticline, interval B contains beds of conglomerate and of slightly arkosic sandstone.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

The Hugoton embayment of western Kansas was a shallow open sea north of the Anadarko basin during accumulation of most of interval B. The shoreline periodically retreated southward as indicated by terrigenous detrital units, including sparse amounts of sandstone, that are interbedded with limestone, which was mainly deposited in areas remote from shore. The sea probably maintained only moderate to low levels of energy, inasmuch as coarse-grained detrital and bioclastic carbonate rocks are not widespread, and oolitic limestone is sparse.

The present margin of the interval from Finney County northward was probably only a short distance northwest of the shore during deposition of much of interval B. The preponderance of detrital rocks in this region, the variegated coloring of mudstones, and the local occurrence of frosted sand grains and of conglomeratic sandstones support this conclusion. Waters of the Hugoton embayment, therefore, probably extended only a short distance into southwestern Nebraska. From Finney County southward into Oklahoma, clayey limestone is the principal lithology, even to the featheredge of the interval at most places. In this region shorelines probably were a few tens of miles east of the present limits of interval B. Small amounts of arkosic sandstone in the interval at a few places in the westernmost counties of Kansas indicate that moderately coarse detrital material was transported eastward from exposed granitic rocks in Colorado.

The alternations of rock types found in interval B in eastern Kansas indicate that environments of deposition in the Forest City basin during accumulation of interval B ranged from terrestrial and marginal marine to shallow-water marine. Terrestrial conditions prevailed around the south and west margins of the basin, but the carbonate rocks in northwestern Lyon County and southeastern Morris County suggest that shallow marine conditions persisted in that area. Shorelines probably were only a few miles beyond present limits of the interval, even during periods of maximum inundation. Interbedding of sandstone, mudstone, limestone, and coal is evidence of frequent and widespread variations in depth of water and repeated shifting from shallow-marine to swamp or other terrestrial environments.

Although accumulation of sediment exceeded 300 feet at some places, the sea probably never was as deep as 300 feet. The water apparently circulated freely, was of normal salinity, but mostly was turbid with suspended clay-sized detritus. The shore probably was bordered by lagoons of brackish water, and onshore areas were characterized by deltas and alluvial plains, lakes, and densely vegetated swamps. Water agitation probably ranged from low most of the time, as indicated by abundant thick units of clayey mudstone, to moderate, as suggested by the small amounts of coarse-grained sandstone and by local intraformational breccia (Lee and Payne, 1944, p. 102).

Relief in the land areas probably was very low except on the northern part of the Nemaha anticline, where Precambrian rocks were exposed and where some hills may have stood at heights of a few hundreds of feet. Coal deposits suggest that the prevailing climate was warm and humid.

Most of the sand of the sandy clayey mudstone facies of interval B in eastern Kansas probably came from lower Paleozoic and Precambrian rocks that were exposed in land areas on the northern part of the Nemaha anticline of Kansas and Nebraska (pl. 2). Much clay and minor amounts of sand were transported eastward to the Forest City and Cherokee basins from the part of the Nemaha anticline south of Geary and Wabaunsee Counties. Erosion of a moderately large area of low-lying Mississippian carbonate-rock terrane in southeastern Kansas, and exposed Mississippian and Ordovician-Cambrian carbonate rocks of the Ozark region apparently provided large amounts of clay.

**PALEOTECTONIC IMPLICATIONS**

The Hugoton embayment was slightly emergent at the end of interval A time but gradually subsided during deposition of interval B. Deposits of interval A were overlapped from the west. Subsidence probably was interrupted by recurrent uplift along the major anticlinal
flexures that bounded the embayment in Kansas, including the Pratt antcline, Central Kansas uplift, and Cambridge arch (fig. 19). Thinning of interval B in Morton, Hamilton, Stevens, Grant, and Haskell Counties indicates that a small arch extended northeastward from the Keyes dome of the Oklahoma Panhandle at least to Finney County. Cumulative subsidence within the embayment was at least 400 feet.

Following accumulation of interval B, emergence and moderate erosion probably occurred near the east margin of the Hugoton embayment. Downwarping in the interior of the embayment was continuous from interval B into interval C.

During the time between the end of deposition of Mississippian rocks and the beginning of deposition of interval B, the region that includes the Forest City basin, the northern part of the Cherokee basin, and the Nemaha antcline was emergent and was eroded to very low relief; slow downwarping of depositional areas was then renewed to bring them intermittently below sea level.

The most pronounced deformation of rocks in Kansas consisted of faulting and monoclinal folding along the east flank of the Nemaha antcline (Lee, 1943, p. 116), probably along its full length in Kansas (Jewett and Abernathy, 1945, pl. 1-4; Merriam, 1963, fig. 146, p. 252-254). The Forest City and Cherokee basins seem to have subsided differentially, producing the Bourbon arch, a broad feature with little closure and low structural relief. This poorly defined arch trended northwestward from Bourbon County through Allen, Anderson, Coffey, and Lyon Counties to Chase County (figs. 19, 20). Its presence is indicated by an area of thinning of interval B (pl. 6A). Northward the interval thickens more than 450 feet in Jackson County, and southward it thickens to more than 200 feet in northern Butler, southern Chase, and northern Greenwood Counties.

The absence of interval B in the southeasternmost counties of Kansas (pl. 6A) and in parts of northeastern Oklahoma probably is due to nondeposition. Atoka seas probably transgressed northward from Oklahoma through southwestern Missouri, thence northwestward to the Forest City basin and southward across the Bourbon arch into the Cherokee basin of Kansas. The presumed land area in southeastern Kansas probably was a structural platform that extended westward to the Nemaha antcline and southward into central Oklahoma.

Little evidence exists to show that interval B of eastern Kansas was elevated enough for marked dissection by erosion before deposition of interval C. Rather, the regional stratigraphy indicates continued sedimentation, alternating between marine and nonmarine environments. The Nemaha antcline, Forest City basin,
Interval C thickens from zero on the Nemaha anticline near the Kansas-Nebraska border to more than 400 feet in the Salina basin to the west and to as much as 460 feet in the Sedgwick basin to the southwest. The interval thins to as little as 250 feet at places between the Salina and Sedgwick basins.

Interval C thickens westward and southwestward from central Kansas (Cambridge arch, Central Kansas uplift, Pratt anticline) into the Hugoton embayment and is about 800 feet thick on the axis of the embayment near the Kansas-Oklahoma border. Local thinning occurs on the northeast flank of the Keyes dome in the southwest corner of Kansas, and regional thinning extends northward from the Keyes dome through about two-thirds of western Kansas.

**LITHOFACIES TRENDS**

Interval C is composed mainly of interbedded clayey mudstone, sandstone, limestone, and coal. The Cherokee Group contains much clayey mudstone, appreciable amounts of sandstone, at least 13 thin but widespread coal beds, and a few thin limestone beds, some of which are lenticular. The Marmaton Group, in contrast, contains four widespread limestone and three mudstone formations; in addition, it includes relatively minor amounts of sandstone and coal. The Pleasanton Group is chiefly clayey mudstone but contains sandstone, one widespread limestone unit, and small amounts of coal.

Several limestone units, some coal beds, and a few sandstone beds of interval C are widely distributed at the surface in eastern Kansas. The addition of limestone and mudstone units in the subsurface in the eastern and central basin areas and a decrease in the number of units on the intervening anticlinal areas (fig. 26) make difficult the tracing in the subsurface of more than a few of the outcropping beds. The most easily recognizable units are limestone, and only the most distinct and persistent can be reliably correlated between eastern and western Kansas.

The Forest City, Cherokee, Salina, and Sedgwick basins contain many limestone units and large numbers of intervening detrital units, but fewer units lap onto the Cambridge arch-Central Kansas uplift-Pratt anticline area and the Nemaha anticline. Changes in facies of interval C westward, from clayey mudstone to dominantly clayey limestone (pl. 7A), are due, in general, to moderate thickening of large numbers of limestone units relative to detrital units, rather than thickening of a few limestone units.

Interval C is mainly clayey to sandy mudstone in the Cherokee and Forest City basins, limy mudstone on anticlinal terraces and in intervening basins of central, southern, and most of northern Kansas, and clayey limestone in the Hugoton embayment in southwestern Kansas (pl.

**THICKNESS TRENDS**

In eastern Kansas, interval C thickens westward from an erosional edge in Cherokee County to more than 800 feet (pl. 7A) in the deepest part of the Cherokee basin in Labette, southern Montgomery, and southeastern Chautauqua Counties, where the section is complete. The interval is more than 600 feet thick in the deepest parts of the Forest City basin in Brown, Atchison, northwestern Jefferson, and eastern Jackson Counties.

A short distance westward from the Forest City basin, interval C thins to a featheredge on the crest of the Nemaha anticline in northern Nemaha and northeastern Marshall Counties. In southeastern Kansas the interval thins westward from more than 800 feet in the Cherokee basin to less than 100 feet above local anticlines in the southern part of the Nemaha anticline (pl. 7A). Interval C thickens to as much as 320 feet along the strike of the plunging Nemaha anticline from Nemaha County southwestward to the Oklahoma line.

Rocks of interval C are absent from the crest of the Central Kansas uplift in Rush and Barton Counties. They are less than 100 feet thick throughout a large area of central Kansas and from there northwestward along the crest of the Cambridge arch into Nebraska. The interval also is thinner than 100 feet over part of the Pratt anticline south of the Central Kansas uplift (fig. 19; pl. 7A).

Figure 25.—Distribution of types of lithology at upper boundary of interval C in Kansas: Area W, Gray to black locally carboniferous or sandy clayey mudstone (interval C) overlain by chiefly buff to gray or white aphanitic or finely crystalline limestone (interval D). Area X, Red and gray, locally variegated, locally grayish-green, clayey or sandy mudstone (interval C) overlain by gray, buff, or locally white aphanitic to finely crystalline locally chert-bearing limestone (interval D). Area Y, Gray, dark-gray, black, or, rarely, variegated clayey mudstone (interval C) overlain by gray to buff mainly chert-bearing aphanitic or finely crystalline limestone (interval D). Area Z, Gray mudstone (interval C) overlain by buff to gray locally chert-bearing limestone (interval D).
7B). Many local modifications of facies occur within each of the three major areas. Among the more conspicuous modifications are a change from clayey limestone to limy mudstone in the Hugoton embayment of southwesternmost Kansas and in the vicinity of the Keyes dome, and the extension of a tongue of clayey limestone of the Hugoton embayment into the dominantly mudstone area of the Sedgwick basin.

The mudstones in eastern, central, and northwestern Kansas are mainly gray, black, red, green, or variegated. In southwestern Kansas they are mostly gray and red. In that area sandy mudstone locally is common. Limestones throughout the State are aphanitic to finely crystalline, and are brown to gray. In the central and western parts of the State they also are locally cherty and oolitic. Conglomerates are present at the base of the interval in some areas, and in eastern Kansas, red to buff sandstone lenses are interbedded with the dominant mudstone. Siderite concretions and coal also occur in the lower beds in eastern Kansas.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

As already indicated, marine deposition was continuous between intervals B and C in the Hugoton embayment of southwestern Kansas and in the Forest City and Cherokee basins in eastern Kansas. The sea gradually spread outward from these areas, across the central part of the State, interrupted only by brief episodes of partial regression. The progressive encroachment of the sea is indicated by overlapping of depositional units from the enlarging basins onto the adjacent land areas. Periodic retreat of the sea is indicated by interbedding of fluvial and marginal-marine rocks with rocks deposited in open-marine environments. As transgression proceeded, all of Kansas became a region of comparatively shallow seas filling basins and covering shelves, separated only by greatly diminished positive elements.

In eastern Kansas, sediment of interval C in the Forest City and Cherokee basins was derived from shrinking land areas along the Nemaha anticline where older Paleozoic and Precambrian rocks were exposed and from land northeast and east of Kansas. Because clayey mud was the most abundant sediment deposited in interval C, the shallow interior seaway is believed to have maintained only low to moderate levels of energy and to have been prevalently turbid; the few sandstones are mostly clayey and fine to medium grained (Lee, 1943, p. 81-84; Howe, 1956). Along the east flank of the northern part of the Nemaha anticline, small amounts of conglomerate and arkosic sandstone suggest occasional strong currents in that area. Locally, conglomeratic or brecciated limestones also indicate wave or current action (Howe, 1956, p. 24).

Depth of water in much of eastern Kansas, as indicated by fusulinids (McCrone, 1964, p. 275), was probably about 50-75 feet, and at no time during interval C did it greatly exceed 100 feet. Relief of coastal land areas may have ranged from a few feet to a few tens of feet. Relief on the surface of unconformity at the Missouri-Des Moines contact, where topographic features can best be measured, is about 50 feet (Lee, 1943, p. 88).

Where interval C crops out in the Cherokee basin of southeastern Kansas, the Cherokee Group at the base contains about 13 cycloths (fig. 27). The cycloths represent alternations of marine and nonmarine conditions near the east margin of the basin (Abernathy, 1937, p. 18-23; Moore, 1960; Howe, 1956). Generally the cycloths comprise five units, starting with dark mudstone and irregularly bedded limestone at the base, overlain in turn by lighter gray mudstone, sandstone or limestone, underclay, and, at the top, coal.

The basal dark mudstone and irregularly bedded limestone of the Cherokee cycloths contain a variety
of marine fossils, mostly brachiopods, corals, conodonts, scattered fish remains, and macerated plant material (Howe, 1956; Moore, 1950, p. 8; Williams, 1938). The mudstone and limestone are interpreted as representing marginal-marine to shallow-marine conditions and the beginning stages of transgression of the sea. Gray even-ly bedded clayey sparsely fossiliferous mudstone overlies the basal dark mudstone and forms the thickest unit within each cyclothem (Howe, 1956, p. 24). Probably this mudstone was deposited at greater depth and distance from shore than was the basal mudstone.

Sandstone or limestone, or a mixture of these rock types, lies above the gray mudstone and beneath underclay. The sandstone normally ranges from massive to thin bedded and is fine grained. Both sandstone and limestone contain marine fossils. Good sorting of the sandstone and local conglomeratic and brecciated limestone is considered evidence of shallow marine to lagoonal conditions (Howe, 1956, p. 24) during regression of the seas. The underclay contains parts of fossil root systems, stigmarian remains, and carbonized plant material (Howe, 1956) which, together with the overlying coal beds, are evidence of nonmarine depositional environments.

The position of the sandstone and clayey mudstone beds in individual cyclothem sequences differs somewhat from place to place, and the environments of deposition did not everywhere follow the same pattern in detail. Not all the sandstone beds are of marine origin; some locally contain stigmarian roots and root casts, other plant material, and the tracks of land animals (Howe, 1956, p. 33, 57; Moore, 1950, p. 8).

Alternating shallow-marine and nonmarine conditions prevailed not only in the Cherokee outcrop area of far eastern Kansas but also in most of the Forest City and Cherokee basins and in the bordering Nemaha region to the west. Thin coal beds are numerous in Cherokee rocks in the interior and western parts of the Forest City and Cherokee basins (Johnson and Adkison, 1967, p. 9; Johnson and Wagner, 1967, p. 132; Bass, 1936; Mudge and Burton, 1959, p. 122). Limestones containing marine fossils and shoestring sands that probably were offshore bars deposited on the north and west margins of shallow bays in the Cherokee sea are also present (Bass, 1936, p. 26).

In eastern Kansas, east of the Nemaha anticline, the Marmaton Group consists of alternating sequences of clayey mudstone, thin units of limestone and channel sandstone, and thin beds of coal, which were deposited in alternating marine and nonmarine environments similar to those of the underlying Cherokee Group (Jewett, 1945; Lee, 1943). Corals, algae, and a few fusulinids are contained in some limestone beds. The coral genus Chaetetes is especially abundant and locally produces marked increases in thickness of the Fort Scott.
Limestone (table 20) (Moore, 1949a). Clayey mudstone beds, some of which are black and carbonaceous, contain mainly marine fossils, including brachiopods, bryozoans, crinoids, mollusks, and corals (Jewett, 1945, p. 13). In the subsurface of the Forest City basin, the Marmaton has progressively greater proportions of detrital rocks where it overlaps the Cherokee Group westward onto the Nemaha anticline. Coal beds are present in the Marmaton at least as far west as Shawnee County on the eastern flank of the Nemaha anticline (Johnson and Adkison, 1967, p. 10; Johnson and Wagner, 1967, p. 133).

In the Pleasanton Group of eastern Kansas a marginal-marine environment of deposition is indicated by clayey mudstone that contains land plants, brachiopods, crinoids, bryozoans, and pelecypods (Jewett and others, 1965). The Checkerboard Limestone (table 20), which consists of fine-grained nodular or crossbedded coquinitid limestone, contains arenaceous foraminifers and other marine fossils. Associated sandstone locally contains marine fossils, land plants, or a mixture of both. In the interior of the Forest City basin and in areas bordering the Nemaha anticline, the Pleasanton consists chiefly of clayey mudstone, siltstone, and fine-grained sandstone and minor amounts of limestone (Johnson and Adkison, 1967, p. 10; Johnson and Wagner, 1967, p. 133; Lee, 1956, p. 91, 92).

West of the Nemaha anticline, cyclothsems as developed in eastern Kansas are not present in interval C, coal beds are few and seemingly discontinuous, and sandstone is present only in minor amounts (Adkison, 1963). Clayey mudstone beds in the Cherokee Group where it overlaps from the Salina and Sedgwick basins onto the Central Kansas uplift are predominantly red and gray (Lee, 1956, p. 87).

Rocks of interval C west of the Nemaha anticline are believed to have accumulated in shallow marine, littoral, and, locally, subaerial environments, and they overlap onto the positive elements. At times, moderately high energy prevailed in the sea, as shown by an abundance of oolitic limestone. The water was turbid periodically but was seemingly clear enough at other times to allow corals and algae to flourish. The slightly emergent region in central and north-central Kansas was the source of much fine detritus, but arkosic sandstones, some of which are 15-20 feet thick, and coarse-grained quartz sandstones in southwestern and northwestern Kansas, having well-rounded frosted grains, suggest the proximity of exposed basement rocks in land areas of eastern Colorado.

PALEOTECTONIC IMPLICATIONS

The interior parts of the Forest City and Cherokee basins and the Hugoton embayment subsided without appreciable interruption from interval B into interval C, and downwarping continued intermittently throughout deposition of interval C. All of Kansas except the crest of the Central Kansas uplift and the northern part of the Nemaha anticline was covered by marine waters at one time or another during accumulation of interval C. Subsidence was interrupted periodically by minor episodes of uplift centered along the trend of the Central Kansas uplift and Cambridge arch in the central and north-central parts of the State and along the Nemaha anticline farther east.

In the Hugoton embayment, thinning of interval C on the Keyes dome and along a band trending northeastward from the dome (pl. 7A) are evidence of continued slight movement of these low-amplitude folds (pl. 3A). Total subsidence in the Forest City and Cherokee basins probably was on the order of 800 feet (pl. 7A). Although rocks of interval C are more than 400 feet thick in the Salina and Sedgwick basins, these regions apparently had been downwarped to some extent before interval C time; accordingly, total subsidence of these basins during accumulation of interval C probably was somewhat less than 400 feet.

Folding of local poorly defined anticlines and synclines accompanied uplift of the Nemaha anticline and the Central Kansas uplift (Lee, 1956, p. 149-150). The Nemaha anticline was faulted at many places along its east flank; both normal and reverse faults seem to be present (Merriam, 1963, p. 182).

INTERVAL D FORMATIONS INCLUDED

Interval D is exposed in eastern Kansas and is present in the subsurface throughout the State except in the southeasternmost few counties (pl. 8A; fig. 28). The interval consists of the Kansas City Group and the overlying Lansing Group, both of Missouri age (table 20; pl. 13).

UPPER BOUNDARY OF INTERVAL D

The upper boundary of interval D is placed at the contact of the Stanton Limestone of the Lansing Group (below) with the Stranger Formation of the Douglas Group (above); this is the Missouri-Virgil boundary in Kansas. Throughout most of Kansas, uppermost rocks of interval D are limestone, and lowermost rocks of interval E are mudstone (fig. 28).

THICKNESS TRENDS

In southeastern Kansas, interval D thickens westward from an erosional edge (pl. 8A). Within the basin area east of the Nemaha anticline, the interval increases in thickness southwestward from about 400 feet in Douglas County to more than 800 feet in Chautauqua County. In the Forest City basin, interval D thins westward from about 400 feet to less than 150 feet in Nemaha County on the northern part of the Nemaha anticline. Farther south, the anticline is only weakly
defined by thickness changes, but the Cambridge arch and the Central Kansas uplift are clearly defined. Interval D is 300 feet thick or less in most of north-central Kansas, thinning to less than 200 feet in parts of Phillips County and adjacent counties. The interval thickens slightly in the Salina basin to the east and thickens markedly, to almost 700 feet, in the Hugoton embayment of southwestern Kansas.

Interval D shows pronounced local variations in thickness across the south edge of Kansas, westward from the Sedgwick basin. Irregular lobate areas of thickening and thinning trend generally northeastward across the northwest-trending axis of the Hugoton embayment. On the flank of the Keyes dome, in the southwest corner of the State, the interval is slightly less than 400 feet.

LITHOFACIES TRENDS

Interval D is composed mainly of limestone and clayey mudstone but contains minor amounts of sandstone and, in the Kansas City Group, small amounts of coal. Fifteen formations are traceable at the surface, and most are recognizable in the subsurface; some units can be followed in the subsurface across the State (Merriam, 1963; Parkhurst, 1959). Limestone units thicken generally westward and southward in the subsurface and compose an increasing proportion of the interval; at some places they thicken abruptly into algal-mound complexes and locally are as much as four times the normal thickness (Merriam, 1963; Heckel and Cocks, 1969).

In southeastern and easternmost Kansas, interval D consists mostly of limy clayey mudstone (pl. 8B), but it is largely clayey limestone in southern Greenwood and northern Elk Counties and both clayey mudstone and sandy clayey mudstone in parts of Chautauqua, Cowley, and Sumner Counties. Most of the clayey mudstone is gray and red, but local thin beds are black and carbonaceous. Beds of buff or tan sandstone at some places are more than 10 feet thick. The limestone is mostly buff to gray, chert bearing, and oolitic.

Interval D is chiefly clayey limestone in an arcuate band that extends from the Forest City basin westward across the Nemaha anticline into the Salina basin and thence across the Cambridge arch into Colorado and southward into the Sedgwick basin (pl. 8B; fig. 19). Limestone and sandy clayey limestone are abundant at several scattered localities. Limestone extends across the Cambridge arch, mostly in Decatur, Sheridan, and Graham Counties. In extreme northwestern Kansas, interval D consists of limy mudstone.

Interval D is predominantly limestone in southwestern and central Kansas, in the region of the eastern flank of the Hugoton embayment. Facies of sandy, clayey limestone, limy mudstone, and clayey limestone border the Keyes dome on the northeast. Large areas of clayey limestone occur surrounded or partly surrounded by pure limestone in southern Finney County and vicinity, in Kiowa, Barber, and Harper Counties, and in southern Clark County.

Limestone in northeastern, central, and western Kansas is buff to gray or locally white, some is chert bearing, oolitic, or dolomitic, and most is aphanitic to finely crystalline. In the clayey limestone facies, clayey mudstone is interbedded with the limestone and is red, gray, grayish green, or black in the area of the Nemaha anticline and Forest City basin and locally contains beds of carbonaceous clayey mudstone. In the Salina basin and from there westward to Colorado, the clayey mudstone is red and gray or variegated and includes black clayey mudstone at several places. Beds of sandstone are present in the clayey mudstone across some parts of the Cambridge arch and in Cheyenne, Sherman, Wallace, and Greeley Counties in northwesternmost Kansas; clayey mudstone on the flank of the Keyes dome is locally sandy or silty and contains beds of white to gray fine-grained limy sandstone.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Sedimentation apparently was uninterrupted from interval C into interval D time in Kansas. Soon after the onset of interval D, land areas near the crests of the Nemaha anticline and Central Kansas uplift were sub-

---

1 Recent work by Stanton M. Ball (written commun., 1972) on uppermost Missouri and lowermost Virgil rocks in area U has shown that extensive sandstone strata recorded in this and other publications as lowermost Virgil are part of the uppermost Missouri. The reader should bear in mind that within area U the upper boundary of interval D probably is somewhat below the Missouri-Virgil contact.
merged, and shallow marine to paralic conditions prevailed across the State. These conditions were interrupted periodically by terrestrial environments during temporary regressions of the sea. Shallow-marine to locally terrestrial environments probably prevailed cyclically throughout most of the region during the time of this interval.

Most of the terrigenous detritus in interval D came from outside Kansas, probably from the region of the Arbuckle Mountains and Ouachita Mountains to the south, the Ozark area to the east, and, to some extent, land areas to the north and northeast (Heckel and Cocke, 1969; Crowley, 1969; Walton, 1960; Ball, 1964). A progressively greater influx of terrigenous detritus from the south into south-central Kansas through interval D time is indicated by the fact that carbonate units of the Kansas City Group persist farther southward than do those of the Lansing Group (Lukert, 1949, p. 146, cross section A-B; Merriam, 1963, p. 124).

Coarse clastic carbonate rock and abundant oolitic limestone in parts of the interval indicate some moderate- to high-energy wave and current action in marine waters in Kansas; in contrast, persistent black fissile carbonaceous and clayey mudstone beds and some nonoolitic limestones that extend far westward in Kansas suggest that many other parts of the sea were quiet. Widespread shallow-water conditions are indicated by gastropods, corals, brachiopods, fusulinids, crinoid stem fragments, and small amounts of algal-encrusted shell fragments on the Central Kansas uplift (Stafford County) and in northwestern Kansas (Rawlins County) (Harbaugh and Davie, 1964); by crinoids, fusulinids, brachiopods, brachipods, ostracodes, gastropods, and pelecypods in the Sedgwick basin (Adkison, 1963); and by fusulinids in many other places and algal and crinoidal limestone in southwestern Kansas (Morton, Seward, and Stanton Counties).

The waters ranged from turbid, as shown by the presence of clayey mudstone beds, to clear, as shown, for example, by beds rich in algae, whose habitat presumably was well lighted and reasonably free of fine suspended mud.

In the outcrop region in eastern Kansas, interval D consists of formations made up chiefly of limestone members interbedded with relatively thin clayey mudstone members, alternating with thick formations of clayey mudstone (fig. 29). Several cyclothems are recognized within this assemblage. Neither the order of occurrence nor the faunal content of the beds is the same among different cyclothems. This complexity seems to be due to partial development of simple cyclothems, each of which is different from the others and all of which are contained in a single large complex cyclothem (fig. 30), termed a megacyclothem (Moore, 1936, p. 29; 1950, p. 11).
Three types of partially developed cyclothems can be distinguished in megacyclothems of interval D. The first type (type A, fig. 30) generally consists of sandstone overlain by beds of shaly, clayey mudstone or, locally, by coal and by limestone that contains mollusks, fusulinids, or brachiopods. The second type (type B, fig. 30) comprises (1) dark carbonaceous fissile clayey mudstone overlain by light-colored clayey mudstone, both of which contain thinly scattered conodonts, brachiopods, pelecypods, or phosphatic concretions with marine fossils; overlain, in turn, by (2) thin- to medium-bedded limestone in a thick sequence containing in its upper part fusulinids and oolites, nodular chert, abundant algae, and some mollusks and brachiopods; and (3) sandy, clayey, and shaly mudstone that generally is unfossiliferous. The third type (type C, fig. 30) locally includes in its lower part a sandy unit bearing plant fossils, overlain by sandy, clayey, and shaly mudstone containing mollusks, and limestone that includes in its upper part fusulinids, mollusks, and algae (Moore, 1949b, 1950).

In eastern Kansas all the limestone formations within the cyclothem sequences of interval D locally contain thick and extensive algal mounds. These reeflike masses have been subjects of much study in recent years (Wilson, 1957; Davis, 1959; Harbaugh, 1959, 1960, 1962, 1964; Wray, 1964, 1965; Harbaugh and others, 1965; Crowley, 1969; Frost, 1968; Heckel and Cocke, 1969). In the central parts of the mounds, leaflike algae are especially abundant. The mounds are chiefly massive to thick-bedded calcilutite and carbonate spar (Heckel and Cocke, 1969, p. 1061); in some of them the core areas contain invertebrate fossils, including brachiopods, bryozoans, crinoid remains, horn corals, fusulinids, encrusting algae, gastropods, tubiform foraminifers, or sponges (Crowley, 1969; Wray, 1964; Davis, 1959; Harbaugh, 1962).

Crossbedded calcarenitic limestone, including both fragmental skeletal material and thick oolitic deposits, commonly overlies the central parts as well as the flanks of the mounds. Some of these mounds are exceedingly thick in comparison with the units within which they are developed, and they cover large areas. For example, a large algal mound in the Spring Hill Limestone Member of the Plattsburg Limestone (table 20) in Wilson County is more than 80 feet thick, whereas the Spring Hill normally is about 3 feet thick. This mound complex covers an area of about 140 square miles (Harbaugh, 1962, p. 43).

Algal mounds within interval D mostly occur near the southern extremities of limestone units, at relatively short distances from places where carbonate beds of interval D grade into terrigenous rocks. In some areas, mounds occur in several different units of interval D. For example, in southeastern Kansas, algal mounds are in the Hertha Limestone (table 20) and also in the Bethany Falls Limestone Member of the Swope Limestone and in the Winterset Limestone Member of the Denni Limestone (Heckel and Cocke, 1969, p. 1063).

Algal mounds seem to have formed mostly on high areas on the floors of shallow seas, including mud bars (Harbaugh, 1964) or, as in the Wyandotte Limestone, on submerged deltaic platforms (Crowley, 1969, p. 47). Their growth probably kept pace with subsidence (Heckel and Cocke, 1969, p. 1067). At times the upper parts of the mounds were barely awash, perhaps even slightly emergent (Harbaugh, 1960, p. 233), and strong
currents deposited calcarenites as spits and bars; the calcilutites probably formed in quiet lagoons.

The algal-mound limestones and associated rocks of interval D grade northward and westward in eastern Kansas into rocks deposited in open-marine environments and southward and southwestward in south-central Kansas and Oklahoma into rocks deposited on or near shore. The environments in which these facies accumulated existed side by side but shifted position continuously as a result of periodic advances and retreats of the sea.

Although most beds of interval D contain abundant marine fossils, a shore environment is recorded at a locality in central eastern Kansas (Anderson County) where a mixed assemblage of marine invertebrates, land plants, coelacanth fishes, terrestrial arthropods, and reptiles has been recovered from the Rock Lake Shale Member of the Stanton Limestone (fig. 29) (Moore, 1936; Peabody, 1952, 1957, 1958; Eaton and Stewart, 1960; Cridland and others, 1963). These fossils are judged to have accumulated in a lagoon near the shore (Peabody, 1952; Eaton and Stewart, 1960; Cridland and others, 1963; Moore, 1964; Ball, 1964, p. 231-232).

Tracks probably of amphibians have been reported from the Kansas City Group at Kansas City, Mo. (Branson and Mehl, 1932, p. 391-393). In addition, a few beds of coal, sandstone, and clayey mudstone that contain plant fossils as well as other features indicating a littoral environment (Gentile, 1969; Ball, 1964) suggest that periodically during the time of interval D eastern Kansas was a region of low-lying swampy coastal plains crossed by sluggish streams.

Marine beds containing algal mounds intertongue with terrestrial beds and extend an unknown but considerable distance west of the outcrop of interval D in Kansas, perhaps as far west as Butler and Cowley Counties on the west side of the Cherokee basin (Merriam, 1963, p. 124; Ball, 1964, p. 196-197). One coal bed occurs in Butler County (Adkison, 1963, p. 10), and coaly, clayey mudstone occurs in Shawnee County (Johnson and Wagner, 1967, p. 133-134). Farther west, the interval was largely or entirely deposited in shallow-marine environments.

The southward change in interval D from predominantly limestone and clayey limestone in central Kansas to predominantly limy mudstone, clayey mudstone, and sandy mudstone in south-central Kansas in the general area of Wilson, Montgomery, Chautauqua, Elk, Cowley, and Sumner Counties may be a combination of changes in facies and of erosion of limestone in south-central Kansas followed by deposition of sandstone and mudstone (Lukert, 1949, p. 145-151; Pate, 1959, p. 44-48; Rascoe, 1962; Merriam, 1963, p. 125; and especially Winchell, 1957; Schulte, 1958; and Ball, 1964, p. 196-200). The relative importance of the two in explaining the facies patterns is unknown.

At the end of deposition of interval D sediments, all of Kansas was submerged except perhaps parts of south-central Kansas (Montgomery, Wilson, Cowley, and Sumner Counties), where uppermost rocks of interval D may have undergone intermittent subaerial erosion.

**PALEOTECTONIC IMPLICATIONS**

The central parts of the Cherokee basin and the Hugoton embayment subsided almost continuously from the time of deposition of interval C sediments through that of interval D sediments. Total subsidence exceeded 700 feet in the central part of the Hugoton embayment in southwestern Comanche County and 900 feet in the Cherokee basin in southern Chautauqua County. However, during periods of maximal submergence, marine waters probably were only a few tens of feet deep.

Rocks of interval D are no thicker in the central parts of the Forest City and Salina basins as defined during interval C than on the flanks of these basins, showing that no differential subsidence occurred during interval D time. During the deposition of interval D, the area of these basins was an intermittently subsiding shelf. Broad structural embayments extended from the deep part of the Cherokee basin northeastward into the southern part of the Forest City basin and northwestward across the southern part of the Nemaha anticline into the Sedgwick basin.

The northern part of the Nemaha anticline was uplifted moderately during interval D time, but only slight uplift took place within the southern two-thirds of the anticline. The Cambridge arch and Central Kansas uplift continued to be differentially elevated and together composed a large anticlinorium of very low amplitude that occupied most of northwestern Kansas. The crest of the structure was located on the Cambridge arch in Norton, Graham, and western Phillips and Rooks Counties, where interval D is thinner than 200 feet. Thinning of interval D in Barber, Harper, and Sedgwick Counties indicates that a large northeast-trending anticline extended from the northeast flank of the Hugoton embayment to the Sedgwick basin and lay directly across the southeastward-plunging part of the Central Kansas uplift.

Active folding of the Keyes dome influenced sedimentation during interval D time; the interval thins across the north and east flanks of the dome and includes sandstone and mudstone next to the dome. Several anticlinal flexures in the Hugoton embayment were uplifted slightly; most of them plunged southward and southwestward from the northeastern shelf toward the axis of the embayment and the Keyes dome.
FIGURE 31.—Generalized section of uppermost Pennsylvanian (interval E) and lowest Permian rocks in eastern Kansas. Section not drawn to scale. (From Jewett and others, 1965, pl. 1; Johnson and Wagner, 1967, pl. 3.)

INTERVAL E
FORMATIONS INCLUDED

Interval E crops out in eastern Kansas and is present in all of the State except in a few counties near the east border. The interval includes the Douglas, Shawnee, and Wabaunsee Groups, in ascending order (pl. 13); formations in these groups are listed in table 20.

UPPER BOUNDARY OF INTERVAL E

No widespread clear physical difference exists between the uppermost units of the Pennsylvanian and the lowest units of the Permian in Kansas. The two systems are conformable at most places, both at the surface and in the subsurface. In the outcrop belt, the upper boundary of interval E is placed arbitrarily at a conformable contact of the Brownville Limestone Member of the Wood Siding Formation with the overlying Permian Towle Shale Member of the Onaga Shale (fig. 31). Clayey mudstone units and channel or sheet sandstones occur both in the upper part of the Wabaunsee Group of Pennsylvanian age and in the lower part of the Admire Group of Permian age. Some channel sandstones of Permian age within or above the Towle Shale Member cut deeply into underlying strata; for example, sandstone-filled channels of the Towle Shale Member have cut as deeply into Pennsylvanian rocks as the Zeandale Limestone (table 20) (Johnson and Wagner, 1967, p. 175-180, pl. 3). At these places the contact is an unconformity at the base of the sandstone. Channel sandstones are numerous in this part of the stratigraphic section and are important to interpretation of environments of deposition; however, they are discontinuous and only local at the base of the Permian System. (For further discussion, see Mudge, 1956; Mudge and Yochelson, 1962, p. 19-21, 119-120.)

Because of the lithic similarity of uppermost Pennsylvanian and lowest Permian rocks, the lack of an exhaustive study of fossils contained in them, and the absence of a well-defined and extensive erosion surface in the stratigraphic section, the boundary between the systems was placed at several different stratigraphic positions in Kansas during the period from 1859 to 1936 (Moore, 1940, p. 300; Mudge and Yochelson, 1962, p. 118). The present boundary at the top of the Brownville Limestone Member was established by Moore (1936, p. 14) and is stratigraphically lower than earlier ones.

Difficulty in exact placement of the boundary is caused by uncertainty regarding the base of the Permian in the North American standard sequence of western Texas (Douglas, 1962, p. 120-122; Oriel and others, 1967, p. 26-27). The base of the *Pseudoschwagerina* zone has been accepted by some writers as the base of the Permian System and has been questioned by others (Douglas, 1962, p. 121). The lowest occurrence of *Pseudoschwagerina*, taken to be diagnostic of the *Pseudoschwagerina* zone, is in the Five Point Limestone Member of the Janesville Shale (fig. 31), which in Kansas is about 60 feet above the base of the Permian as defined since 1936. Finally, because distinct breaks are lacking in the evolutionary sequence of the closely studied fossils from uppermost Pennsylvanian and lowest Permian rocks of Kansas, solution of the problem remains difficult.

Faunas in the rocks at the Pennsylvanian-Permian boundary include fusulinids, corals, echinoids, crinoids, brachiopods, pelecypods, gastropods, both nautiloid and ammonoid cephalopods, and ostracodes (Douglas, 1962, p. 120-122; Duncan, 1962, p. 122-123; Gordon, 1962, p. 123, 126). Arbitrary assignment of the top of the Brownville Limestone Member as the top of the Permian System in Kansas is as logical, on the basis of present knowledge of these faunas, as assignment of any other horizon in this part of the section. It is justified mainly by practicality and longtime usage (Mudge and Yochelson, 1962, p. 116-127).

The Wood Siding Formation can be traced into the
subsurface west of the outcrop (Merriam, 1963, fig. 43) but is not sufficiently lithologically distinctive and persistent to be easily recognizable throughout central and western Kansas. Lithologic relations at the contact in different parts of Kansas are shown in figure 32.

THICKNESS TRENDS

In eastern Kansas, interval E thickens from an erosional edge westward into a trough that flanks the east edge of the Nemaha anticline. Thicknesses increase southward in the trough area from 800 feet on the north to about 1,400 feet on the south, extending from southeasternmost Nebraska to Oklahoma (pl. 9A). The interval thins markedly farther west to about 1,200 feet on the southern part of the Nemaha anticline and about 900 feet on the central part. A slight westward thinning from 800 to about 500 feet occurs over the northern end of the anticline in Nemaha, Marshall, Pottawatomie, Riley, and Geary Counties. Rocks of interval E are more than 700 feet thick in most of Jewell and Mitchell Counties in the central part of the Salina basin, and they thicken southward along the west flank of the Nemaha anticline to more than 1,500 feet in Harper and Sumner Counties and to more than 1,100 feet farther west in Clark County. The interval thins generally northwestward from Harper and Clark Counties to less than 400 feet in the northwesternmost counties of Kansas. Rocks of interval E are thinner than 600 feet in Morton County in the vicinity of the Keyes dome.

LITHOFACIES TRENDS

In eastern Kansas, interval E contains clayey mudstone, limestone, sandstone, and minor amounts of coal (pl. 9B). The Douglas Group consists of two formations made up of thick units of mudstone interlayered with sandstone and a few beds of coal, limestone, and conglomerate. The Shawnee Group comprises four limestone and three shale formations, within which members are arranged in distinctive cyclic sequences. The Wabannee Group consists of mudstone, sandstone, thin limestone units, and a few thin beds of coal. Channel sandstones are common in the upper part of the Wabannee Group. A few beds in interval E are excellent markers and can be traced for long distances in the subsurface of central and western Kansas.

Interval E is mostly clayey mudstone and sandy clayey mudstone east of the Nemaha anticline. The lime content increases near the anticline, especially in parts of Nemaha and Pottawatomie Counties, in Jackson, Wabannee, Geary, and Morris Counties, and in south-central Cowley County. Throughout most of the area of the Salina and Sedgwick basins, the interval is primarily limy, locally sandy, clayey mudstone. Clayey limestone and pure limestone constitute interval E throughout almost all of western Kansas. Limy, clayey mudstone is the dominant facies in parts of Cheyenne and Rawlins Counties in northwesternmost Kansas and in parts of Seward, Meade, and Clark Counties in southwestern Kansas.

Clayey mudstone in interval E is mostly red, gray, or, except in western Kansas, greenish gray; variegated mudstone occurs in central Kansas; beds of black fissile clayey mudstone are numerous and widespread in the eastern part of the State, especially in formations of the Shawnee Group, but black mudstones are few and local in central and western Kansas. At some places in eastern Kansas, clayey mudstone units contain sandstones several tens of feet thick.

Limestone of the interval everywhere is buff to gray and anaphastic to finely crystalline; locally, limestone is dolomitic or cherty; oolitic limestone is common, especially in western Kansas. The limestone contains crinoids, ostracodes, bryozoans, brachiopods, algae, and abundant fusulinids (Adkison, 1963, p. 8-9; Collins, 1947, p. 4; Maher, 1947, p. 6), all of which are useful in interpreting environments of deposition. Some of the fusulinids are weathered and worn (Collins, 1947).

Thin discontinuous coal beds are in interval E in Butler, Sedgwick, and Reno Counties (Adkison, 1963, pl. 1); plant fossils and impure coal beds are present in the southermost townships of Seward and Meade Coun-
ties. Thin beds of arkosic sandstone are in the subsurface in Hamilton and Clark Counties.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

Sedimentation apparently was continuous from interval D time into interval E time throughout all of Kansas except perhaps the south-central part (area U, fig. 28). Oscillations of the sea occurred periodically. In eastern Kansas, terrestrial environments developed at many places during the regressive phases.

Source areas of terrigenous detritus deposited in eastern and southern Kansas were mainly the Arbuckle, Ouachita, and Ozark Mountains (Ball, 1964). In the youngest beds of the interval, detritus probably came partly from the north and northwest (Mudge, 1956, p. 676). Thin beds of arkosic sandstone in southwestern Kansas suggest the proximity of exposed basement rocks on the Apishapa and Sierra Grande uplifts in eastern Colorado.

Comparisons of lithofacies and faunas show remarkably similar depositional environments throughout Kansas during interval E. Faunas in the carbonate rocks suggest that all the limestones were deposited in shallow open seas. These seas probably persisted longer in central and western Kansas than in the east, but coal beds in south-central Kansas and both plant fossils and impure coals in southwestern Kansas indicate that the strandline may have fluctuated widely. Probably most of Kansas was emergent for short periods during interval E, and the region apparently consisted largely of broad coastal plains at such times.

The following discussion of environments of deposition in each of the three groups in interval E includes descriptions of the lithology where this information is pertinent to environmental interpretation.

**DOUGLAS GROUP**

In the area of outcrop the Douglas Group (table 20) is mostly a thick section of clayey mudstone and sandstone containing a few thin limestones and small amounts of conglomerate and coal (fig. 33). It thickens southward, and the minor components largely disappear in southern Kansas and northern Oklahoma (Ball, 1964).

The clayey mudstones are mostly gray or greenish gray; a few beds are reddish brown. Fauna of the mudstone beds includes gastropods, arenaceous foraminifers, ostracodes, nautiloid and ammonoid cephalopods, conodonts, bryozoans, and abundant pelecypods (Ball, 1964; Wagner, 1954). Silty and sandy mudstones locally contain remains of land plants. In general, the pelecypod-bearing sandy, clayey mudstone beds of the Douglas Group in eastern Kansas grade northward into brachiopod-bearing clayey mudstones in Nebraska. This change in lithology and fauna is evidence that nearshore depositional environments graded northward into deeper water offshore (Ball, 1964, p. 278).

Sandstones of the Douglas Group probably were deposited in terrestrial and marine environments, mostly near the shores of the sea (Ball, 1964, p. 306-311). Lenses of fine-grained tan to brown quartzose sandstone and siltstone occur within mudstone beds at many places in the outcrop area. Some units of sandstone are as thick as 65 feet (Lins, 1950, p. 124). Channel sandstones extend through parts of several counties in eastern Kansas and thicken generally southward, indicating a source to the south. The sandstone commonly is thin bedded or crossbedded and locally contains ripple marks, contorted bedding, flow casts, and related sedimentary structures (Lins, 1950; O'Connor, 1960; Ball, 1964, p. 90-91). Some of the sandstone beds contain flakes of muscovite mica, a few percent feldspar grains, and plant fossils.

Conglomerates are common in the Douglas Group, occurring chiefly in the lower parts of the thicker sandstone units (fig. 33). Pebbles in the conglomerates are mostly limestone and mudstone derived from beds of the Lansing and Douglas Groups (Ball, 1964, p. 283-284; O'Connor, 1960, p. 29; Lins, 1950, p. 119-120). The matrices are quartz sand and silt. These beds contain invertebrate fossils, mostly reworked, and plant fossils (Ball, 1964, p. 293; Lins, 1950, p. 120-121).

Coal beds of the Douglas Group generally are thin and lenticular and extend only a short distance into the subsurface in eastern Kansas (Ball, 1964, p. 235). Many coal lenses occur within mudstones or limestones bearing marine fossils. Plant material of these coals probably was transported to the sites of deposition and accumulated in littoral environments (Bowsher and Jewett, 1943, p. 38; Ball, 1964, p. 304). At least one coal lens, however, contains stumps and trunks of trees in growth positions and twigs that cut across bedding planes of the overlying mudstone and sandstone (Bowsher and Jewett, 1943, p. 28; Ball and others, 1963, p. 25). This coal bed obviously is autochthonous.

All limestone beds in the Douglas Group probably were deposited in shallow marine waters, perhaps less than 100 feet deep. Some of them originated as calcareous sands or lime muds in nearshore environments (Ball, 1964, p. 272-273). The limestones are only a few feet thick and wedge out southward. Contained fossils commonly include algae, brachiopods, corals, bryozoans, crinoids, gastropods, fusulind, and pelecypods (Ball, 1964). Remains of fish occur locally in the lower part of the Lawrence Formation (table 20) (Miller and Swineford, 1957; Twenhofel and Dunbar, 1914). One limestone member in northeastern Kansas contains streaks and disseminated bits of coal. Some of
the limestone units are crossbedded locally (Ball, 1964, p. 105-109, 262).

Numerous disconformities are recorded from the outcrop area of the Douglas Group, especially in the southern part of eastern Kansas. However, most of these disconformities were caused by erosion by nearshore currents (Ball, 1964, p. 41-42, 255-263).

The prevailing climate as inferred from fauna, flora, and type of deposits probably was warm and humid (Miller and Swineford, 1957). In summary, the outcropping part of the Douglas Group, which thins northward and northwestward, includes mostly terrigenous detrital rocks deposited in a marine environment, but some of the sediments were accumulated in marine-border and swampy terrestrial environments.

Topography of the sea bottom and the coastal region was gentle, and during times of maximal submergence the sea probably was less than about 100 feet deep.

**SHAWNEE GROUP**

The Shawnee Group includes two complete megacyclothem and parts of two others (fig. 34) (Moore, 1949b, p. 143-145; 1950, p. 10-11), which record a variety of environments repeated in sequence. The megacyclothemes are "cycles of cyclothems," each including four or five completely or partly developed cyclothem (Moore, 1936, p. 29). The sequence of cyclothems, the lithologies of beds that compose them, and the faunas that they contain are generally repeated from megacyclothem to megacyclothem (fig. 34; table 21).

The Shawnee Group, like the Douglas, thickens southward and southeastward from northwesternmost Kansas. It grades in the same directions from red and gray sandy mudstone into gray and tan limestone, in-

**TABLE 21.—Rock types and fauna of cyclothem 1 of the Shawnee Group**

[From Moore, 1936, 1949b, 1964; Troll, 1969; Toomey, 1966; Evans, 1966]

<table>
<thead>
<tr>
<th>Cyclothem</th>
<th>Unit</th>
<th>Generalized Lithology and Fauna</th>
</tr>
</thead>
<tbody>
<tr>
<td>E</td>
<td>3</td>
<td>Clayey mudstone containing marine invertebrates in lower part, plant fossils in upper part.</td>
</tr>
<tr>
<td>D</td>
<td>4</td>
<td>Clayey mudstone containing mollusks in lower part, land plants in upper part.</td>
</tr>
<tr>
<td>C</td>
<td>4</td>
<td>Clayey mudstone containing marine mollusks.</td>
</tr>
<tr>
<td>B</td>
<td>2</td>
<td>A single thin but very extensive bed of uniformly fine grained limestone containing abundant foraminifera, including arenaceous forms and fusulinids, and other marine fossils, mainly brachiopods, gastropods, and pelecypods.</td>
</tr>
<tr>
<td>A</td>
<td>4</td>
<td>Thick limestone, locally oolitic, containing fusulinids, mollusks, brachiopods, echinoids, crinoids, algae, and foraminifers.</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>Clayey mudstone containing nearshore invertebrate fossils, especially pelecypods and brachiopods.</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>Sandstone, locally conglomeratic and locally containing plant fossils.</td>
</tr>
</tbody>
</table>
terbedded with clayey mudstone, that occupies much of the rest of the State (Rascoe, 1962, p. 1364-1365).

Some formations can be correlated for long distances in the subsurface (Adkison, 1963, pl. 1). A few rock units of some cyclothemes, especially the single-bedded limestones and overlying black fissile mudstones (units
B2 and C1, fig. 34), are traceable throughout vast areas. The black fissile mudstone of the Oread Limestone, for example, underlies 150,000 square miles and has a uniform lithology throughout this large area (Evans, 1966, fig. 3, p. 60-61).

Limestones and most of the mudstones of the Shawnee Group accumulated in marine waters; thin coal beds and some of the sandstones accumulated in terrestrial and marine-border environments. The remarkably uniform thickness and lithology of some units across wide areas suggest little variation in topographic relief or in other factors that controlled sedimentation. Members of a few cyclothems of the Shawnee Group have been studied closely, and their depositional environments are reasonably well known.

The genesis of the sandstone that forms the basal unit of a type A cyclothem (unit A1, fig. 34) in a megacyclothem is difficult to interpret. This sandstone generally contains plant fossils but no marine fossils (Johnson and Adkison, 1967, p. 83; O'Connor, 1960, p. 42, 45-47). It is judged to be nonmarine, and perhaps it accumulated as detritus on the flood plains of coastal rivers. Clayey mudstone and coal that overlie the sandstone (unit A2) represent deposition in swamps, probably located near the margins of the seas (Troell, 1969, p. 21). Clayey mudstones (unit A3) that overlie the coal beds were deposited near the shoreline of a transgressing sea. Thick limestone (unit A4) probably was laid down during an episode of rapid marine transgression followed by a stillstand and the onset of a regression (Troell, 1969; Johnson and Adkison, 1967, p. 83). The limestone was deposited in an environment that ranged from tidal flat to open marine. In open-marine environments the water depth averaged perhaps 50 feet (Troell, 1969).

The lower part of the clayey mudstone that comprises unit B1 (fig. 34) is mostly nonmarine. It probably was deposited in coastal swamps and estuaries during regression of the sea (Troell, 1969, p. 23; Johnson and Adkison, 1967, p. 81-84; Moore, 1936, p. 148, 163). The upper part of the clayey mudstone records the onset of an episode of transgression; the topmost beds probably accumulated in shallow water near the shorelines (Moore, 1964, p. 311; Johnson and Adkison, 1967, p. 81-83; Troell, 1969, p. 23). With deepening of the seas, exceptionally uniform single-bedded limestone of a type B cyclothem (unit B2) was deposited conformally upon the clayey mudstones. The limestone accumulated on extensive platforms, probably in quiescent relatively shallow open-marine water (Toomey, 1966).

The depositional environment of the basal black fissile clayey mudstone comprising unit 1 of a type C cyclothem (fig. 34) is problematic. The environment has been described as (1) a marine swamp (Moore, 1950, p. 11); (2) a vast expanse of open water only a few metres deep (Moore, 1964, p. 344); (3) a shallow sea blocked by exceedingly dense seaweed and having weak circulation of water (Wagner, 1964, p. 583, 584, 590); and (4) deep water having restricted circulation in the central part of a basin (Evans, 1966, p. 120-126). The first three explanations suggest that the mudstone was deposited during a regression of the sea; the deep-water explanation requires that the shale was deposited far from shore during maximal transgression. The evidence compiled by Evans (1966) for the deep-water explanation is impressive. If Evans’ conclusions are correct, cyclothems B and C (fig. 34) record a single cycle of transgression and regression, and megacyclothems of the Shawnee Group may be composed of only four cyclothems.

Gray clayey mudstone overlying the black fissile mudstone (unit C2, fig. 34) is believed to have been deposited in shallow marine waters. It may record a short-lived episode of regression. The overlying limestone (unit C3) probably accumulated in clear marine water less than about 60 feet deep but 50 miles or more from shore (Moore, 1964, p. 318). Uppermost beds of a type C cyclothem (unit C4) are clayey mudstones deposited near shore during regression of the sea (Johnson and Adkison, 1967, p. 81-82).

Basal deposits of a type D cyclothem (fig. 34) are sandstone and mudstone (unit D1) that may be partly terrestrial in origin, overlain by mudstone deposited near shore in shallow water (unit D2) (Johnson and Adkison, 1967, p. 81-82). The mudstone is overlain by limestone (unit D3) probably deposited in shallow clear water. Regression of the sea apparently began during deposition of the limestone, and the upper part of the limestone may have accumulated above wave base (Johnson and Adkison, 1967, p. 81-82). This limestone grades upward into nearshore, perhaps estuarine, clayey mudstone which in turn grades up into nonmarine clayey mudstone (unit D4) (Johnson and Adkison, 1967, p. 82-83).

The lower part of a type E cyclothem is composed of mudstone and sandstone, some deposited on flood plains and swamps (unit E1, fig. 34) (Johnson and Adkison, 1967, p. 82-83). The mudstone grades upward into argillaceous limestone (unit E2) probably deposited near shore in shallow water. Uppermost beds of a type E cyclothem (unit E3) are nearshore marine clayey mudstones.

The environments described here were repeated with local variations for the other cyclothems and megacyclothems of the Shawnee Group.

WABAUNSEE GROUP

The Wabaunsee Group includes several cyclothems (fig. 35) in eastern Kansas, which are not grouped into megacyclothems. As in the Shawnee Group, the cyclothems record alternate marine and nonmarine
deposition, chiefly of sandy and clayey mudstone and sandstone but including many extensive thin limestones to the north and west of the area of maximum thickness of the group in southeast Kansas.

Transgressions during deposition of the Wabaunsee Group in eastern Kansas probably were across an extensive platform. The sea was quiescent, clear, and warm, and maximum depths are believed to have been only

---

**Figure 35.** Generalized section of Wabaunsee Group in outcrop area of eastern Kansas. Stratigraphic positions of about 12 cyclothsms (C) are shown in right column (not drawn to scale). The 10 units (left column) that compose a Wabaunsee-type cyclothem are:

1. Coal
2. Root Shale
3. Stotler Limestone
4. Pillsbury Shale
5. Zeandale Limestone
6. Willard Shale
7. Emporia Limestone
8. Auburn Shale
9. Bern Limestone
10. Scranton Shale
11. Howard Limestone
12. Sever Shale

(From Moore, 1936, p. 23, 25, 26; 1948b, p. 181; Jewett and others, 1968, pl. 1; Johnson and Wagner, 1967, pl. 3.)
specific details of depositional environments represented by most members of the Wabaunsee Group are not known. However, several members have been studied closely (Mudge and Yochelson, 1962; Mudge, 1956; Owen, 1959), and a few exceptions to the typical cyclothem have been described (Mudge and Yochelson, 1962, p. 102). The general interpretation of Moore (1936) for depositional environments of Wabaunsee cyclothems seems to be confirmed.

Mudstones of the Wabaunsee Group are mostly bluish gray to yellowish gray, although, in the upper part of the group, some beds are red, green, or maroon. Fossils are not abundant in the mudstone, but they include brachiopods, pelecypods, gastropods, crinoid columnals, bryozoans, ostracodes, a few corals, and sparse fucoidal markings (Moore, 1949b; Mudge and Yochelson, 1962; Mudge and Burton, 1959; Johnson and Wagner, 1967, p. 141-142). Evidence of marginal-marine or nonmarine environments includes fossilized wood and plant fragments, at least some of which came from land plants, in several mudstone members of the Wabaunsee Group and generally in association with sandstone lenses (Moore, 1949b; Mudge and Yochelson, 1962, p. 11, 13; Johnson and Wagner, 1967). Several thin coals are in mudstone of probable nonmarine origin (Moore, 1936, p. 25; 1964, p. 289).

Fine-grained gray quartzose sandstone that composes much of the Wabaunsee Group occurs mostly as lenses and channel fillings within mudstone members (fig. 35). The channel fillings range in size from a few feet wide and deep to one channel as wide as 3.5 miles and another as deep as 105 feet (Mudge, 1956).

The channel sandstones commonly contain muscovite and chlorite mica, and some beds have minor amounts of frosted quartz grains, feldspar, garnet, tourmaline, and opaque heavy minerals (Mudge, 1956; Mudge and Burton, 1959, p. 26; Mudge and Yochelson, 1962, p. 11). Many sandstones contain macerated plant fragments, fossilized wood, and, locally, thin lenticular beds of coal. Crossbedding and ripple marks are common locally (Mudge, 1956; Mudge and Yochelson, 1962; Johnson and Wagner, 1967).

Conglomerates are common in channel fillings, and pebbles of limestone and clayey mudstone are their main constituents. Both marine and nonmarine fossils, including fish and amphibians, have been recovered from conglomerate that probably was deposited in a fluvial-estuarine environment (Rasmussen and others, 1971). A minor proportion of the sandstones and mudstones filling channels in uppermost Wabaunsee rocks may be marine (Mudge, 1956, p. 674-675).

Limestones of the Wabaunsee Group are generally thin, but they extend throughout large areas. Their average thickness probably is less than 3 feet. The limestone is mainly gray and finely crystalline to aphanitic; however, some beds are argillaceous, coquinaid, or, locally, conglomeratic, breciicated, or crossbedded (Johnson and Wagner, 1967; Mudge and Yochelson, 1962; O'Connor, 1953). Most limestones are thin bedded to medium bedded. Small bioherms composed mainly of pelecypods and gastropods constitute a limestone facies of one thick clayey mudstone unit. One bioherm is 10 feet thick and underlies approximately 15 square miles (Owen, 1959; Johnson and Wagner, 1967, p. 147-148).

Limestones of the Wabaunsee Group contain abundant marine fossils, especially encrusted algae, fusulinids, brachiopods, crinoid stems, corals, gastropods, pelecypods, bryozoans, echinoid spines, and ostracodes (Johnson and Wagner, 1967; Mudge and Yochelson, 1962; O'Connor, 1953). Local occurrences of amphibian tracks and raindrop impressions in one limestone member of the Wabaunsee (Schoewe, 1956; Johnson and Wagner, 1967, p. 138) suggest that some of the limestone beds may have been deposited in the intertidal zone.

**PALEOTECTONIC IMPLICATIONS**

The major basins of Kansas continued to subside during the time of interval E. The Forest City basin in northeastern Kansas and the Cherokee basin in the southeast formed a shallow synclinal depression along the east flank of the Nemaha anticline. Subsidence along the axis of the basin was generally more than 1,000 feet and reached more than 1,400 feet in northern Cowley County.
A wide shelf that subsided moderately developed on the site of the former Salina basin. The Sedgwick basin subsided about 600 feet at the Kansas-Nebraska boundary in Jewell County and about 1,500 feet near the Kansas-Oklahoma boundary in southwestern Sumner and southeastern Harper Counties. Maximum subsidence in the Hugoton embayment was more than 1,100 feet, near the Kansas-Oklahoma boundary in southeastern Clark County. The northern flank of the embayment was broad and poorly defined. Minor differential subsidence occurred throughout southwestern Kansas.

The Nemaha anticline subsided appreciably less than adjacent areas during the time of interval E. It probably had little effect on sedimentation (Ball, 1964, p. 222-227). Subsidence was retarded, also, on the Cambridge arch and Central Kansas uplift, but both were submerged. These positive elements composed a broad anticlinorium that extended from south-central Kansas through the northwestern counties into southwestern Nebraska. The Keye dome was folded slightly, but apparently was not an important source of detrital sediments.

Periodic transgressions and regressions of the sea during time of interval E probably were mostly caused by eustatic fluctuations of sea level. Such fluctuations may have been due to subsidence or uplift within the northern midcontinent region or to changes in amounts of sediment transported from source areas to the south, southeast, and north. Thin but relatively uniform and extremely widespread beds indicate that any differential tectonic movement was small. At the end of interval E time all of Kansas probably was submerged.

TOTAL THICKNESS OF PENNSYLVIANIAN ROCKS

THICKNESS TRENDS

Pennsylvanian rocks in Kansas range in thickness from an erosional edge in southeastern Kansas to at least 3,000 feet in the axis of the Hugoton embayment in Morton and Stevens Counties (pl. 11). At least in the subsurface of Kansas, Pennsylvanian rocks were not beveled by erosion before deposition of Permian rocks. Therefore, variations in thickness are interpreted as being caused by depositional thickening in synclines and thinning on anticlines. In the area of outcrop in eastern Kansas, variations in thickness are caused partly by post-Pennsylvanian erosion.

PALEOTECTONIC IMPLICATIONS

Major negative structural features that were active during the Pennsylvanian Period were the Cherokee and Sedgwick basins in southeast and south-central Kansas, the Forest City and Salina basins in northeast and north-central Kansas, and the Hugoton embayment of the Anadarko basin in the southwest part of the State (fig. 19). Total subsidence in the deepest parts of the Cherokee and Sedgwick basins during Pennsylvania time was more than 2,500 feet. Total subsidence in the deepest part of the Hugoton embayment was more than 3,000 feet.

Major active positive elements during the Pennsylvanian were the Nemaha anticline and the anticlinorium that includes the Central Kansas uplift and Cambridge arch (fig. 19). Pennsylvanian rocks are thinner than 700 feet on the crests of these structures. Moderately extensive faulting occurred intermittently along the east flank of the Nemaha anticline during the Pennsylvanian Period.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVIANIAN SYSTEM

UNITS OVERLYING PENNSYLVIANIAN

Westward from the area of outcrop, Pennsylvanian rocks are overlain conformably by the Admire Group (Wolfcamp age) of the Permian System (pl. 12). The Admire Group is mostly clayey mudstone and sandstone but contains thin limestones and thin coal beds.

Tertiary and Quaternary deposits rest on Pennsylvanian rocks in small areas in Kansas (not shown on pl. 12). Within the area of Pennsylvanian outcrop thin Tertiary deposits of chert gravel in brownish-red clay overlie the Pennsylvanian locally as high terrace deposits. Pleistocene and Holocene terrace deposits and alluvium lie upon the Pennsylvanian in valleys of the larger streams. Pleistocene tills and outwash of Kansas and Nebraskan age overlie Pennsylvanian rocks in several of the northeasternmost counties.

PALEOTECTONIC IMPLICATIONS

Deposition of the Pennsylvanian continued into the Permian without significant interruption. The original eastern limit of Permian rocks in Kansas is not known, but all Pennsylvanian rocks probably were covered (McKee and others, 1967, pl. 9A, B). Westward tilting of Pennsylvanian rocks occurred during Permian time.

Post-Permian history of Kansas indicates only broad gentle uplift and downwarping, which are reflected in alternate deposition and erosion of Mesozoic and younger rocks. Kansas was emergent during the Triassic, and only westernmost Kansas subsided enough to be covered by Jurassic rocks (Jewett and others, 1968, p. 53-54). Permian rocks probably were eroded from eastern Kansas mostly during Triassic and Jurassic time. Cretaceous rocks may have overstepped the eroded Pennsylvanian rocks from the south and extended into Missouri (O'Connor, 1971). The exposed edges of Pennsylvanian strata were eroded to a peneplainlike “rock plain,” probably during Miocene and Pliocene time (Ham, 1939; Melton, 1959, p. 369). The larger streams of eastern Kansas probably were superimposed on Pennsylvanian and younger rocks.
from the upper Tertiary rock plain. Northeastern Kansas was glaciated during the Nebraskan and Kansan Glaciations of the Pleistocene.

REFERENCES


Adkison, W. L., 1963, Subsurface geologic cross section of Paleozoic rocks from Butler County to Stafford County, Kansas: Kansas Geol. Survey. Oil and Gas Inv., v. 28, p. 90.


Charles, H. H., 1927, Oil and gas resources of Kansas; Anderson County: Kansas Geol. Survey Bull. 6, pt. 7, 95 p.

Collins, J. B., 1947, Subsurface geologic cross section from Trego County, Kansas, to Cheyenne County, Colorado: Kansas Geol. Survey Oil and Gas Inv., Prelim. cross section 5, 6 p.


Davis, J. C., 1959, Reef structure in the Plattsburg and Vilas formations (Missourian) in southeast Kansas: Compass, v. 36, no. 4, p. 319-335.


___1953, Subsurface geologic cross section from Meade County to Smith County, Kansas: Kansas Geol. Survey Oil and Gas Inv., v. 9, pt. 22.


Lukert, L. H., 1949, Subsurface cross sections from Marion County,


O'Connor, H. G., 1953, Rock formations of Lyon County, Pt. 1 of Geology, mineral resources, and ground-water resources of Lyon County, Kansas: Kansas Geol. Survey [Rept.], v. 12, p. 5-24.


1971, Geology and ground-water resources of Johnson County, northeastern Kansas: Kansas Geol. Survey Bull. 203, 68 p.


Trolle, A. R., 1969, Depositional facies of Toronto Limestone Member (Oread Limestone, Pennsylvanian), subsurface marker unit in Kansas: Kansas Geol. Survey Bull. 197, 29 p.


Wilson, F. W., 1957, Barrier reefs of the Stanton formation (Missourian) in southeast Kansas: Kansas Acad. Sci. Trans., v. 60, no. 4, p. 429-436.
Arkansas and Northern Louisiana

By ERNEST E. GLICK

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-1
**CONTENTS**

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>157</td>
</tr>
<tr>
<td>Region defined</td>
<td>157</td>
</tr>
<tr>
<td>Paleogeology</td>
<td>159</td>
</tr>
<tr>
<td>Units underlying Pennsylvanian</td>
<td>159</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>160</td>
</tr>
<tr>
<td>Interval A</td>
<td>160</td>
</tr>
<tr>
<td>Formations included</td>
<td>160</td>
</tr>
<tr>
<td>Upper boundary of interval A</td>
<td>161</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>161</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>162</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
<td>163</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>164</td>
</tr>
<tr>
<td>Interval B</td>
<td>164</td>
</tr>
<tr>
<td>Formations included and stratigraphic relations</td>
<td>164</td>
</tr>
<tr>
<td>Upper boundary of interval B</td>
<td>165</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>165</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>166</td>
</tr>
<tr>
<td>Interval B—Continued</td>
<td></td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
<td>166</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>168</td>
</tr>
<tr>
<td>Interval C</td>
<td>168</td>
</tr>
<tr>
<td>Formations included</td>
<td>168</td>
</tr>
<tr>
<td>Upper boundary of interval C</td>
<td>169</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>169</td>
</tr>
<tr>
<td>Lithofacies trends</td>
<td>169</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
<td>170</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>172</td>
</tr>
<tr>
<td>Total thickness of Pennsylvanian rocks</td>
<td>172</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>172</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>172</td>
</tr>
<tr>
<td>Geologic units directly above Pennsylvanian System</td>
<td>173</td>
</tr>
<tr>
<td>Units overlying Pennsylvanian</td>
<td>173</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>173</td>
</tr>
<tr>
<td>References</td>
<td>173</td>
</tr>
</tbody>
</table>

**ILLUSTRATIONS**

[For listing of plates (in separate case) see volume "Contents"]

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>Map showing Pennsylvanian and post-Pennsylvanian structural features of Arkansas and adjacent areas referred to in text and present geomorphic provinces</td>
<td>158</td>
</tr>
<tr>
<td>37</td>
<td>Map showing counties, towns, and geomorphic provinces in Arkansas referred to in text</td>
<td>159</td>
</tr>
<tr>
<td>38</td>
<td>Map showing Pennsylvanian tectonic provinces referred to in text</td>
<td>160</td>
</tr>
<tr>
<td>39</td>
<td>West-east cross section showing rocks of interval A in northern Arkansas</td>
<td>162</td>
</tr>
</tbody>
</table>
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES,
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

ARKANSAS AND NORTHERN LOUISIANA

By Ernest E. Glick

ABSTRACT

Arkansas and northern Louisiana received detritus from northeastern and eastern sources nearly continuously from the beginning of Pennsylvanian time until about middle Des Moines time. The total volume deposited was about 125,000 cubic miles. Probably only a little more than 1,000 feet of sediment accumulated on the northern part of the shelf area of northwestern Arkansas, but as much as 30,000 feet may have accumulated in the axial part of the Ouachita trough. At the same time, a southward-thickening sequence of sediment accumulated in the ancestral Mississippi embayment and probably merged with sediments of the Ouachita trough.

A deepwater passage across central Arkansas—a trough inherited from the Mississippian Period—received turbidites and other sediments during Morrow and early Atoka time. Eventually it was filled. The northern part became an intermittent coal swamp, and the southern part probably became a shallow marine basin.

In northern Arkansas, deposition during Morrow time was largely that of an open-marine environment. Coal swamps developed in early and in late Morrow time along fluctuating coastlines of the northern Ozark area. Later, during Atoka and early Des Moines time, depositional cycles became more regional, and coal swamps extended southward across the present Arkansas Valley. Most of the post-Morrow deposits are the products of cyclic regressive environments in which apparently few marine animals lived. At times when the water was especially shallow in temporary but extensive nearly sea level swamps of a subsiding basin, coal-producing vegetation flourished. A northeastern source nourished most of the detritus that was deposited in the cyclic sequences.

Rocks of early Des Moines age are the youngest Paleozoic rocks that remain and perhaps are the youngest that were deposited in northern and central Arkansas. Farther south, in southern Arkansas and northern Louisiana, rocks of Des Moines age are known only from sparse and inconclusive subsurface data.

The Ouachita orogeny, which ended Paleozoic deposition in northern and central Arkansas, probably did not begin in Arkansas on a major scale before middle Des Moines time and may have begun much later. As much as 8 miles of vertical uplift in the core area of the Ouachitas generated gravity-induced slidding of enormous segments of rock. Uplift in the Ozark area and Mississippi embayment, along with that in the Ouachitas, left the Arkansas Valley as a large compressed westward-plunging trough. Approximately 100,000 cubic miles of Pennsylvanian and older rock (an average of 1 vertical foot per 20,000 years) was eroded from Arkansas before Cretaceous time. Erosion, which nearly kept pace with the uplift, was greatest in the Ouachita area and was probably least in southern Arkansas and northern Louisiana.

Rocks of Cretaceous and Tertiary age overlap the truncated Pennsylvanian sequence on the south side of the Ouachitas and to the north in the Mississippi embayment. The post-Des Moines erosional break was of relatively short duration in southern Arkansas and northern Louisiana.

REGION DEFINED

Arkansas and northern Louisiana contain a wide variety of Pennsylvanian rocks. The sequence is exposed in the Ozark area, the Arkansas Valley, and the Ouachita Mountains and is buried beneath sediments of the Gulf Coastal Plain in the southern part of the region and beneath the contiguous Mississippi embayment to the east (figs. 36, 37).

Rocks of Pennsylvanian age underlie, and resist erosion on, the several uppermost “dissected plateaus” of the Ozark area. In that area, the rocks are well displayed in the valleys of steep obsequent streams that cut through the Pennsylvanian sequence on the north slope of a regional cuesta whose broad linear crest is the Boston Mountains. The rocks dip southward from the crest into the Arkansas Valley and along the way are displayed in stream valleys that locally cut through the entire sequence. The Pennsylvanian rocks thicken southward, partly from depositional thickening of individual units but largely from the addition of progressively younger beds.

The structural change from gently dipping rocks of the Ozark area to faulted and folded rocks of the Arkansas Valley is partly gradational but takes place generally along the Mulberry fault zone (fig. 36), an east-trending series of branching and coalescing down-to-the-basin normal faults that have a total displacement of about 3,000 feet. Southward across the Arkansas Valley, the amplitude of the folds and the displacement of the faults progressively increase, as does the depositional thickness of the Pennsylvanian sequence. The southern part of the Arkansas Valley is marked by a series of south-dipping high-angle reverse faults, many of which die out in large truncated anticlines that expose as much
as 17,000 feet of the Pennsylvanian sequence. For the northern part of the Arkansas Valley, subsurface data are excellent, but for the southern part they are almost nonexistent.

The structural boundary between the Arkansas Valley and the Ouachita Mountains could logically be placed at the surface trace of any one of several large reverse faults. An apparent lithofacies boundary is mapped at the surface trace of the Ti Valley fault of Oklahoma and western Arkansas and its approximate but largely unmapped extension across central Arkansas. Abruptness of this lithofacies change at that boundary may be exaggerated in the mapping, however, inasmuch as correlative units close to the fault on either side are at no place available for comparison. Undoubtedly, the facies contrast in that area has been made more abrupt by thrusting.

The Ouachita Mountains consist of folded and faulted Paleozoic rocks, as old as Cambrian (?) along the east-trending Broken Bow-Benton uplift of the core area, and as young as Pennsylvanian along either flank (Miser, 1959). Detailed lithologic, thickness, and age data are
sparse for the area. Sedimentary strata of the Gulf Coastal Plain and of the Mississippi embayment on the south and east, respectively, overlap the Ouachitas. Paleozoic rocks beneath the overlapping strata have not been extensively drilled in southern Arkansas and northern Louisiana and are too little known to justify any but the most general speculations.

**PALEOGEOLOGY**

**UNITS UNDERLYING PENNSYLVANIAN**

Rocks of Mississippian age underlie the Pennsylvanian System in all of Arkansas and in northern Louisiana (pl. 2). In most of that area, the Pennsylvanian rests conformably on rocks of late Chester age. However, in the central and western Ozark area a slight angular unconformity separates the two systems. There, the Pennsylvanian rests on the truncated edge of the Pitkin Limestone of late Chester age and extends northward onto the Fayetteville Shale of early and middle Chester age.

Southward from the Ozark area the Pitkin Limestone grades into the upper part of a thick fissile mudstone unit that also includes the equivalent of the
Fayetteville Shale. No name other than Mississippian and locally older shale is applicable to most of that sequence of rock because several older units down to and including the Chattanooga Shale (Devonian and Mississippian) grade basinward into it. In the Ouachita Mountains and southward, that sequence, locally sandy, is a major part of the Stanley Shale and possibly of the Jackfork Sandstone, which in this publication is mapped as Pennsylvanian. All those units are conformably overlain by younger rocks of the Pennsylvanian System.

**PALEOTECTONIC IMPLICATIONS**

During the latter part of the Mississippian Period the shallow sea of northern Arkansas withdrew southward from the shelf area (fig. 38) in which mud and calcareous organic debris were accumulating. Faunal evidence (Easton, 1942; Lane, 1967) indicates that Mississippian deposition in northern Arkansas continued on the outer areas of the shelf at least as long as, and probably longer than, it did in the Mississippian type area 200 miles to the northeast. The western and central parts of the Ozark area, however, were exposed to subaerial erosion and locally lost, through erosion, perhaps as much as 150 feet of Chester deposits. The evidence for loss by erosion is not entirely conclusive; offlap deposition during late Chester time may account for much of the basinward addition of progressively younger Mississippian strata. The southern and eastern parts of the present Arkansas Valley and the area to the south remained covered by the sea.

Local tectonic events of Late Mississippian and Early Pennsylvanian time did not change the relative position of trough, embayment, and shelf areas in Arkansas (fig. 38). Regional tectonic events in the Eastern and Central United States brought about a marked increase in the volume of coarse detritus (silt and sand) delivered to Arkansas and northern Louisiana. The Illinois basin and other basins "upstream," which had intercepted and held virtually all coarse detritus moving southwestward toward Arkansas during the Mississippian Period, began to release some of the material stockpiled there and to permit a through-transport of detritus. Thus, a large volume of sediment was delivered by rivers to northern Arkansas for the first time since the Ordovician Period.

In the Ouachita Mountains area and south of it in northern Louisiana, deposits of mostly clay during Late Mississippian time gradually were succeeded by deposits of clay mixed with much sand during Early Pennsylvanian time. Perhaps that change reflected tectonic uplift in an eastern or southeastern (or even a southern) source area. However, a change in the topography of the trough itself may have been required for deposition of sand to take place. The mechanics of spreading thick beds of relatively coarse sand over a broad area are not well understood, and the possibly related tectonic events during this time are vaguely known.

The depositional basin that during Mississippian and early middle Pennsylvanian time occupied the present Ouachita Mountains area is herein called the "turbidite trough." The term is intended to imply that (1) the basin was continuously occupied by deep marine water; (2) no wave energy disturbed any of the detritus as or after it came to rest on the deep-sea floor; and (3) detritus slumped, slid or was otherwise transported into the trough, creating or perpetuating density or turbidity currents that then spread and deposited much of the detritus longitudinally down the trough. The trough was elongate in a generally east-west direction and slightly convex northward; the long axis of the trough maintained a slight westward slope. As each density current gradually lost energy on the nearly flat sea floor, its load settled to the bottom to form an extensive layer (a turbidite), generally only a few inches thick and made up of detrital grains that tended to grade upward from coarser to finer. Each turbidite tended to be scoured intermittently by density currents until it was covered by the next turbidite; the orientation of the scours indicates paleocurrent directions.

**INTERVAL A FORMATIONS INCLUDED**

The Hale Formation and the overlying Bloyd Shale, which comprise the Morrow Series in its type area of Washington County, Ark. (fig. 37), in the eastern part of
the Ozark area and in the Arkansas Valley to the south make up interval A (pl. 13). Farther south, across the thrust faults along the northern boundary of the Ouachita Mountains, the Jackfork Sandstone and the overlying Johns Valley Shale are thought by some geologists to be approximately equivalent in age to the Hale and Bloyd and are included in interval A.

The Cane Hill Member of the Hale Formation rests unconformably on the Pitkin Limestone (upper Chester) or on the Fayetteville Shale (middle and upper Chester) where the Pitkin has been removed by erosion in the northern shelf area or where the Pitkin has graded basinward into the Fayetteville Shale. The hiatus between the Cane Hill and the underlying rocks of Mississippian age decreases southward and eastward from the west-central Ozark area. In the Ouachita Mountains of Arkansas, the Jackfork Sandstone, which is here considered equivalent to the Hale, rests conformably on the Stanley Shale (Upper Mississippian).

A regional unconformity within interval A of Arkansas separates the Cane Hill Member from the overlying Prairie Grove Member of the Hale Formation. The contact between the two members is marked by local channeling throughout the area of outcrop. One major pre-Prairie Grove channel in Newton County (loc. 123) cuts through as much as 300 feet of the Cane Hill and into the underlying rocks of Mississippian age (pl. 10E, sec. DD-D 'D'). The unconformity dies out basinward in the Arkansas Valley.

Another unconformity at the top of the Woolsey Member of the Bloyd Shale in northwestern Arkansas (Henbest, 1962, p. D43) separates the caprock of the Baldwin coal from underlying rocks. Field investigations in 1969-70 suggest to the author that this unconformity may be much more extensive than previously considered and may have been incorrectly identified as post-Morrow in much of north-central Arkansas.

UPPER BOUNDARY OF INTERVAL A

The Bloyd Shale is unconformably overlain by the Atoka Formation (pl. 10E, sec. DD-D 'D', fig. 39); its truncated northern limit is overlapped by the Atoka in the northernmost outcrops of the central Ozark area. In that area, the Atoka rests on the Hale Formation (Purdue and Miser, 1916).

For reasons not obvious, the greatest concentration of resistant coarse-grained basal Atoka sandstone is in the vicinity of, but skewed slightly westward from, the area in which rocks of Morrow age are most deeply truncated. Perhaps the Atoka sea advanced northward against and progressively destroyed a low sea cliff of Morrow rocks in that area, leaving coarser grains as beach and nearshore deposits that were subject to a certain amount of westward longshore transport. Where interval A is most deeply truncated, its upper surface is smooth and almost devoid of significant channels. Farther west in Washington County, where interval A is scarcely truncated, the contact between the Trace Creek Shale Member of the Bloyd and the Greenland Sandstone Member of the Atoka is described as a "channeloid to interfingering" type (Henbest, 1962, p. D43).

Where a complete sequence of interval A is present in northwestern Arkansas, the post-Morrow unconformity is inconspicuous. Sandstone of the Atoka Formation overlying the Bloyd Shale of interval A is generally silty and fine grained. From west-central Madison County eastward, however, an unconformity which has been considered as post-Morrow is obvious, and in this area coarse-grained sandstone overlying the Bloyd has been assigned to the basal Atoka. Limestone mapped locally as part of the lower Atoka, however, is lithologically similar to the Kessler Member of the Bloyd, and the fauna directly beneath the so-called post-Morrow unconformity of north-central Arkansas is consistently of early Bloyd age (Gordon, 1964). These facts suggest to the author the strong possibility that the "basal Atoka sandstone" of north-central Arkansas is of Morrow (middle Bloyd) age and is equivalent to the basal unit of the Dye Member (Henbest, 1962), the caprock of the Baldwin coal. If so, the boundary between the Morrow and Atoka Series is probably within virtually conformable strata in north-central Arkansas, rather than at the unconformity shown at the top of figure 39, which was used as the boundary in constructing maps of intervals A and B and is designated as the pre-Atoka unconformity in this publication.

The Johns Valley Shale has been mapped in Arkansas only along the western part of the northern boundary of the Ouachita province, where it is assumed to be conformably overlain by the Atoka Formation. Most if not all of its exposed contacts are fault contacts. Elsewhere in the Ouachita Mountains of Arkansas, the upper boundary of interval A is at the top of the Jackfork Sandstone, which is conformably overlain by probable Atoka Formation (pl. 10E, sec. FF-F 'F').

THICKNESS TRENDS

Rocks of Morrow age thicken because of deposition eastward and southwestward from their type area in Washington County, Ark., where the complete section is about 350 feet thick (loc. 116). Depositional thickening is partly offset by "pre-Atoka" truncation of the upper units eastward along the area of outcrop. From the type area of the Morrow east-southeastward, in contrast, the thickness increases nearly tenfold in a distance of 175 miles (fig. 39). This rate of thickening appears, on the basis of extremely sparse data, to continue from east-central Arkansas southward into the Ouachita Mountain area. Beneath the Cretaceous and Tertiary rocks of
the Mississippi embayment, rocks of interval A are truncated by a major pre-Cretaceous unconformity. The truncated edge of the rocks swings appreciably southward before reaching the eastern boundary of Arkansas (pl. 3A).

From the type area of the Morrow, southward along the western boundary of Arkansas, the thickness of interval A doubles in a distance of about 60 miles. Farther south, for about the next 20 miles to the south edge of the Arkansas Valley, rocks of Morrow age have not been reached by drilling. In the adjacent Ouachita Mountains, the apparent equivalent of the type Morrow sequence is nearly 10 times as thick as the greatest section yet penetrated along the western border of the State. If the gradual southward rate of thickening continues entirely across the Arkansas Valley, then an abrupt increase in thickness takes place across the fault-delineated boundary of the Arkansas Valley and the Ouachita Mountains area (pl. 10E, sec. FF-FF'). The maximum thickness of the Jackfork Sandstone in the Ouachita Mountains of Arkansas is about 6,600 feet.

**LITHOFACIES TRENDS**

Rocks of the Ouachita Mountains of Arkansas assigned to interval A—the Jackfork Sandstone and the overlying and geographically restricted Johns Valley
Shale—are entirely detrital and are mostly sandstone (pl. 3B). Less than 10 percent shale in a section of Jackfork 5,000-6,600 feet thick, in the central and western part of the southern Ouachitas (loc. 306) was recorded by Miser and Purdue (1929, p. 76); slightly less than 20 percent shale was measured by Glick (in Danilichik and Haley, 1964) in a 4,300-foot section of Jackfork in the east end of the southern Ouachitas (loc. 255); and “little more than half sandstone in Scott County, Ark.” was reported by Reinemund and Danilichik (1957) (loc. 307) in the upper 2,500-3,000 feet of Jackfork exposed in the western part of the southern Ouachitas. Sand grains of the Jackfork mostly range in size from fine to coarse; locally, especially in the lower part of the formation, the sandstone contains rounded milky quartz granules and sparse quartz pebbles.

Rocks of Morrow age in the Ozark area and Arkansas Valley grade from generally calcareous coarse detritals in the north to mostly noncalcareous fine detritals in the south. The sequence grades westward—toward the Oklahoma border—into shale containing thick interbeds of sandy, fossil-fragmental and oolitic limestone (fig. 39). The Cane Hill Member of the Hale Formation, the lower part of interval A, is less limy than the upper part. It is made up of shale, siltstone, fine-grained sandstone, locally derived conglomerate, and local beds of “sideritic limestone” containing total iron reported as Fe₂O₃ equivalent in the amount of 5-20 percent (T. W. Carney and P. A. Bailey, Arkansas Geol. Comm., written commun., 1970). Only a trace of coarse sand and granules, such as are common in the overlying units, is present in the Cane Hill. A local coal bed and a few lenses of phosphorite have been reported from the basal part of the Cane Hill of the central Ozark area (Glick and Frezon, 1963; Arkansas Geol. Comm., 1964, 1965). The most extensive coal bed of interval A in Arkansas is the Baldwin coal of the Woolsey Member of the Boyd Shale in Washington County and adjacent areas.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

The sources of virtually all the sediments that form rock of Morrow age in northern Arkansas were distant; the major contributors were, in the order of their importance, to the northeast, north, and northwest. In the Ouachita Mountains region, the sediments were deposited in an elongate depression aligned slightly north of west and slightly convex northward, down which detrital sediments moved from east to west. Land areas furnishing these detritals were both to the north and to the south of the trough (Klein, 1966). Material entering the trough from the north probably was transported across northern Arkansas. Detritus coming from the south probably entered the trough as far east as southeastern Alabama, which probably was the site of a major source of sediments during Morrow time.

During all of Morrow time the southwest-flowing ancestral Mississippi River delivered detritus that was deposited in a vast deltaic and marine sequence in the northeastern and north-central parts of Arkansas. During Hale time a relatively small volume of detritus may have entered the area in streams flowing from the northwest. But, for the most part, detrital sediments deposited in the northwestern part of Arkansas during Morrow time were swept westward by ocean currents that crossed the main delta front (fig. 39). The west-central part of the State remained a shallow shelf and tended to accumulate muds and both shelly and oolitic lime sediments, along with a few beds of sand that extended westward from the deltaic sequence.

The deltaic sequence in the northeastern part of the State thickened southward and partly merged with a turbidite sequence in the deep trough to the south. Sediment accumulated along the delta front until unstable parts of the mass were detached and pulled down the slope by gravity. These slides or flows probably occurred about one in 100-1,000 years. Similar density currents also undoubtedly came down the trough from the east.

Within the Ouachita Mountains area of Arkansas, turbidite deposition began before Morrow time and continued without a significant interruption into Atoka time (Walthall and Bowsher, 1966). Sedimentation in parts of the trough may have been affected by a post-Cane Hill 300-foot drop of sea level that is convincingly recorded in the rocks of Morrow age in the Ozark area and the Arkansas Valley (Frezon and Glick, 1959, p. 182). Possibly the Jackfork Sandstone exposed north of Little Rock, “suggestive of a deltaic environment” (Briggs and Cline, 1967), was deposited as a result of this substantial lowering of the sea level. If so, the Jackfork may be slightly younger than earliest Morrow.

The southern limit of the turbidite trough (fig. 38) in Arkansas is placed beyond the southermost outcrop of Jackfork in the Ouachita Mountains because the excellent exposures at the DeGray damsite (Stone and others, 1966) and other localities along the southern Ouachita border clearly indicate that deepwater deposits extended southward beyond that area. However, a southward shoaling was required to channel the flows westward. For that reason, and not on the basis of evidence of a possible southern source of sediments, the trough is believed to have been bordered by a subsea “rise” in southern Arkansas. That is, the trough was a linear depression along the shoreward edge of the continental shelf.

Southward retreat of the shoreline early in Pennsylvanian time, coupled with the influx from the north of a large volume of fine sand and clay, probably appreciably changed the regimen of the turbidite trough. The ages of the deepwater sediments are not precisely
 known, but the work of C. G. Stone and Mackenzie Gordon, Jr. (oral commun., 1969), among others, leaves little doubt that much of the thick Carboniferous sequence of central and southern Arkansas is of Early Pennsylvanian age. The partial withdrawal of the sea at the end of Mississippian time may have corresponded to the deposition of the sandy upper part of the Stanley Shale; and the even more extensive withdrawal of the sea during post-Cane Hill time is believed to have corresponded to the deposition of much of the Jackfork Sandstone. As will be discussed later, the "post-Morrow or middle Boyd" retreat of the shoreline perhaps can be correlated with the deposition of the sandy lower part of the Atoka Formation of the northeastern Ouachita area, in and beyond the lower end of the ancestral Mississippi embayment. This linking of events is in part based upon the concept (R. T. Hazzard, oral commun., 1956) that during transgression offshore areas tend to be starved of sediment, and during regression, the reverse occurs.

In Newton County (loc. 123), river gravel of Prairie Grove age was deposited in the bottom of a channel that cut through nearly 300 feet of marine rocks of the Cane Hill (pl. 10E, sec. DD-D') The Prairie Grove Sea filled this and similar channels and then spread across the coastal plain, depositing a complex channel-fill facies overlain by widespread limy medium- to coarse-grained sandstone that grades upward and northward westward into sandy limestone. Coarse sand and milky quartz granules first appeared in abundance in northern Arkansas in the basal Prairie Grove and are especially abundant in the lower part of channel-fill debris. The coarse material indicates an increase in the carrying power of streams delivering the detritus, resulting perhaps from uplift in the source area. A sea of high energy winnowed the coarse material on the shelf area and transported the fine clastics into the basin to the south. Only after the turbidite basin filled, during middle Atoka time, was clayey sand deposited on the shelf.

During deposition of the Boyd, the ancestral Mississippi River continued to bring detritals to northeastern Arkansas; and mud and both organic and oolitic carbonate sediments were deposited on a shallow shelf in western Arkansas. A substantial coal basin developed in and around Washington County, Ark., during the deposition of the middle part of the Boyd Shale. The Baldwin coal, "rarely thicker than 14 inches" (Henbest, 1962), and the overlying beds of the Woolsey Member of the Boyd are unconformably overlapped by the caprock of the Baldwin coal. This caprock is assigned to the Dye Member, which overlies the Woolsey. The caprock was deposited as the sea readvanced into the northwestern part of the State, initiating latest Morrow sedimentation.

### PALEOTECTONIC IMPLICATIONS

The pre-Cane Hill southward retreat of the shoreline may have been coupled with uplift to the north in the Illinois basin region, as indicated by the flood of fine-grained sandstone probably derived from Chester rocks north of Arkansas. A 300-foot drop in sea level that accompanied the shoreline retreat in post-Cane Hill time is believed to be eustatic in origin (Frezon and Glik, 1959). Retreat of the shoreline after Boyd time has been thought to have accompanied uplift in northern Arkansas. But, it seems more likely that the retreat took place in middle Boyd time and that it was accompanied by very little uplift.

As a result of tectonic or other changes in surrounding areas, the supply of detritus to both northern and central Arkansas increased greatly in Late Mississippian time and was abundant throughout the first half of the Pennsylvanian Period. The maximum grain size of the sand delivered increased from fine to coarse during Morrow time. Even the 20,000 cubic miles of detritus believed to have been deposited in Arkansas during interval A time failed to either fill or shoal the basins—the sea floor subsided as sediment accumulated. A postulated large normal fault or a sharp flexure is assumed to have been intermittently active along the western part of the northern limb of the turbidite trough (fig. 38). The fault was activated only where and when the shelf was receiving a light load of sediments and the basin a heavy load.

In the opinion of many geologists, a local arching of north-central Arkansas followed deposition of the Boyd. A substantial part of the Morrow sequence of northwestern Arkansas is not recognized beneath the so-called "basal Atoka sandstone." In the vicinity of Carrollton dome (fig. 36) in southwestern Boone County, normal faults cut rocks of Morrow age, but not the overlying "basal Atoka sandstone" (J. M. Clark, oral commun., 1961). The faulting may have occurred during the deposition of the upper part of the Boyd Shale. But if the "basal Atoka sandstone" of north-central Arkansas is of middle Boyd age (as discussed above), the movement was earlier than has been generally supposed and the Morrow was only locally truncated.

### Interval B

**FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS**

Interval B consists of the Atoka Formation in Arkansas (pl. 13) and its apparent equivalent in northern Louisiana. This rock-stratigraphic unit is several thousand feet thick throughout an extensive geographic area in which it consists of shale and subordinate amounts of siltstone and sandstone. For convenience and uniformity, the term Atoka has been extended to designate
roughly equivalent rocks farther and farther from the type area. A complementary time-stratigraphic designation is the term "Atoka Series," which is now used far beyond the geographic limits of the formation.

Where the Atoka Formation is lithologically distinguishable from underlying rocks (Bloyd Shale) of latest Morrow age and overlying rocks (Hartshorne Sandstone) of earliest Des Moines age, the names Atoka Formation and Atoka Series refer to the same rock sequence. At places, however, rocks assigned to the Atoka Formation are older or younger than those lower and upper boundary markers, respectively. For instance, some rocks of Morrow age are included in the Atoka Formation of the southern Arkansas Valley and the Ouachita Mountains of Arkansas. Faunal data are still too meager to assess the increase in thickness resulting from including some rocks of Morrow age in the Atoka Formation in the compilation of interval B.

Since 1929 virtually all geologists working in Arkansas have used the name Atoka Formation in lieu of Winslow Formation. The sequence mapped as interval B in this publication and herein called Atoka Formation is thought to include the entire Atoka Series and probably includes some rocks of Morrow age in the Ouachita Mountains; undoubtedly there are intraformational unconformities in northern and central Arkansas.

UPPER BOUNDARY OF INTERVAL B

The Atoka has been truncated, dissected, and weathered by post-Pennsylvanian to Holocene erosion in the Ozark area and probably would have been virtually removed from the area except that very resistant beds of sandstone at the base of and within the formation have protected it. In the Ozark area, most of the Atoka is exposed in hills and valley walls as much as 1,500 feet high.

Rocks of Des Moines age overlie the Atoka Formation in the central and western parts of the Arkansas Valley. There, the upper surface of the Atoka was channeled and truncated locally by pre-Des Moines erosion. In an area a few miles southeast of Fort Smith (fig. 37, near locs. 170, 241, and 247), as much as 300 feet of the upper Atoka sequence may have been truncated locally prior to the deposition of the overlying Hartshorne (Hendricks and Parks, 1950, p. 72-73, fig. 6; Haley, 1966, p. 8). The amount of truncation in that area has been determined, for the want of a more accurate method, by tracing a recognizable bed of upper Atoka sandstone (which may or may not be bounded by time lines or be parallel to the youngest bed of Atoka deposited in the area) across a distance of several miles and measuring its convergence or divergence with the base of the Hartshorne Sandstone. The regional angularity so measured between the two units is so slight that measurable local angularity is attributed to local disturbance.

In the southwestern part of the Arkansas Valley a thick sequence of rocks of Des Moines age overlies the Atoka. In that area, the upper boundary of the Atoka is difficult to determine precisely or even to the nearest 300 feet (Reinemund and Danilich, 1957), suggesting that the contact may be gradational. Thus, as stated by Reinemund and Danilich, "The problem of regional correlation of the Hartshorne, earlier emphasized by Hendricks and Read (1934, p. 1052) is much in need of further study."

Interval B of the Ozark area and the Arkansas Valley is overlain progressively eastward by Tertiary and Upper Cretaceous rocks of the Mississippi embayment. The hiatus between Atoka and Late Cretaceous is about 200 million years.

In the Ouachita Mountains of Arkansas, rocks assigned to interval B are the youngest Paleozoic rocks of the area; they crop out along the south flank of the mountains and in the cores of major synclines (Miser, 1959). Cretaceous and, in one small area, Tertiary rocks of the Gulf Coastal Plain of southern Arkansas lap over the beveled edges of the erosional remnants of interval B along the southern Ouachitas.

Farther south, in the subsurface of southern Arkansas and northern Louisiana, rocks of possible Des Moines age have been reached in two wells (loc. 343, Arkansas; loc. 9, Louisiana). From this limited information one can speculate that in this area, older rocks of Pennsylvanian age, including perhaps some that could be assigned to interval B, are buried at a considerable depth.

THICKNESS TRENDS

Interval B thickens generally southward from the Ozark area into and across the Arkansas Valley (pl. 64); however, local details are uncertain. Erosion has removed parts of the unit and has greatly modified the original depositional thickness and so the thickness trends largely reflect post-Pennsylvanian beveling and dissection.

Although interval B is overlain by younger rocks in a considerable area of the western part of the Arkansas Valley, the thickness data do not fit a smooth isopach pattern. Appreciable local variations in thickness commonly are found across normal faults, some of which may have been active during deposition. In general, the thickness of interval B beneath the protective cover of the Hartshorne Sandstone ranges from slightly less than 5,000 feet in the northern part of the Arkansas Valley to at least 12,000 feet (and perhaps as much as 15,000 ft) at the southern limit of the Hartshorne, about 40 miles to the south.
To the north beyond the limit of interval C, interval B loses about half its thickness through truncation, where it passes northward from the downstream block to the upthrown block of the postdepositional Mulberry fault system. The remaining 2,000 feet or more of the Atoka is lost northward, between the Mulberry fault system and the Missouri border, by further truncation. To the east, thinning also takes place, but slightly more gradually. The northeast limit of interval B is along a line extending generally southeastward from the outcrop in the eastern Arkansas Valley to the eastern border of the State (pl. 6A).

Interval B thickens generally southward in the Arkansas Valley beyond the limit of Hartshorne cover, except where the interval is truncated on structural highs. In a strip about 20 miles wide just north of the Ouachita Mountains, a largely unmapped sequence of folded rock is segmented by high-angle reverse faults. Within the fault blocks, rocks assigned to interval B are as much as 15,000 feet thick. Subsurface data in this area generally are inadequate, but tend to indicate that wells drilled here begin and end in the Atoka Formation.

Along the south side of the Ouachita Mountains, as much as 6,000 feet of rock is assigned to the Atoka Formation (Miser and Purdue, 1929). The sequence lies conformably on the Jackfork Sandstone. The Atoka Formation of the Ouachita Mountains, which may in places include in its basal part beds equivalent to the Johns Valley Shale, is herein assigned to interval B because there is no convincing evidence that it is not mostly Atoka in age.

LITHOFACIES TRENDS

Interval B, or the Atoka Formation, of Arkansas consists largely of shale, siltstone, and sandstone. Minor constituents, which make up much less than 1 percent of the formation but which are significant as marker beds are bentonite(?), clay ironstone, limestone, and coal. Sandstone appears to be evenly distributed in the lower, middle and upper parts of complete sections of the Atoka. The distribution of sandstone in truncated sections is assumed to be similar in most places. In much of the Ozark area and in some other places where all but the basal sandstone is eroded away, the resulting high sandstone percentage fails to show the originally deposited proportions of sandstone to mudstone. Sections less than 1,000 feet thick, therefore, were generally not used in making the lithofacies maps.

Lithofacies divisions of interval B in Arkansas have boundaries that trend slightly north of east (pl. 6B), roughly parallel to pre-erosional isopachs (pl. 15A). The sand percentage decreases from north to south in the direction that the thickness increases, though locally the actual thickness of sandstone may increase. In the western Ozark area, slightly more than 50 percent of the truncated section is sandstone; probably a similar percentage of sandstone was deposited in the now eroded upper part of the formation. Farther south in the western part of the northern Arkansas Valley and in the eastern Ozark area, 20-50 percent of the unit is sandstone. Southward, no fairly thick section of Atoka is reported to have more than 20 percent sandstone, except possibly the southernmost belt of the Arkansas Valley (the "frontal Ouachitas") where the Atoka is reported to contain a thick sandy "lower Atoka lithic unit" (C.G. Stone, oral commun., 1963). The increase in proportion of sandstone in this region may be due to the inclusion of sandy beds equivalent to the Johns Valley Shale in the basal part of interval B.

Some of the sandstone and siltstone in the Atoka is calcareous, and some is very fossiliferous, but only a small amount has even close to the 50-percent calcium carbonate content. A few thin beds of relatively pure limestone are found in wells; these represent very special local environments. Most limestone beds in the lower part of the formation contain marine fossils; the limestone beds in the upper part of the formation are less obviously marine.

Clay ironstone occurs as nodules and thin beds in many shale units of the Atoka Formation; it is especially evident in fresh exposures in the southern Ozark Mountains and in the northern Arkansas Valley. Siderite is recorded on many well logs of the formation and likely was the original matrix of a few heavily iron-stained fossiliferous beds of the upper part of the Atoka. The abundance of limonite on the surface of certain weathered sandstone cliffs suggests that iron from siderite or some other relatively unstable mineral is being leached and redeposited.

Coal is present in thin beds in the upper half of the Atoka Formation in the Arkansas Valley. Haley (1960, p. 818) stated: "The coal beds in the Atoka formation are not mined on a large scale because of the thinness and poor quality of the coal, but two of them have been mined on a small scale to supply local demand. The most extensively mined coal bed in the Atoka formation is near Centerville in Yell County**. * * . Reserves have not been calculated for any of the coal beds in the Atoka formation."

Some coal beds of the Atoka Formation are thought to extend as far as 30 miles (Haley, 1966, p. 11). All wells drilled through the upper half of the formation have a good chance of penetrating one or several coal beds. The beds are not known to be of any special use in correlation but they are of use as environment indicators.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Analyses of the environment of deposition of the Atoka have been provided by Miser (1929, 1934), Crones (1930), Hendricks (1937a, b), Hendricks and Parks
reached as far west as east-central Oklahoma during Atoka time.

The presence and distribution of invertebrate fossils in sandstone beds in the Atoka offer some evidence of an intermittent marine environment. The probable oscillation of the marine-nonmarine boundary is reflected in the presence of both transgressive and regressive sandstones in the Atoka Formation, though most sandstone units of the Atoka are, in the author's opinion, regressive. Some local sandstone units that rest on minor unconformities and grade upward into mudstone are obviously transgressive. Such units are most common in the Ozark area.

The invertebrate fossils also suggest changes in the character of the marine environment on the shelf in the Ozark area during Carboniferous time. Clear open-marine water above a mud bottom probably prevailed during Late Mississippian time. The water may have become muddy or brackish, starting at about the beginning of Atoka time, inasmuch as open-marine fauna did not survive much past early Atoka time. Few, if any, cephalopods are found in the Atoka more than 1,500 feet above the base of the formation; remains of other marine animals are found somewhat higher in the formation, but they also decrease in abundance upward.

After not more than a third of interval B had been deposited, the deepwater trough that crossed central and southern Arkansas ceased to receive turbidites. Deposition, however, outstripped subsidence and the northern part of the trough became filled with sediment and the environment changed from deep to shallow marine. North of Little Rock about 2,000 feet of shale overlies the turbidite sequence and underlies the first shallow-water sandstone—the lowest of the beds informally called the "traceable 3" sandstone complexes. (C. G. Stone, oral commun., 1965). Ultimately, the basin was occupied by shifting coal swamps.

Farther south in the Ouachita Mountains and in the Coastal Plain area of southern Arkansas and northern Louisiana, correlatives of middle and upper interval B either have been mostly eroded or are buried deeply in the subsurface; the regional thickness relations suggest that the eastern source continued to supply coarse clastics to those areas during at least middle Atoka time (C. G. Stone, oral commun., 1970).

As the southern part of the trough filled, the flooded shelf on which the ancestral Mississippi emptied formed an embayment that opened more to the west or actually shifted westward. The depocenter of the trough-shelf system shifted toward west-central Arkansas, and the environment in that area changed from generally muddy open-marine to delta-complex. During this shift the ancestral Mississippi River meandered freely across northern Arkansas. As a result, "considerable difficulty is encountered in correlating what might be called the
middle Atoka. This section, more than the upper and lower Atoka, exhibits not only extreme thickness variations within short distances but also lacks laterally persistent marker beds” (Bacho, 1961, p. 66).

Environments of deposition during middle and late Atoka time were similar to the environment in early Atoka time except that marine incursions seem to have become less numerous; a few marine fossils occur throughout the sequence. Reference to these is made by Hendricks and Parks (1950, p. 70) as follows:

Marine fossils and well-preserved plant fossils have been found at only a few localities, and none of the fossiliferous zones can be traced laterally beyond a single exposure.

The lithologic character of the formation indicates that the sediments were mostly deposited on wide, comparatively level surfaces that stood close to sea level.

The origin of some sandstone beds in the middle and upper part of the Atoka Formation has long been an enigma. The problem was discussed by Seull (1961, p. 156) as follows:

For the writer, the most difficult feature to explain, in resolving the depositional history of the Atoka, is the origin of the massive-beded sandstones. Some of these beds, up to 300 feet thick, can be traced east-west for several miles, yet the north-south extent cannot be determined from outcrops because of high dips. The maximum trace was slightly less than a mile. They have no depositional structures such as cross-bedding, do not vary in thickness over a few feet, contain clay balls, but otherwise are uniform in texture, and do not contain plant or animal remains. These sands may represent broad, deep river channels almost at grade in which the creep of the water-saturated sandfill resulted in grain size segregation.

The scarcity of marine fossils, the channeloid contact at the base of many massive sandstones, and the presence of underlying or overlying local lenses of coal, however, do not rule out a marine origin for these sandstones. Regionally, the entire upper part of the Atoka grades southward and thickens rapidly into a sequence in which the sandstone is finer grained and less abundant. Where the sandstone beds are thick they tend to be elongate east-west or roughly parallel to the axis of the basin, as would be expected of offshore bars. Where subsurface control is close, each sandstone body can be traced in the subsurface for several tens of miles. Determination of the origin of these sandstones still awaits extensive detailed regional investigation.

As interval B ended, a few streams meandered across a slightly emergent coastal plain in northern Arkansas, and the shore of a shallow muddy sea extended westward across central Arkansas. Possibly the central part of what was later to become the Ouachita Mountains formed low islands in the southern part of the State, though land in that area may not have emerged until a little later, in interval C time.

**PALEOTECTONIC IMPLICATIONS**

Tectonic activity in Arkansas during Atoka time was largely restricted to downwarping and downfaulting of the east-west trough across the central part of the State. During deposition of the upper two-thirds of interval B, the rate of subsidence seems to have lagged behind the rate of accumulation, causing the trough to be filled.

In middle Atoka time the ancestral Mississippi River shifted westward across the shelf in northern Arkansas. The cause of that change, which brought about the beginning of “typical Atoka deposits,” is not evident. A postulated early arching in the Ouachita Mountains to the south hardly seems to account for the change, an arching in the embayment area itself would have been more likely to deflect the river westward.

During middle Atoka time the shelf area of west-central Arkansas received more sediment than during all of previous Carboniferous time. The high rates of accumulation and subsidence were balanced, which kept the surface near sea level through early Des Moines time. Subsidence seems to have been assisted by the sedimentary load. Normal faults, mostly downdropped to the south, probably were active during deposition (Shields, 1961), with semicontinuous movement, rather than a single episode of displacement.

No marked tectonic activity is known to have occurred at the end of Atoka time in the Arkansas region. However, mild uplift in the Ouachita Mountains region, predating the later, more vigorous uplift in that area, may have taken place as early as middle Atoka time. If so, the evidence of such early uplift in Arkansas has been largely destroyed (C. G. Stone and B. R. Haley, oral commun., 1970).

**INTERVAL C FORMATIONS INCLUDED**

Interval C in the Arkansas Valley consists of remnants of the Krebs Group (pl. 13) of the Des Moines Series. The four Des Moines formations which are present and which crop out in Arkansas are the Hartshorne, McAlester, Savanna, and Boggy. The basal unit, the Hartshorne Sandstone, is conformably overlain by the McAlester Formation. The Savanna Formation rests with an erosional contact on the McAlester in southwestern Sebastian County, Ark. (Hendricks and Parks, 1950, p. 76), and interfingers with the upper McAlester “in eastern Logan County, Ark.” (Haley, 1961b, p. 118). Probably Haley meant “central” Logan County, inasmuch as the Savanna does not extend into eastern Logan County. The Boggy Formation, which in Arkansas consists of only a few truncated outliers, rests unconformably on the Savanna (Haley, 1961b, p. 122).

In southern Arkansas and northern Louisiana, substantial evidence indicates that rocks of Des Moines age are present deep in the subsurface. A well in Hempstead County, Ark. (loc. 343, pl. 1), completed in 1960 at a depth of 10,346 feet, is reported to have reached the top of Paleozoic rocks at a depth of 8,700 feet. Evidence from
that well on the possible presence of Des Moines rocks at depth is furnished in a letter of April 25, 1966, from Chevron Research.

Our palynologic study of cores from the Humble Oil and Refining Company (Carter Division) No. 1 G. B. Royster, Sec. 31-T10S-R24W, Hempstead County, Arkansas, indicates pre-Mississippian and pre-Missourian rocks are present at a depth of 9488-9524 feet.

Present in this well are *Laevigatosporites* which indicates a post-Mississippian age and *Densosporites* which is not reported from rocks younger than the Desmoinesian Series. In addition to this generic evidence for the age, *Laevigatosporites obacurus*, *L. thiesseni*, *Punctatosporites obliquus*, and *Lutosporites globosus*, species which become extinct at approximately the Desmoinesian-Missourian boundary, were recovered from the core listed above.

Our data are such that it is not possible to say with certainty that this core is from Desmoinesian strata. The rocks are of pre-Missourian Pennsylvanian Age.

A well that furnished additional evidence in support of possible Des Moines rocks in the subsurface was completed at a depth of 10,475 feet in Morehouse Parish in northern Louisiana (loc. 9). This well was reported to have reached the base of the Triassic Eagle Mills Formation (Scott and others, 1961) at a depth of 9,285 feet. The underlying Morehouse Formation consists of at least 1,190 feet of marine silty shale and siltstone. Imlay (1940, p. 7-8) reported the following on the Morehouse:

**Correlation.** The Morehouse formation is probably Jurassic age as indicated by the occurrence of a fossil sponge obtained at a depth of 9,304-9,305 feet in the type section. This sponge probably belongs in the family Stauaroerdinae which is known only from the Jurassic. Some pelecypods obtained from the type section are reported to be Mesozoic forms, but the writer has not had the opportunity to examine them.

Later, the Morehouse in the same well was classified by Imlay and Williams (1942) on the basis of its gastropod fauna as upper Paleozoic. Still later, plant spores in a core sample of the interval 10,243-10,253 feet of that well were studied by Hoffmeister and Staplin (1954, p. 154-159), who assumed them to be of Middle to Late Pennsylvanian age.

In the present report, the rocks of pre-Missouri Pennsylvanian age of locality 343, Hempstead County, Ark. (Chevron Oil Co., written commun., 1966), and the Morehouse Formation of Middle to Late Pennsylvanian age at locality 9, Morehouse Parish, La. (Hoffmeister and Staplin, 1954), are arbitrarily assigned to interval C.

**UPPER BOUNDARY OF INTERVAL C**

Except where covered by alluvium, rock of the Des Moines Series in the Arkansas Valley is the present ground surface. Rocks in the subsurface of southern Arkansas that may be of Des Moines age probably are conformable with overlying younger rocks, but the boundary with the younger rocks has not been delimited precisely.

**THICKNESS TRENDS**

The thickness of rocks of Des Moines age as originally deposited in the Arkansas Valley increased regionally from north to south (pl. 7A). However, the sequence can be measured now only in erosional remnants, the thickest of which are in synclines or in downfaulted blocks.

Of the four formations of Des Moines age in the Arkansas Valley, the McAlester Formation shows, on the basis of published data, the most pronounced original southward thickening. It is less than 500 feet thick in the northern part of the valley and more than 2,000 feet thick in the southern part (Haley, 1961b, fig. 3). The underlying Hartshorne Sandstone is less than 100 feet thick in many areas and may be as much as 270 feet thick in northern Scott County, Ark. (Reinemund and Danilchik, 1957). The Savanna is less than 2,000 feet thick in all areas (Hendricks, 1937b, p. 1417; Haley, 1961b, p. 118, 122) except perhaps in north-central Logan County, where it is about 2,200 feet thick (Haley, 1961c, p. 8). As much as 900 feet of the lower part of the Boggy is preserved in one remnant (Hendricks and Parks, 1950, p. 76). The thickest section of the combined formations is about 4,500 feet, preserved in a synclinal mountain in northern Scott County (pl. 10E, sec. FF-'F').

Neither the top nor the bottom of rock sequences of southern Arkansas and northern Louisiana that may be of Des Moines age is, as yet, well defined. The prospects of establishing meaningful thickness trends for interval C are indeed dim in that area. The well at Arkansas locality 343 (pl. 1) in Hempstead County penetrated 1,464 feet of Paleozoic of which at least a 36-foot-thick section in the middle is thought to have pre-Missouri Pennsylvanian fossils. The well at Louisiana locality 9 in Morehouse Parish penetrated 1,190 feet of Morehouse of which a 10-foot-thick section in the lower part has yielded fossils of Middle and Late Pennsylvanian age.

**LITHOFACIES TRENDS**

Lithofacies trends plotted on data from the remnants of interval C in the Arkansas Valley are deceptive unless one is cognizant of the differential erosion of that area. The lowest unit, the Hartshorne Sandstone, is far more resistant to erosion and is more widespread than the overlying units. Thus, the lithofacies trends shown on the small-scale regional map (pl. 7B) reflect little other than the amount of interval C cover remaining on the Hartshorne. Lithofacies trends restored for individual formations can help establish the depositional environments and direction to the source areas.

In much of the northern, central, and western parts of the Arkansas Valley, the Hartshorne rests with a channeloid erosional contact on the underlying Atoka Formation. The thickness and lithology of the Hartshorne change abruptly within short distances at the edge of each of the broad channels. The unit was described as follows by Hendricks and Parks (1950, p. 73): "Where it is thick it is generally coarse-grained,
clean, and thick-bedded or massive; where it is thin it is fine-grained, thin-bedded, and shaly.”

Hendricks (1937b, p. 1419) stated, concerning the orientation of the channels: “Along the southern margin of the Arkansas-Oklahoma coal field and at some places north of that margin, thick, coarse-grained, channeloid parts of the Hartshorne sandstone appear to trend northward and to indicate that the streams that deposited sands of that formation entered the basin of deposition from the south.”

The Hartshorne loses its channeloid nature and virtually disappears southeastward in the Arkansas Valley.

The lithology of the McAlester Formation of Arkansas has never been plotted regionally in enough detail to establish lithofacies trends. Many geologists have, however, furnished detailed partial sections and general descriptions of the formation in their areas of study. A diagram entitled “Sections of lower part of McAlester shale encountered in five diamond drill holes in T. 4 N., R. 32 W., Arkansas” was provided by Hendricks and Parks (1950, fig. 8) and Hendricks (1937a, fig. 5). Their sections are selected from “about 600 diamond-drill records” and each of the five sections depicts about 1,100 feet of the McAlester. An excellent description of the entire formation (2,033 ft 9 in.) from T. 5 N., R. 32 W., also is available (Hendricks, 1937b, p. 1416). These sections, and the detailed descriptions of the McAlester by Haley (1961a, b, c, 1966, 1968), Merewether and Haley (1961, 1968), and Merewether (1967), provide a general, but still far from complete, indication of the lithofacies trends of the McAlester.

Data on the McAlester in the western part of the Arkansas coal field have been summarized by Hendricks and Parks (1937, p. 199). Haley (1961b, p. 118) described the McAlester in the Arkansas coal field as “predominantly shale with minor amounts of sandstone and siltstone, a few thin beds of limestone, and eight coal beds.” The Upper Hartshorne coal bed and an underlying sandstone are also present locally, but according to Haley, “These units pinch out to the north and east in south-central Sebastian County.”

The lithology of the Savanna Formation of Arkansas has been best described from the Bloomer syncline (fig. 36; Hendricks and Parks, 1950, p. 76, 77; Haley, 1966, p. 9, 10). According to Haley (1966, p. 9) “The formation is predominantly shale with minor amounts of sandstone and siltstone, four coal beds, and four limestone beds.” The Savanna Formation of the Waldron quadrangle on the south side of Poteau Mountain in northern Scott County was described by Reinemund and Danilchik (1957).

Apparently the Savanna is everywhere poorly exposed and individual units are thinner, the lithology is more varied, and the beds are more lenticular than those of the McAlester. The most recent map of this area (Haley, 1966, pl. 1), however, show moderate continuity of five of the lower mappable sandstones. Two economically important coal beds, the Charleston in the lower part of the formation and the Paris in the upper part of the formation, are, with some uncertainty, correlative from one area to another.

Limestones, all in the upper part of the formation, are very local. They contain some freshwater fossils (Hendricks, 1937b, p. 1417), but “mostly brachipods, some pelecypods, and a few crinoid fragments” (Haley, 1966, p. 10).

The lithology of the Boggy Formation of Arkansas has not been described in detail. Probably the most complete general description is that by Hendricks and Parks (1950, p. 76), who stated: “The part of the formation that is present in the [Fort Smith] district consists of dark clay shale and gritty shale, with three sandstone beds about 760 to 800 feet above the base. The shales are poorly exposed, and it is possible that they contain some coal beds that were not seen. The sandstone is coarse-grained, medium-bedded, and buff to brown in color.”

SOURCES AND ENVIRONMENTS OF DEPOSITION

One of the constructive early statements concerning the general subject of Pennsylvanian sedimentation in Arkansas is the following by Branner (1896, p. 235-236):

If we inquire into the reason for the great thickness of Coal Measures sediment in the Arkansas Valley, I believe it is to be found in the drainage of the continent during Carboniferous times. The rocks of this series in Arkansas contain occasional marine fossils, and these marine beds alternate with brackish or freshwater beds whose fossils are mostly ferns and such like land or marsh plants. This part of the continent was, therefore, probably not much above tide level. The drainage from near the Catskill Mountains in New York flowed south and west. The eastern limit of the basin was somewhere near the Archean belt extending from New England to Central Alabama. This Appalachian watershed crossed the present channel of the Mississippi from Central Alabama to the Ouachita uplift, or to a watershed still farther south and now entirely obliterated and buried in Northern Louisiana. In any case the drainage flowed westward through what is now the Arkansas valley between the Ozark Island on the north and the Arkansas Island on the south.

Branner’s conclusion that the major source of Pennsylvanian sediments lay far to the northeast and east is virtually the same conclusion reached in this report.

Later workers proposed that a major source area to the south had furnished a large quantity of clastic material to the Ouachita area and the Arkansas Valley during Carboniferous time. This theoretical source area was supported by Miser (1921, p. 89), as indicated by the following statement:

A land area, which has been called Llano by Willis, Schuchert, and Ulrich and Llanoria by Dumble and Powers, existed in Louisiana and eastern Texas during much if not most of the Paleozoic era and during the Triassic and Jurassic periods of the Mesozoic era. It varied in outline from time to time. It may have occupied a part of the area of the
present Gulf of Mexico; at times it was doubtless connected with large land areas that occupied at least much of central and northern Texas, southern Oklahoma, and southern Arkansas, and for short periods it may have extended eastward across the lower Mississippi Valley and joined the southwest end of the Appalachian area. It furnished most of the sediments that formed the elastic rocks of Pennsylvanian age in north-central Texas and those of Ordovician, Silurian, Mississippian, and Pennsylvanian age in the Ouachita Mountains and Arkansas Valley of Arkansas and Oklahoma.

In addition to Llanoria, the Ouachita area itself has been considered by some to have been a southern source during part of Pennsylvanian time. Many geologists have assumed that the Ouachita area had ceased to be a basin and had become a source area at least as early as early Des Moines time. Apparently Llanoria was thought to have been high enough to continue sending clastics across the uplifted and eroding Ouachita area. An excellent summary of data leading to that point of view has been provided by Hendricks (1937b) in his paper on the Pennsylvanian sedimentation in the coal field of Arkansas. He stated (Hendricks, 1937b, p. 1403) that: "The major source of the sediments appears to have been farther south, in and south of the Ouachita Mountains, but some sediments in the lower part of the section probably came from the east."

Inasmuch as the Ouachita area did become a major source of detritus following its uplift, it is important to determine the date of that uplift. Evidence has already been presented in this chapter to support the idea that rocks of Morrow and Atoka age had a northeastern source. If a southern source was a factor in Arkansas during early Des Moines time, currents carrying detritus should have had a northerly direction and paleocurrent direction indicators should furnish such evidence. In other words, the first significant uplift of the Ouachita area should coincide with the reversal of, or at least a major change in, the paleocurrent direction. This change should be from southward or southwestward to northward or northwestward.

Reinemund and Danilchik (1957) showed that most paleocurrent directions for every stratigraphic unit have a southward component. Their own summary, quoted below, probably is as definite as the data allow:

The observations made thus far in Scott County indicate that the sediments of the Atoka formation were mostly in transport southwestward when they were deposited in that area, but *** the kinds of structures and their orientations in the post-Atoka sequence are somewhat different from those in the Atoka, probably reflecting a change in depositional environment that accompanied the initial accumulation of coal-bearing strata overlying the Atoka. Speculation on the significance of these structures is best deferred, however, until they have been mapped and studied much more extensively, and it is hoped that geologists working with these Pennsylvanian formations elsewhere in Arkansas and Oklahoma will be encouraged to do so.

Recent work leaves little doubt that paleocurrents of Atoka and Des Moines time flowed southwest in most of southeastern Oklahoma (Briggs and Cline, 1967; Agterberg and Briggs, 1963). In the author's opinion, similar currents carried detritus southward and southwestward in Arkansas during Des Moines time, and a source to the south, if it existed, was inconsequential.

The Arkansas Valley was occupied by a near-sea-level basin in which deposition and subsidence were rapid and approximately equal. In some places coal swamps alternated with fluvial sands in vertical sequence. A very low energy sea invaded the area periodically from southern Arkansas and northern Louisiana, but rarely left a record of a marine fauna.

The Hartshorne Sandstone is believed to have been deposited by streams on a coastal plain. Streams apparently channeled into a newly emerged plain of low relief that sloped gently southward toward the coastline of central Arkansas in earliest Des Moines time. Sand filled the stream channels and spread across the interfluvies. Much of the ultra-fine carbonaceous material associated with the sand probably was subaerially oxidized, giving the unit its light-gray—nearly white—color. The clay that was carried southward by the streams likely was deposited, along with a steadily increasing amount of sand, under marine conditions.

The regional blanket of Hartshorne Sandstone, 10-300 feet thick, underlies the most extensive bed of coal known in Arkansas. In most areas, a few feet of carbonaceous shaly mudstone separates the nongradational top of the sandstone from the overlying Lower Hartshorne coal bed.

Deposition of coal was followed by transgression of the sea. Rivers delivered detritus to the shore, and the detritus was distributed by marine processes to form a blanket deposit. Mud was progressively overlain by sand in a seaward direction. The resulting sand-capped platform, only slightly below sea level, provided an ideal site for a future coal swamp—just as did the subaerially formed platform of the Hartshorne.

The sandstone units in interval C above the Hartshorne are characterized by gradational bases. These sandstones are the upper parts of depositional units that grade upward from mudstone into very fine grained thin-bedded shaly sandstone, then into slightly coarser grained, thicker bedded sandstone. Unoxidized carbonaceous debris gives the sandstones a gray color, contrasting with the ashly white of the Hartshorne. The tops of the sandstone units are relatively even and nongradational. On them rest carbonaceous mudstones and beds of coal, which in turn are overlain by shaly mudstone that grades upward into another similar unit of sandstone. The stratigraphic interval from the top of one unit of sandstone to the top of the next unit averages about 200 feet in the McAlester Formation and about 100 feet in the Savanna Formation. Each unit represents
one oscillation of the sea. A unit is considered to have been deposited in approximately 100,000 years.

The rivers carrying detritus basinward flowed through brackish-water swamps formed behind the prograding shoreline. The swamps and their coal-producing vegetation probably advanced seaward as the platform was progressively built up. The product of the regressive beach environment, the highest energy part of the cycle, is directly overlain, without gradation, by the product of the low-energy coal-swamp environment. Deposition in the swamps failed to keep pace with subsidence, and the swamps were succeeded by interdistributary lakes in which the coal-producing vegetation died out. As subsidence proceeded, natural levees and barrier beaches were not maintained, marine water advanced over the delta, and the locus of river-mouth deposition shifted landward.

The Des Moines Sea of central Arkansas apparently was rarely deeper than 100 feet. Several deltaic distributaries probably helped in maintaining the shallow-sea floor, though they may not have contributed sediment simultaneously. Of the later cycles of deposition, no one cycle affected the entire basin as did that of the Hartshorne Formation. Throughout Des Moines time regressive deposition exceeded transgression.

PALEOTECTONIC IMPLICATIONS

At the end of Atoka time the southern part of Arkansas is believed by some geologists, including the author, to have remained below sea level. However, other geologists believe the absence of beds of Des Moines age in the central part of the Ouachita Mountains to indicate that the area was above sea level and was undergoing erosion. The northern part became slightly emergent. In the Arkansas Valley—the only place in Arkansas where the details of pre-Des Moines structure can be investigated—local structures with as much as 300 feet of differential uplift may have developed.

After a period of early Des Moines subaerial deposition in the Arkansas Valley, subsidence resumed. Sediment aggregating about 5,000 feet was deposited there by the end of Boggy time. No appreciable tectonism other than downwarping is evident in the early Des Moines record of Arkansas. If Paleozoic rocks younger than those of approximately middle Des Moines age were deposited in northern and central Arkansas, they have been removed by erosion.

Vertical uplift in the Ouachita Mountains may have begun as early as middle Atoka time or the initial movement may have been Des Moines or later Pennsylvanian, and perhaps may have produced, by later Pennsylvanian time, as much as 8 miles of vertical movement along the Broken Bow-Benton uplift (fig. 36). Gravity-induced sliding of enormous proportion probably resulted largely from the primary uplift. A compressive force from the south might have aided that process, but if so, it apparently caused little or no uplift in southern Arkansas and northern Louisiana.

At about the time that the Ouachita area was raised, the Ozark area to the northeast also was uplifted; the Arkansas Valley area, between two areas of uplift, was an enormous westward-plunging trough, containing compressed and wrinkled rocks. One of the larger folds is an anticycle with about 17,000 feet of closure. In southeastern Missouri, western Tennessee, and northeasternmost Arkansas, the Pascola arch rose within the area of the ancestral Mississippi embayment and shed a large quantity of detritus derived from Mississippian and Pennsylvanian rocks.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS
THICKNESS TRENDS

Pennsylvanian rocks generally thicken in the Arkansas region from their erosional edge in the Ozark area southward to the south side of the Arkansas Valley. Pennsylvanian rocks beneath the Mississippi embayment probably thicken to the southwest. A maximum thickness of at least 20,000 feet is estimated along the south edge of the Arkansas Valley; about half that thickness is estimated to be under the Mississippi embayment. Strata within the Ouachita Mountains, although believed to have once included the thickest deposits in the Arkansas-Louisiana region, are now generally reduced by erosion to much less than the Arkansas Valley maximum. Little is known of the thickness of Pennsylvanian rocks that may be present under the northern Gulf Coastal Plain.

PALEOTECTONIC IMPLICATIONS

Through Morrow and much of Atoka time the tectonic elements of the Arkansas region were a relatively stable shelf on the north and a subsiding trough in the present areas of the Arkansas Valley and Ouachita Mountains. The trough probably was bordered on the south by a submerged part of the continental shelf. Subsidence in the trough, assisted by intermittent faulting at least along the northern margin of the trough in western Arkansas, exceeded the rate of filling until late Atoka time. In late Atoka and Des Moines time subsidence was generally slower than filling. Moderate uplift may have occurred to the north in the Illinois basin region at the end of Mississippian time and in northern Arkansas between times of Morrow and Atoka deposition; but much of the cyclic character of Early and Middle Pennsylvanian deposition in the Ozark area is attributed by the author to eustatic shifts in sea level or to shoreline changes that accompanied shifts in dominance of subsidence or sedimentation rates.

The absence of any evidence of change from the dominant northeast-to-southwest direction of sediment
transport in the Pennsylvanian rocks of Arkansas suggests no major uplifts before Des Moines time. The present structural pattern of the region—positive elements in the Ozark area and the more strongly elevated Ouachita Mountains area flanking the westward-plunging, folded, and faulted trough of the Arkansas Valley—is thought to have developed no earlier than middle to late Des Moines and possibly substantially later.

GEOLeIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

Pennsylvanian rocks in the greater part of the Ozarks, Arkansas Valley, and Ouachita Mountains are at the surface or are covered only by alluvium. No Permian rocks are known in the region, and Triassic (?) and Jurassic rocks have been identified only south of the Ouachita Mountains. Cretaceous rocks extend across the Gulf Coastal Plain and into the Mississippi embayment and are, in turn, overlapped by Tertiary deposits where the Mississippi embayment abuts the Ozarks, Arkansas Valley, and Ouachita Mountains areas.

PALEOTECTONIC IMPLICATIONS

Geologic events in northern and central Arkansas cannot be precisely dated within the span from middle Des Moines to Cretaceous time—about 200 million years, or about half as long as the entire Paleozoic Era. The author believes that the entire orogenic development of the Ozark and Ouachita Mountain areas and the Arkansas Valley area was long drawn-out and that erosion kept pace with most of the uplifting. About 100,000 cubic miles of Pennsylvanian and older rock were eroded from Arkansas prior to Cretaceous time. That figure represents an average rate of 1 vertical foot of rock per 20,000 years for the entire State and about 1 foot per 5,000 years in the central Ouachitas.

REFERENCES


1964, In eastern Arkansas exploration is shifting to Paleozoic prospects: World Oil, v. 158, no. 4, p. 72, 75-76, 84.


Haley, B. R., and Frezon, S. E., 1965, Geologic formations penetrated by the Shell Oil Company No. 1 Western Coal and Mining Co. well on the Backbone anticline, Sebastian County, Arkansas: Arkansas Geol. Comm. Inf. Circ. 20-D.


Imlay, R. W., 1940, Lower Cretaceous and Jurassic formations of southern Arkansas and their oil and gas possibilities: Arkansas Geol. Survey Inf. Circ. 12, 64 p.


Lane, H. R., 1967, Uppermost Mississippian and Lower Pennsylvanian conodonts from the type Morrowan region, Arkansas: Jour. Paleontology, v. 41, no. 4, p. 920-942.


Texas Panhandle and Oklahoma

By SHERWOOD E. FREZON and GEORGE H. DIXON

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-J
CONTENTS

Abstract ................................. 177
Regional features ....................... 177
Paleogeoogy ............................ 178
Units underlying Pennsylvanian ...... 178
Lower boundary of Pennsylvanian ..... 178
Paleotectonic implications .......... 179
Interval A ................................ 179
Formations included ................. 179
Upper boundary of interval A ......... 181
Thickness trends ....................... 181
Lithofacies trends ..................... 181
Sources and environments of deposition 182
Paleotectonic implications .......... 182
Subinterval A ............................ 183
Formations included ................. 183
Upper boundary of subinterval A ...... 183
Thickness trends ....................... 183
Lithofacies trends ..................... 184
Interval B ................................ 184
Formations included ................. 184
Upper boundary of interval B ......... 184
Thickness trends ....................... 184
Lithofacies trends ..................... 184
Sources and environments of deposition 185
Paleotectonic implications .......... 185
Interval C ................................ 186
Formations included ................. 186
Age of interval C ....................... 186
Interval C—Continued
Upper boundary of interval C ...... 187
Thickness trends ....................... 187
Lithofacies trends ..................... 187
Sources and environments of deposition 188
Paleotectonic implications .......... 188
Ouachita orogeny ........................ 188
Interval D ................................ 189
Formations included ................. 189
Upper boundary of interval D ......... 189
Thickness trends ....................... 190
Lithofacies trends ..................... 190
Sources and environments of deposition 190
Paleotectonic implications .......... 190
Interval E ................................ 191
Formations included ................. 191
Upper boundary of interval E ......... 191
Thickness trends ....................... 191
Lithofacies trends ..................... 191
Sources and environments of deposition 192
Paleotectonic implications .......... 192
Total thickness of Pennsylvanian rocks 192
Thickness trends ....................... 192
Paleotectonic implications .......... 193
Geologic units directly above Pennsylvanian System 193
Units overlying Pennsylvanian ....... 193
Paleotectonic implications .......... 193
References ............................. 193

ILLUSTRATIONS

For listing of plates (in separate case) see volume “Contents”

Figure 40. Map showing geographic features in the Texas Panhandle and Oklahoma mentioned in text ....................... 178
41. Map showing Pennsylvanian structural features in the Texas Panhandle and Oklahoma mentioned in text 179
42. Map showing Pennsylvanian intervals and subintervals at base of Pennsylvanian System in the Texas Panhandle and Oklahoma 180

TABLES

Table 22. Rock-stratigraphic units included in interval C in northeastern Oklahoma ....................................................... 186
23. Rock-stratigraphic units included in interval D in eastern Oklahoma ................................................................. 189
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

TEXAS PANHANDLE AND OKLAHOMA

By SHERWOOD E. FREZON and GEORGE H. DIXON

ABSTRACT

Rocks of Pennsylvanian age in the Texas Panhandle and Oklahoma region overlie rocks as young as latest Mississippian (Chester) and as old as Precambrian. Rocks older than the Mississippian are present beneath the Pennsylvanian along the trend of the Amarillo, Wichita, and Criner uplifts and along a trend extending from the Arbuckle uplift northward to the Kansas border along the Nemaha anticline.

The Pennsylvanian System in the region includes rocks assigned to five intervals that correspond approximately to the five provincial series of the system in the midcontinent region. The series and their approximate interval equivalents are: Morrow, interval A; Atoka, interval B; Des Moines, interval C; Missouri, interval D; and Virgil, interval E. The lower part of interval A is treated separately at some places and called subinterval A₁. Each series has a diagnostic fusulinid fauna.

The depositional history of the Pennsylvanian can be interpreted from the thickness, distribution, and lithology of rocks of each interval. During Morrow, Atoka, and early Des Moines time (intervals A, B, and early C) the sea gradually encroached over land areas. Lower Morrow deposits were mainly restricted to an arcuate area that included the Ouachita geosyncline and the Ardmore and Anadarko basins. Upper Morrow sediments overlapped lower Morrow deposits northward in eastern Oklahoma and the Texas Panhandle. Upper Morrow sediments also were deposited in the Palo Duro, Hardeman, and Hollis basins. Atoka sediments were largely coextensive with areas of earlier Pennsylvanian sediments within the region.

In most areas the Pennsylvanian sea reached its maximum extent early in Des Moines time, when lower interval C was deposited over all the area except the Amarillo and Wichita uplifts. Apparently sedimentation during early Des Moines time was cyclic. In northern Oklahoma the cyclic sequence contains coal beds.

Although some tectonism occurred in the Ouachita region as early as late Morrow time, extensive uplift and folding were not apparent in eastern Oklahoma until the Ouachita orogeny in the early Des Moines (interval C). Tectonic movements in this area continued throughout the remainder of the Pennsylvanian. The Arbuckle uplift also became a positive area during the Des Moines.

The western part of the region was the site of marine sedimentation during late Des Moines, Missouri, and Virgil time. In the northeast, cyclic deposits containing coal beds were deposited through late Des Moines and into Missouri time. In most areas of deposition the younger sediments became increasingly calcareous.

Except in eastern Oklahoma, deposition of marine rocks continued from the Late Pennsylvanian into the Permian.

REGIONAL FEATURES

The region described in this chapter includes all of Oklahoma and the panhandle of Texas (fig. 40). Since the turn of the century more than 100,000 wells have been drilled in the region in the search for oil and gas, and stratigraphic data from a representative group of these wells provide most of the data that we used.

The present concepts of Pennsylvanian stratigraphy for the region have been developed in many papers, and the abundant subsurface data have been the major source of information used in these papers.

Structural features that were active within this region during Pennsylvanian time are shown in figure 41. The most prominent positive trend—the Amarillo, Wichita, and Criner uplifts—extends from southern Moore County, Tex., southeastward for more than 300 miles into Love County, Okla. A hyphenated combination of Amarillo, Wichita, and Criner is used when the whole trend is referred to; when only part of the trend is discussed the name for that part is used. The Muenster arch is a southeast extension of the Wichita-Criner uplift and is separated from the Criner uplift by the Marietta basin (fig. 41). Other negative elements around the Amarillo-Wichita-Criner uplift include, on the north, the Anadarko and Ardmore basins and, on the south, the Palo Duro, Hollis, and Hardeman basins (fig. 41).

The Matador arch in the southwestern part of the region forms the south edge of the Palo Duro basin. The Cimarron uplift and the Keyes dome extend northward from the west end of the Amarillo uplift. The Dalhart basin lies west of the Cimarron uplift and east of a positive area that includes the Bravo dome.

The Nemaha anticline, a subsurface structure extending from the area of Cleveland County, Okla., northward through Kansas, was the highest part of an extensive positive area in northern Oklahoma. The entire positive area is here called the Nemaha highlands.
In Early Pennsylvanian time the east-trending Ouachita geosyncline and its northern shelf occupied the eastern part of the region. In Middle Pennsylvanian time the southern part of the geosyncline developed into the Ouachita Mountains. The northern part of the geosyncline and part of the northern shelf retained their structural configuration and have been designated as the Arkoma basin. The Ozark uplist, mainly northeast of Oklahoma, is an older structural feature that had been submerged during Early Pennsylvanian but was rejuvenated at the time that the Ouachita Mountains developed.

The Arbuckle uplist in south-central Oklahoma became a positive feature at about the same time as the Ouachita Mountains.

Two areas of study (fig. 40) are combined in the present regional discussion. The western part of the region is the responsibility of Dixon; the eastern part, that of Frezon.

**PALEOGEOLOGY**

**UNITS UNDERLYING PENNSYLVANIAN**

Rocks directly beneath the Pennsylvanian in most of the region are Mississippian (Chester, Meramec, or Osage). The distribution of these rocks is shown on plate 2. The Oklahoma part of this map is modified from a map by Jordan, Bellis, and Rowland (1962). Where rocks older than Mississippian directly underlie the Pennsylvanian, paleostructural highs are indicated. On the Amarillo uplift the oldest rocks directly beneath the Pennsylvanian are Precambrian, and to the east they are Cambrian. Along the Criner segment of the uplift east of Stephens County, Okla., rocks beneath the Pennsylvanian are Cambrian and Ordovician (Ham and others, 1964). Between the Arbuckle Mountains and southern Logan County, Okla., strata beneath the Pennsylvanian range in age from Mississippian and Devonian to Cambrian and Ordovician. Along the Nemaha anticline from southern Logan County northward, rocks as old as Cambrian and Ordovician underlie the Pennsylvanian on a series of isolated highs.

In much of north-central Oklahoma Osage rocks underlie Meramec rocks. The southern limit of the Osage strata cannot be accurately determined from available information. The Osage is assumed to be present on all highs along the Nemaha anticline northward from northern Logan County, but the scale of plate 2 does not permit differentiating rocks of Osage age on these highs.

Small areas in which Precambrian rocks underlie Pennsylvanian are present both northeast and south of the Palo Duro basin (Totten, 1956). These Precambrian rocks are present in local truncated folds of pre-Pennsylvanian age. The largest fold, on the northeast side of the basin, in Hall and Donley Counties, Tex., may be related to the Amarillo-Wichita uplist. The four folds on the south side of the basin are local highs on the Matador arch.

**LOWER BOUNDARY OF PENNSYLVANIAN**

Because of onlap the base of Pennsylvanian strata is not of the same age everywhere in Oklahoma. The distribution of the various units at the base of the Pennsylvanian is shown in figure 42.

The oldest Pennsylvanian rocks in most of the mid-
continent region are of Morrow age. The base of the Morrow Series in northeastern Oklahoma corresponds to a rock-stratigraphic boundary at the base of the Hale Formation (pl. 13). Generally, the basal unit of the Morrow in the outcrop area is equivalent to the Prairie Grove Member of the Hale Formation, but locally the Morrow includes older mudstone equivalent to the upper part of the Cane Hill Member of the Hale. These rocks are unconformably underlain by rocks of Chester age.

In other parts of Oklahoma, where the Springer Formation (Pennsylvanian and Mississippian) is present, the time-stratigraphic boundary separating Mississippian from Pennsylvanian is within the lower part of the Springer and does not correspond to the formation boundary. For convenience in mapping, all of the Springer Formation is herein included in interval A because the precise base of the Pennsylvanian within the Springer cannot be ascertained. The Springer is underlain by rocks of Chester (Mississippian) age stratigraphically equivalent to the Pitkin Limestone.

The distinction between shale of Early Pennsylvanian age and older gray shale units is subtle. Morrow or Springer shales are noncalcaceous and commonly include thin beds of sideritic mudstone; locally they contain thin beds of fine-grained sandstone. Older, Mississippian shales are generally calcareous and locally have thin interbeds of limestone. Although the older shales are also dark gray, they are generally recognizable by a brownish cast and a slightly brownish streak.

In areas where Pennsylvanian rocks younger than Springer or early Morrow overlie Mississippian rocks little or no difficulty is encountered in determining the base of the Pennsylvanian. In these areas the younger, Pennsylvanian detrital rocks are coarser grained than the Mississippian rocks, and the contact is an unconformity.

In this report the base of the Pennsylvanian in the Ouachita Mountains area is placed stratigraphically lower than it is currently placed by many workers. This problem is discussed in the section concerning interval A.

PALEOTECTONIC IMPLICATIONS

At the close of the Mississippian a general uplift resulted in low-lying land areas and restricted seas. Seas continued to exist and marine sedimentation persisted in an elongate embayment that included the Ouachita geosyncline, the Ardmore basin, and the Anadarko basin. Low-lying land areas to the north, west, and south of the restricted seas probably were sources of the Lower Pennsylvanian fine detrital sediments.

Deposition probably was continuous in the Ardmore and Anadarko basins and in the Ouachita geosyncline from Mississippian into Pennsylvanian time. Thus, pre-Springer Mississippian and Springer sediments are similar.

INTERVAL A
FORMATIONS INCLUDED

For the most part interval A in the Oklahoma and Texas Panhandle region is the rock unit that cor-
responds approximately to rocks of the Morrow Series in the midcontinent region. In Pennsylvanian rocks Millerella is a long-ranging fusulinid genus, but it is the only fusulinid genus generally found in rocks of Morrow age (Wanless, 1963, p. 26).

Formations assigned to interval A for the region are shown on plate 13. In northeastern Oklahoma, outcropping rocks of interval A are the Boyd and Hale Formations, lateral equivalents of the same formations at their type sections in western Arkansas, where they are part of the Morrow Series. The interval includes the Wapanucka Limestone and the Springer Formation in the northern Arbuckle Mountains and the Golf Course Formation and the Springer Formation in the southern part of the Arbuckle Mountains and the Ardmore basin. Also included in the interval are the lower part of the Dornick Hills Formation and the Springer Formation in the area west and south of the Arbuckle Mountains.

Despite the presence in the Springer of some beds of youngest Mississippian age, all the Springer of the Arbuckle and Ouachita Mountains is included in interval A for practical reasons that will be discussed. Also, the Springer has been separated from the rest of interval A and is treated in a separate series of maps as subinterval A1. The stratigraphic reasons for this procedure are discussed in the section entitled "Subinterval A1."

In the subsurface, interval A includes rock-stratigraphic equivalents of the Springer. Only in central Oklahoma are Springer and rocks of Morrow age differentiated in the subsurface, in the panhandles of both States interval A generally is not divided into units.

In the Ardmore and Anadarko basins the Mississippian Goddard Shale Member at the base of the Springer may be included in the interval at localities where the Goddard cannot be clearly distinguished from the overlying part of the Springer with available data.

Two sets of stratigraphic nomenclature are used for the rocks that crop out in the Ouachita Mountains. In the frontal Ouachita Mountains, the area of extreme
overthrusting along the north edge of the mountains, the Wapanucka Limestone and the Springer Formation are included in interval A. In the central Ouachita Mountains; south of the frontal zone, the Johns Valley Shale and the Jackfork Sandstone are included.

Many workers currently place the lower boundary of the Pennsylvanian within the Johns Valley. This usage is in accordance with the fossil age determinations made by Cline (1956, 1960). On the other hand, Gordon and Stone (1969) believe, on the basis of fossil evidence, that the Johns Valley is entirely Pennsylvanian because the underlying Jackfork Sandstone contains Early Pennsylvanian and possibly Late Mississippian fossils. The age of the Johns Valley and the Jackfork have not been established unequivocally. The Stanley Shale, which underlies the Jackfork, has a tuff bed (Hatton Tuff Lentil) at the base that may be Pennsylvanian in age according to radiometric studies made by Mose (1969); but on other evidence, the Stanley is still considered to be Mississippian in age by many geologists and, therefore, is not mapped with Pennsylvanian rocks in this publication.

**UPPER BOUNDARY OF INTERVAL A**

Rocks of interval A are overlain by the Atoka Formation (interval B) and equivalent units or, where the Atoka is absent, by rocks of the Des Moines Series (interval C). In most of eastern Oklahoma relatively coarse sandstone at the base of the Atoka Formation is easily distinguished from underlying limestone, calcareous mudstone, and calcareous, glauconitic sandstone of the Morrow.

In northwestern Mayes County and northeastern Rogers County, Okla., (fig. 40), calcareous, glauconitic sandstone of the Morrow (interval A) is readily separated from overlying noncalcareous, nonglauconitic sandstone and noncalcareous clay mudstone of the Krebs Group (interval C). In the frontal Ouachita Mountains, the western part of the Arkoma basin, and the Arbuckle Mountain area, the top of the Morrow coincides with the top of the distinctive thick widespread Wapanucka Limestone. In the Ardmore basin the Otterville Limestone Member at the top of the Golf Course Formation marks the top of the Morrow.

In the Anadarko basin in western Oklahoma, the Morrow is overlain by dark shaly mudstones and some thin interbeds of sandstone of the Atoka; the lower part of the Atoka contains a few coal beds. The Morrow rocks consist of calcareous mudstone, limestone, and calcareous sandstone; glauconite is common in both limestone and sandstone.

In the Oklahoma Panhandle and the northern part of the Texas Panhandle, the base of the 13 finger limestone of subsurface usage is also the top of the Morrow (Cunningham, 1961). Farther south, in the Palo Duro basin, the top of interval A is determined by lithic criteria similar to those used in the Anadarko basin of western Oklahoma. In the Palo Duro basin the Atoka has a slightly greater percentage of limestone than in western Oklahoma; the darker mudstone and the coal beds are the most useful criteria for distinguishing the unit from underlying rocks.

**THICKNESS TRENDS**

The maximum thickness of interval A (pl. 3A) is at least 8,900 feet in the Ardmore basin, 1,485 feet in the Arkoma basin, 6,800 feet in the Anadarko basin, and about 500 feet in the Palo Duro basin.

Interval A thickens southward in both eastern and western Oklahoma from an eroded edge to a maximum in the Anadarko and Arkoma basins. The south edges of these basin areas coincide with the faulted north edges of the Wichita-Criner and the Ouachita structural provinces, which also mark the eroded edge of interval A. Information is not available concerning the maximum depositional thickness or the original southern extent of these rocks, but probably the greater thickness in the Ardmore basin indicates a former continuation of the southward thickening. The present distribution of thickness in the Palo Duro basin probably is essentially that of the original basin of deposition.

**LITHOFACIES TRENDS**

Where interval A is thickest in Oklahoma and in the Texas Panhandle, clay mudstone is strikingly dominant among the facies shown on plate 3B and on plate 10D, section W-W'. Northward from the Arkoma basin the sandstone and limestone content of interval A increases as the total thickness decreases. Sandstone and limestone increase also in the Anadarko basin southward, westward, and, to a lesser degree, northward from a center of maximum thickness.

In the basin area southwest of the Amariello-Wichita uplift, interval A contains a higher percentage of limestone and fine detrital rock in the center than on the periphery of the basin, where sandstone increases in abundance. Deposits in the southern part of the Ouachita geosyncline consist of mudstone and arkosic sandstone of the Jackfork Sandstone; because of inadequate data, lithofacies for this area are not shown on plate 3B.

The boulder beds of the Johns Valley Shale are unique for the Pennsylvanian System in Oklahoma. Although the distribution of this formation is not shown on any of the maps, the lithology has great paleotectonic significance. The beds contain exotic clasts that range in size from sand to blocks on the order of several hundred feet in their greatest dimension, enclosed in a matrix as fine as mudstone. The clasts range in age from St. Croixan (Late Cambrian) to Morrow (Early Pennsylvanian) and came from source areas in the Ozark, Ouachita, and Arbuckle uplift regions (Shideler, 1970).
SOURCES AND ENVIRONMENTS OF DEPOSITION

The Springer Formation and equivalent rocks were deposited in an arcurate area across southern Oklahoma. Deposits in the Arkoma basin now isolated from those of other areas indicate the former greater extent of these rocks.

Sources of interval A detritus probably were, in part, low-lying land areas. The Nemaha highlands, a low positive area, contributed fine sediments southwestward into the Anadarko basin and southeastward into the Arkoma basin. The Ozark uplift apparently was the source of the slightly coarser, more sandy deposits in the Arkoma basin.

The sources of sand in the Springer Formation of the Ardmore basin possibly were the Wichita-Amarillo uplift and the Muenster arch (Tomlinson and McBee, 1959, p. 13-14). Southern and southwestern sources for these sediments also seem probable, even though the southern extent of the basin of deposition is not known. Cross section W-W"(pl. 10D) indicates that the Keyes dome was a sand source for the low areas north and south of the dome.

In all source areas the sediments were derived initially from rocks of Mississippian age, exposed at the surface early in interval A time. The exclusively fine grain size of these detrital source rocks probably was the controlling factor in determining the texture of deposits forming interval A, although low topographic relief, with resulting low-energy transport, was a contributing factor.

Sandstone in the Jackfork had both a northern source, possibly the Ozark uplift, which supplied quartz from older sedimentary rock, and a southern source, east of the present outcrop belt, which supplied feldspar from igneous and metamorphic sources (Klein, 1966, p. 316, 318). These sands were mixed in the deep part of the geosyncline and were dispersed to the west by currents along its axis.

Outside the Ouachita geosyncline, sediment in the lower part of interval A was deposited in a rather specialized chemical environment, very different from that of rocks above and below. Mud, silt, and fine sand apparently were carried into seas that were shallow and brackish. Influxes of iron-bearing solutions into the slightly reducing environment caused precipitation of iron carbonate which mixed with the muds to form "ironstone" beds of the Springer.

Detritus in the upper part of interval A is generally coarser and more abundant than that in the Springer, suggesting that greater quantities of coarse-grained material were eroded in the source areas. Probably the land areas were more restricted and sedimentation was more extensive during deposition of upper interval A than before; the centers of deposition, however, were located in about the same places.

The seas in which sediments of the upper part of interval A were deposited apparently had slightly higher pH and Eh than did the earlier, Springer Seas, and the environment must have been sufficiently rich in oxygen to support lime-secreting invertebrate faunas. Glauconite, characteristic of the sandstone in the upper part of interval A, probably formed diagenetically in an anaerobic environment after burial of the sediments.

PALEOTECTONIC IMPLICATIONS

The Amarillo-Wichita-Criner uplift was active during much of the Pennsylvanian, but which parts were most active during the times represented by each of the intervals is difficult to document. Periods of active deformation were discussed by Harlton (1951, 1963); Chase, Frederickson, and Ham (1956); Totten (1956); Adams (1962); Riggs (1957); Edwards (1959); Tomlinson and McBee (1959); Eddleman (1961); Rascoe (1962); Hills (1963); and Ham, Denison, and Merritt (1964).

The lower part of interval A, designated as subinterval A₁, is not present on the central block of the Amarillo-Wichita-Criner uplift, but it is present on immediately adjacent blocks to the north and is involved in the block-faulted structure. Whether the central block was once covered by subinterval A₂ or was a source area for deposits on the flanks of the uplift at that time is not fully established. Distribution and character of conglomerates in the Criner segment of the uplift indicate that movement may have started as early as Springer time (Tomlinson and McBee, 1959, p. 14-15), and if so, the central block was a source area. The oldest conglomerates near the Criner uplift are in strata of late Springer age, and overlying conglomerates of Morrow age contain cobbles and pebbles of Mississippian and Devonian rocks. In the area from which the clasts were derived, the surface, according to Tomlinson and McBee (1959, p. 15), "must have been nearly a mile higher than it was beneath the area where they came to rest."

Northwestward along the uplift some of the rocks in the upper part of interval A are arkosic. Thus, removal of Paleozoic strata from the uplift and exposure of basement rocks is indicated. The western part of the uplift probably was eroded before the eastern part.

Deposition was restricted to a small part of the Anadarko basin in Springer time. Subsidence continued, and by Morrow time the area of deposition expanded northwestward across the panhandle areas. The Amarillo-Wichita-Criner uplift continued as a positive area, and evidence is not available to indicate that the sea crossed it. During Springer deposition, the Palo Duro basin area also was positive, but during Morrow deposition, the area subsided to form the westernmost
end of a shallow marine basin of deposition similar to the generally expanding basin on the north side of the uplift.

The Nemaha highlands were positive, but continuing subsidence in the basins to the southeast in late Morrow time caused the seas to lap onto the margins of the uplift. The highlands probably were low except where the Nemaha anticline formed a line of hills or low mountains extending northward through the central part.

The west end of the Ozark uplift was not positive in this region during interval A time. Sandstone in the lower part of the interval suggests that some part of the uplift may have been a source for sediments, but if so, this source was probably east of Oklahoma.

The Ouachita uplift region apparently was a subsiding trough until late Morrow time, when tectonism produced source areas along the north margin of the trough (Shideler, 1970) and structures in the basin areas.

In the Ardmore basin, sediments accumulated without any apparent interruption during the interval.

SUBINTERVAL A₁ FORMATIONS INCLUDED

In most areas of the country the base of the Pennsylvanian System corresponds to the base of a rock-stratigraphic unit. Along the Amarillo-Wichita-Criner uplift, in the Ouachita tectonic system in the southern United States, and along the Appalachian system in the eastern United States a rock-stratigraphic unit straddles this time-stratigraphic boundary. This rock-stratigraphic unit, which is considered part of interval A, is differentiated from the rest of interval A on a separate series of maps as subinterval A₁ (pl. 4). The relations between the rocks of subinterval A₁ and the oldest Pennsylvanian rocks of Morrow age have been the subject of controversy and misunderstanding, but increased stratigraphic data permit better understanding of the problems.

In the area of the Arbuckle Mountains, northward from the Arbuckle Mountains in the Anadarko basin, and in the Ouachita Mountains the rocks in subinterval A₁ belong to the Springer Formation. In the Arkoma basin and in northeastern Oklahoma the lower or Cane Hill Member of the Hale Formation is equivalent to the Springer of the northern part of the Arbuckle uplift (Frezon, 1962) and is also included in subinterval A₁.

UPPER BOUNDARY OF SUBINTERVAL A₁

In the Ardmore basin outcrops in southern Oklahoma, subinterval A₁ (Springer Formation) is overlain by the Primrose Sandstone Member of the lower part of the Dornick Hills Formation. At the surface north and east of the Arbuckle uplift in Pontotoc, Coal, and Atoka Counties, the Union Valley Sandstone Member of the Wapanucka Limestone overlies the Springer. In northeastern Oklahoma the upper boundary of the subinterval is the contact between the Cane Hill and the overlying Prairie Grove Member of the Hale Formation (Frezon, 1962, p. 26).

In many areas in Oklahoma, beds equivalent to the Springer are not overlain by readily recognizable sandstone units such as the Union Valley, Prairie Grove, or Primrose Member. These areas include surface exposures in southeastern Pontotoc and Coal Counties (Kuhleman, 1951) and subsurface strata of western Coal County (Frezon, 1962, p. 26). Nevertheless, the rocks that do overlie the Springer—the upper part of the Morrow Series—are distinctly different from Springer rocks, which are sideritic and noncalcareous.

THICKNESS TRENDS

In the Anadarko basin, subinterval A₁ ranges in thickness from a thin edge to more than 4,600 feet (pl. 4A); the maximum occurs in Caddo County, adjacent to faults that mark the front of the Wichita uplift. In the Ardmore basin, subinterval A₁ is more than 8,900 feet thick, and in the southwest end of the Arkoma basin in Atoka County, Okla., it is more than 900 feet thick.

In both the Anadarko and Arkoma basins, rocks of subinterval A₁ are progressively thinner toward the north and northwest on the shelf area and terminate as a thin edge. In these basins the subinterval is everywhere covered by younger rocks of interval A.

In the frontal Ouachita Mountains, subinterval A₁ is estimated to be about 2,500 feet thick (Goldstein, 1961, p. 38). This estimate shows the magnitude of thickness, but thicknesses for specific localities are not shown on the map (pl. 4A). Farther south, a few wells indicate thicknesses of more than 6,000 feet.

The edge of Springer rocks in eastern Oklahoma probably is erosional, but in western Oklahoma it may be the result of erosion and depositional thinning. In Muskogee and McIntosh Counties, Okla., rocks of this subinterval are locally absent.

North of the general limit of subinterval A₁, outliers attest to an originally more extensive distribution of these rocks. In one such outlier, in Muskogee County (pl. 4A), 45 feet of the unit is present. Another outlier, not shown on the map, occurs at the west end of Fort Gibson Dam in Wagoner County, Okla. (fig. 40). In this outcrop the unit is 3-10 feet thick and consists of black noncalcareous shaly mudstone and a basal pebble conglomerate that unconformably overlies exposed Pitkin Limestone (Chester). The unit probably is a remnant of the Cane Hill Member of the Hale Formation (S. E. Frezon, unpub. data) and is overlain unconformably by limestone of the Prairie Grove Member of the Hale For-
mation, which in some places is conglomeratic at the base. Henbest (1953) described a similar unconformity at the top of the Cane Hill in its type locality in Washington County, Ark. (fig. 37).

In both the Anadarko and Arkoma basins, subinterval A; thickens southward; but in neither basin can the original southward extent of the interval be determined. In the Anadarko basin the southward thickening trend is abruptly interrupted by the faulting that marks the north edge of the Wichita-Criner uplift; south of this uplift the subinterval is not present.

In the Ardmore basin subinterval A; is overlain by younger rocks of interval A, and its thickness is consistently greater than in the Anadarko and Arkoma basin areas. The greater thickness in the Ardmore basin probably more closely represents original thickness than do the present maximums in the Anadarko and Arkoma basins, where more postdepositional erosion took place.

LITHOFAENCIES TRENDS

Throughout much of the region, subinterval A; consists predominantly of mudstone (pl. 4B) and generally contains less than 20 percent sandstone; nowhere does the subinterval contain more than 5 percent limestone.

In the Arkoma basin, where the total thickness of the unit is less than 600 feet, sandstone-mudstone ratios range from greater than 4:1 in small areas to less than 1:4 in large areas (pl. 4B). The same or greater amounts of sandstone in parts of the Ardmore and Anadarko basins are not reflected in the facies patterns because the sandstone forms much less than half the subinterval, which is dominantly mudstone.

INTERVAL B
FORMATIONS INCLUDED

Interval B is the rock unit equivalent to or approximating the Atoka Series of the midcontinent region. Rocks included in the Atoka Series contain the Profusulinella and Fusulinella fusulinid zones (Wanless, 1963, p. 26).

Interval B in the Texas Panhandle and Oklahoma region includes only the Atoka Formation and equivalent strata (pl. 15). In the area of the Arbuckle Mountains, equivalent rocks assigned to interval B are the Lake Murray Formation and the upper part of the Dornick Hills Formation.

UPPER BOUNDARY OF INTERVAL B

The upper boundary of interval B is placed at the base of light-colored mudstone in overlying interval C (Des Moines). In the eastern part of the area numerous coal-bearing cycles occur in the lower part of the Des Moines Series but not in underlying rocks. Because changes between rocks of Atoka age and the overlying rocks are gradational, the boundary in many places is difficult to determine.

In eastern Oklahoma the base of the Hartshorne Sandstone or, where the Hartshorne is absent, the base of the McAlester Formation is the upper boundary of interval B. These two formations are readily traceable from exposures into the subsurface.

In the Ardmore basin the top of the Lake Murray Formation is placed at the top of the Frensley Limestone Member; this lithology is easily distinguished from overlying detrital rocks.

In the Anadarko basin the Atoka is difficult to differentiate from overlying rocks. Lithic descriptions indicate that the gray and black mudstones of the Atoka are similar in appearance to the mudstones of the overlying Krebs and Cabaniss Groups of Des Moines age (Adkison and Sheldon, 1963). The age of the Krebs is discussed under "Age of Interval C."

Westward, in the Texas and Oklahoma Panhandle areas, the top of the 13 finger limestone of subsurface usage is considered the top of interval B. The status of information and opinion concerning Atoka-Des Moines relations in western Oklahoma and in the two panhandle areas was discussed by Rascoe (1962, p. 1353-1354). In the Palo Duro basin, rocks assigned to interval B are similar to those in the area north of the Amarillo uplift. They consist of a series of sandstone and dark-gray mudstone beds and a few limestone beds. In this area, as in others discussed, the top of the unit is not well defined.

THICKNESS TRENDS

The thickness of interval B (Atoka) is shown on plate 6A. The maximum thickness is in the Arkoma basin, where more than 10,000 feet is present near the east edge of Oklahoma. In the Anadarko basin the maximum thickness is about 5,000 feet; in the Ardmore basin it is about 2,500 feet, and in the Palo Duro basin, 500 feet.

Between the margin of deposits on the north and the limiting structural elements on the south (Amarillo-Wichita-Criner uplift and the Ouachita uplift) a southward thickening occurs (pl. 10D, sec. W-W'). In eastern Oklahoma this trend is interrupted by faults within the northern part of the Ouachita uplift. In western Oklahoma and in the panhandle areas, the thickness reaches a maximum along the axis of the Anadarko basin; southward from the axis the thickness of interval B decreases toward the south edge of the basin.

Deformation and erosion mask original thicknesses and depositional trends in the Ardmore basin. In the Palo Duro, Hardeman, and Hollis basin areas the rocks are thickest toward the centers of the basins, where they are preserved in their originally deposited thicknesses.

LITHOFAENCIES TRENDS

Lithofacies for interval B (Atoka) are shown on plate 6B. Thick sections in the Anadarko and Arkoma basins
are dominated by fine-grained detrital rocks; toward shelf areas to the north the rocks are more calcareous and coarser grained. Semicircular lithofacies bands along the western limits of interval B in Seminole and Hughes Counties, Okla., apparently have no depositional significance. In this area the lower part of the interval is sandy, but the upper part is composed dominantly of mudstone. Thus, the thinner sections near the limits of the formation where the upper part of the interval is eroded away have a higher percentage of sandstone.

In the Anadarko and Ardmore basins, rocks of interval B are dominantly mudstone. From the Anadarko basin northward into the panhandle area, the percentage of limestone in the section increases; this is the area in which the Atoka is represented by the 13 finger limestone. Across the area of the Cimarron uplift and its northward extension, the average size of detrital particles increases, but farther west, in the Dalhart basin, sections are composed dominantly of fine-grained detrital rock interbedded with limestone. These rocks are similar to those east of the Keyes dome and the Cimarron uplift. In the Palo Duro basin, rocks of this interval are dominantly mudstone interbedded with various amounts of limestone. In the Hollis and Hardeman basins coarser detrital rocks are dominant and are associated with a smaller percentage of limestone.

Near the Amarillo-Wichita-Criner uplift, conglomerates are present in interval B. In the Ardmore basin the Bostwick Member occurs at the base of the Lake Murray Formation and is considered the base of interval B. This conglomerate contains pebbles and cobbles of formations as old as late Arbuckle (Early Ordovician) age, and their distribution suggests that the Criner uplift was their source (Tomlinson and McBee, 1959).

Along the Amarillo-Wichita part of the uplift, coarse detrital sediments—the so-called carbonate and granite washes—were deposited in considerable volume. Their distribution and stratigraphic relations in the western part of Oklahoma were described by Edwards (1959).

The conglomerate and sand-sized wash containing virtually unaltered carbonate and feldspar grains indicate that chemical weathering was minimal during degradation of the uplifts and production of sediment for the basins.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Sources of sediments in interval B in the Texas Panhandle and Oklahoma region were several positive areas, all of them near or within the region, except one to the east. The Amarillo-Wichita-Criner uplift was positive throughout interval B time, and sediments from this uplift were deposited in the Hardeman, Hollis, Palo Duro, and Marietta basins to the south and in the Anadarko and Ardmore basins to the north. A low land area to the south was a source of sediments in the Hardeman and Hollis basins; the same area contributed detritus to the Palo Duro basin. The Bravo dome northwest of the Palo Duro basin may have been a source area, and a low-lying island area in the Palo Duro was also a local sediment source.

The Nemaha highlands and the Nemaha anticline contributed sediments to the Anadarko and Ardmore basin areas and to the shelf north and west of the Ouachita geosyncline. Eastern, southeastern, and southern sources, though, furnished most of the sediment in the eastern part of Oklahoma (Scull and others, 1959). In eastern Oklahoma, gradual subsidence and enlargement of the geosyncline resulted in a sequence of sediments that overlapped the northern and western shelf areas of the basin. The Anadarko and Ardmore basins as well as the Ouachita geosyncline apparently were marine basins in which subsidence and deposition were nearly in balance.

In Oklahoma and the panhandle of Texas all deposition apparently took place in a marine environment, and the thickness was controlled by the amount of detrital sediments available for deposition. In basin areas, where detrital sediment dominated, carbonate deposition was negligible, and in these areas swamps developed when basin filling reached temporary maxima. In swamps localized in the Anadarko, Palo Duro, Hardeman, and Arkoma basins coal formed with irregular distribution.

In a few areas, as in the Ardmore basin and on the shelf area northwest of the Anadarko basin where the 13 finger limestone is present, lime-secreting invertebrate animals flourished.

Along the southeast edge of the Arkoma basin in Atoka County, Okla., and in the Ouachita Mountains immediately southeast of that area the Atoka Formation is locally conglomeratic. The conglomerates contain pebbles as much as 4 inches in diameter. These pebbles include cherts from unknown sources as well as Ouachita facies rocks such as spiculites of Morrow age and phosphate pebbles from the Caney Shale of Mississippian age (Hendricks and others, 1947). Because these conglomerates are only local in their distribution they are not quantitatively an important lithic constituent of the formation. Nevertheless, they are important in that they indicate a nearby source area to the south or east.

PALEOTECTONIC IMPLICATIONS

Structural uplift along the Amarillo-Wichita-Criner axis occurred early during deposition of interval B; this activity is referred to as the Wichita orogeny (Tomlinson and McBee, 1959). Although the entire uplift was
positive during interval B time, the Criner segment apparently was uplifted more strongly than the area farther west. Erosion of the Criner uplift resulted in deposition of the Bostwick Member in the lower part of the Dornick Hills Formation. In other rocks of interval B, both along the uplift to the west and around the Nemaha highlands, the absence of conglomerate suggests that those areas had relatively low topographic and structural relief.

The depositional axis of the Ouachita geosyncline was south of present limits of the interval. Postdepositional uplift and erosion of deposits occurred during the Ouachita orogeny. The implications of the Atoka conglomerates and their relation to the tectonism in the Ouachita area are discussed in the section concerning interval C.

In Craig County, in northeastern Oklahoma, outliers of the Atoka Formation (interval B) consist of fine- to medium-grained crossbedded sandstone that fills channels in underlying rocks of Mississippian age (Branson and others, 1965), indicating uplift and channeling before Atoka time.

The configuration of the Anadarko, Palo Duro, Hardeman, and Hollis basins was determined by folding during interval B.

**INTERVAL C FORMATIONS INCLUDED**

Rocks designated as interval C (Des Moines Series of the midcontinent region) crop out across a wide area in eastern Oklahoma, adjacent to the west edge of the Ozark uplift. These exposed rocks are divided into three groups that include 15 formations (table 22). In the Ardmore basin all rocks of interval C are included in the Deese Formation (pl. 13).

Formation names shown in table 22 are rarely used for subsurface studies in most areas. Rocks of interval C generally are designated by group names where the interval is divided; only in south-central Oklahoma are rocks of the interval referred to the Deese Formation. In western Oklahoma and in the panhandle areas, extensions of the Cabaniss and Krebs Groups are designated as an undivided unit or as the Cherokee Group. In the Texas Panhandle, equivalent rocks are generally identified as Des Moines in the western Anadarko basin or Strawn Group in basins to the south or west. Traceable lithic units, such as distinctive limestone and sandstone beds within formations, are recognized in the subsurface and are used as key beds within the group or series penetrated by a borehole.

Individually named formations recognized in northeastern Oklahoma generally are equivalent to rocks assigned to two or more named formations of Kansas or southwestern Missouri. Generally, formations thin northward in Oklahoma, and only relatively attenuated parts extend into Kansas or Missouri. In Oklahoma the Krebs and Cabaniss contain approximately 25 coal cycles (Branson, 1954). Some of the formation units to the north consist of single cycles.

**AGE OF INTERVAL C**

Rocks of the Des Moines Series include the *Fusulinella (Beedeina of current usage)* and *Wedekindellina* fusulinid zone (Willess, 1963, p. 27). In the east-central part of the United States, and especially in Illinois, these genera are restricted to rocks of Des Moines age; *Fusulinella* occurs only in the pre-Des Moines rocks. In areas where this mutually exclusive occurrence of genera exists, a ready differentiation of rocks of Des Moines and pre-Des Moines age is possible; but in Oklahoma such differentiation is precluded because of faunal distribution and taxonomic problems that are not present in the east.

Faunal distribution creates a problem in the type area of the Krebs Group in northeastern Oklahoma. The lower two formations of the group in that area do not contain fusulinids; the fauna in the upper two formations indicates an early but not earliest Des Moines age as recognized in the eastern areas. Thus, in northeastern Oklahoma the base of the Krebs Group (assigned an earliest Des Moines age) cannot be compared faunally with the base of the Des Moines in other areas.

The second problem involves taxonomy. Apparently a sequence of forms transitional from *Fusulinella* to *Fusulinella (Beedeina of current usage)* is present in the Ardmore basin (Waddell, 1966a, b), and this transitional sequence masks the precise position of the base of the Des Moines. Seven fusulinid biozones are recognized in the Ardmore basin (Waddell, 1966a, b). The lowest of these zones, Zone I, is distinguished by the genus *Fusulinella*. The *F. prolifica* fauna of the Atoka

| Table 22.—Rock-stratigraphic units included in interval C in northeastern Oklahoma |
|---------------------------------|--|
| **Group**                      | **Formation**               |
| Marmaton                       | Holdenville Shale           |
|                                | Lenapah Shale               |
|                                | Nowata Shale                |
|                                | Oologah Limestone           |
|                                | Labette Shale               |
|                                | Fort Scott Limestone        |
|                                | Wetumka Shale               |
|                                | Calvin Sandstone            |
| Cabaniss                       | Senora Formation            |
|                                | Stuart Shale                |
|                                | Thurman Sandstone           |
| Krebs                          | Boggy Formation             |
|                                | Savanna Formation           |
|                                | McAlester Formation         |
|                                | Hartshorne Sandstone        |
Formation of eastern Oklahoma is related to Waddell's Zone I fauna (Waddell, 1966b, p. 8). The overlying Zone II fauna in the Ardmore basin is characterized by "one of the most primitive Fusulina faunas to be found in the midcontinent region and by the presence of Wedekindellina" (Waddell, 1966b, p. 8). This zone, assigned an earliest Des Moines age, apparently is not represented in northeastern Oklahoma (Waddell, 1966b, p. 8). The next overlying zone (Zone III) is marked by two species of Wedekindellina and Fusulina (Beebeina of current usage) cf. F. novamexicana.

Fusulina (Beebeina of current usage) novamexicana is found in the Spaniard Limestone Member of the Savannah Formation in northeastern Oklahoma (Alexander, 1954). Though the only species found, it permits correlation of the Spaniard Limestone Member with Waddell's (1966b) Zone III in the Ardmore basin and with the Stonefort Limestone Member of the Spoon Formation in Illinois (Alexander, 1954, p. 49). The Stonefort, which contains B. novamexicana, among other forms (Dunbar and Henbest, 1942), lies above the base of the Des Moines in the Illinois section. Thus, the available data permit but do not prove an earliest Des Moines age assignment for the basal part of the Krebs Group.

All the formations of the Krebs Group are present throughout northeastern Oklahoma, although they thin northward from the area where their maximum thickness occurs, in the Arkoma basin. This thinning is apparently depositional rather than the result of truncation within the group.

**UPPER BOUNDARY OF INTERVAL C**

Interval C (Des Moines) is overlain unconformably by interval D (Missouri) throughout the Texas Panhandle and Oklahoma region, except beyond the eastern outcrop limits of interval D (pl. 8A). In most of southern Oklahoma the base of the Hoxbar Formation, which in some areas conformably and elsewhere unconformably overlies the Deese Formation, defines the top of interval C. In the Ardmore basin, the Confederate Limestone Member at the base of the Hoxbar Formation (pl. 13) overlies the Deese Formation (interval C). Between the Arbuckle Mountains and the Kansas border, sandstone at the base of the Seminole Formation (table 23) marks the top of interval C. The Arbuckle Limestones and the Jones sand of subsurface usage are considered equivalent to the basal sand of the Seminole and mark the top of interval C. In the subsurface where the Seminole is thin and the basal sand is absent, the base of the Checkerboard Limestone is used as the top of interval C.

In the Anadarko basin, limestone beds called informally the Big Lime in subsurface work are assigned to the Marmaton Group or upper part of interval C. Individual limestones in the Big Lime are, in ascending order, the Fort Scott, the Oologah, and the Lenapah (table 22). The lowest limestone, the Fort Scott, is the most extensive (Rascoe, 1962, p. 1357), but where the lowest is missing the next highest is used in determining the top of interval C.

**THICKNESS TRENDS**

The distribution and thickness of interval C are shown on plate 7A. Interval C is the oldest Pennsylvanian unit to extend over all the northern part of the region. The interval reaches a maximum thickness of 4,900 feet in the Ouachita geosyncline, but there the only part represented is the lower third of the interval, which is the part represented by the Krebs Group. The interval is 4,400 feet thick in the Anadarko basin.

In both the Ouachita and Anadarko areas interval C is bounded on the south by structural uplifts. In the Ouachita area the southern limit of rocks in the interval lies in a zone of intense folding immediately north of the Choctaw fault. Here the southern limit is the result of erosion that followed folding. In the Anadarko basin the greatest thickness occurs along an apparent axis within the basin, and the interval thins southward between this axis and the southern limit of the rocks at the faulted north edge of the Amarillo-Wichita uplift. Whether this thinning is mostly depositional or mostly erosional has not been determined.

On both the Cimarron uplift and the Keyes dome, interval C is less than 500 feet thick and is locally absent. The interval increases in thickness westward to more than 1,500 feet in the Dalhart basin.

Interval C is thin and locally absent in the Texas Panhandle but thickens eastward along the south side of the Wichita-Criner uplift to more than 7,400 feet in the Marietta basin. Thickness trends in the Marietta and Ardmore basins are much disturbed by faulting.

**LITHOFACIES TRENDS**

Lithofacies of interval C (Des Moines) rocks are presented on plate 7B. The rocks are dominantly detrital; mudstone is most abundant, and sandstone is generally subordinate. The proportion of carbonate rock is appreciable on shelf areas, where the interval is thin.

Details of the lithostratigraphic sequence cannot be shown by the generalized patterns used on plate 7B. In northeastern Oklahoma, for example, the seven formations of the Krebs and Cabaniss Groups contain 25 coal cycles (Branson, 1954, p. 5-6). In addition the overlying Marmaton Group in the area has at least one recognizable coal cycle in the Labette Shale.

The composition of the Marmaton Group, which is dominantly limestone, is masked on the lithofacies map. Wanless, Tubb, Gednetz, and Weiner (1963) have shown that limestones and associated beds in the Marmaton in northern Oklahoma are marine equivalents of coal-bearing sequences farther north and east.
The seas were more restricted during deposition of lower interval C than during the later stages of Atoka deposition. Regional subsidence was gradual, and deposition probably was virtually continuous in the basin areas during all of interval C time.

Positive areas that were sources of detritus for interval C (Des Moines) time were the Arbuckle and Wichita uplifts, the Muenster arch, the Nemaha highlands, and the Ouachita-Arkoma-Ozark area in the eastern part of the region.

The Nemaha highlands probably was not a major source of detrital sediments during deposition of the Cherokee Group (Cole, 1969). Most of the rocks exposed in the highlands during emergence were mudstones and carbonates of pre-Pennsylvanian age. Along the Nemaha antcline Ordovician and Cambrian sandstones were exposed locally, but these could not have furnished all the sand now in the lower part of interval C. During Krebs and Cabaniss deposition marine waters slowly covered the Nemaha highlands, and by the end of Cabaniss time the highland area was under water for the first time in the Pennsylvanian. This overall marine transgression permitted the development of the varied environments represented in the coal cycles described by Branson (1954). The encroachment process was oscillatory rather than continuous; the depositional history has been described most recently by Berg (1969) and Cole (1969). During most of Marmaton deposition the former highland area remained a submerged area of normal marine limestone sedimentation.

The presence of minor amounts of chert and feldspar in sandstone at different horizons within the interval indicates that the source area remained the same throughout the interval. Although some chert and feldspar may have been shed from the Nemaha antcline, most of the detritus came from the Muenster arch and the Wichita uplift, where chert-bearing Cambrian and Ordovician rocks and feldspar-bearing basement rocks were exposed.

The Arbuckle uplift was another source of detritus for interval C. Limestone pebbles of Silurian age and chert pebbles, probably of Cambrian and Ordovician age, were shed northward to form conglomerates at the base of the Thurman Sandstone in northern Coal, southeastern Hughes, and western Pittsburg Counties, Okla.

After deposition of the Krebs, the Ouachita-Arkoma-Ozark area became a major source of mud and sand. In eastern Oklahoma, where coal beds are present in interval C, the depositional environment fluctuated between normal marine and nonmarine. Over the rest of the area, deposition probably occurred in a marine environment throughout Des Moines time.

The major tectonic event in interval C time was the formation of the Ouachita uplift, discussed below. Other events were a slight reversal of movement in the region of the Nemaha highlands and rejuvenation of the Ozark uplift to the northeast.

In the Nemaha highlands area the cyclicity of the sediments suggests numerous episodes of subsidence, each followed by a brief period of quiescence. The overall effect of these numerous subsidences was a slow depression of the land area below sea level that carried all the highlands area below sea level by the end of Cabaniss time. During the remainder of interval C time slow subsidence in the area continued, never becoming so excessive that the region was depressed below depths where abundant marine faunas could survive.

The Amarillo-Wichita uplift continued to rise during deposition of interval C, at least in its western part. The east end of the positive element may have been submerged.

Uplift of the Arbuckle positive element is recorded by the conglomeratic Deese Formation and the equivalent Franks Conglomerate of former usage. Uplift of the Matador arch during interval C deposition does not appear to have been very great, although the lithofacies map indicates that some local areas shed fine-grained detritus. The Bravo dome in Hartley and Oldham Counties, Tex., and in adjacent New Mexico was possibly a positive feature in interval B and was more certainly uplifted during interval C time. Uplift also probably occurred along the Sierra Grande arch in northern New Mexico.

Ouachita Orogeny

In Middle and Late Pennsylvanian time uplift and intense folding and faulting designated as the Ouachita orogeny occurred in eastern Oklahoma and Arkansas. During the orogenic period the depositional area that had been the Ouachita geosyncline developed into two new structural elements, the Ouachita uplift to the south and the Arkoma basin to the north.

Uplift in the Ouachita region apparently started as early as late Morrow (interval A) time; uplift and, later compressive movements continued intermittently into Early Permian time. These movements caused folding and faulting in rocks as old as interval C (Des Moines) and fracturing in rocks as young as Early Permian. These structures occur within a broad region that includes the east half of Oklahoma.

Boulders in the Johns Valley Shale are the earliest evidence of tectonic activity in the Ouachita uplift in Pennsylvanian time. Various mechanisms have been cited to explain the mixture of boulders and shale in the Johns Valley (see Cline and Shelburne, 1959, p. 203-206, for a summary), but a juxtaposition of active tectonic
elements that produced submarine sliding seems the most likely explanation. The source of the boulders may have been a long arcuate positive element, near the present frontal Ouachita Mountains in eastern Oklahoma, that came into existence as a "rising geanticlinal ridge, fault-bounded on the south, which became the locus of an emergent archipelago" (Shideler, 1970, p. 805).

Early uplift within the Ouachita region might also account for conglomerates that are found in the Atoka Formation in Atoka County, Okla. (Knechtel, 1937; Hendricks and others, 1947). Because some of the pebbles in the conglomerates, particularly the spiculite of Morrow age, are of rock types unknown outside the Ouachita region, a Ouachita source is inferred. The Black Knob Ridge area in the frontal part of the Ouachita thrust belt in northwest Atoka County has been regarded specifically as the source area of cherts in Atoka conglomerates, but chert was not exposed at Black Knob Ridge until after Atoka time, thus ruling out this area (Knechtel, 1937, p. 126). The pebbles may have been derived from the source area that earlier furnished the boulders in the Johns Valley Shale (Shideler, 1970, p. 801). This suggestion is made in the absence of evidence indicating a distinct new orogenic episode in Atoka time.

After the Krebs Group was deposited, both the Arkoma basin and the Ouachita Mountains were extensively folded and uplifted, marking the major phase of the Ouachita orogeny and establishing the structural fabric evident today. Fold axes were initially in an arcuate pattern and folding was more intense in the southeast than in the north and west.

The Cabaniss Group, which overlies the Krebs, is not folded, although it lies close to older folded rocks in western Pittsburg County, Okla. Thus, folding seems to have ended before deposition of the Cabaniss.

Regional uplift accompanied folding and resulted in a change in the depositional strike of post-Krebs sediments. This uplift included an apparent rejuvenation of the Ozark uplift north of the Ouachita Mountains. During deposition of the Krebs and Lower Pennsylvanian, the depositional strike was east; the strike subsequently changed to north-northwest between western Pittsburg County and the Kansas border. This new depositional strike was continued through later Pennsylvanian time.

Overthrust faults are shown on maps of interval C, although they may not have developed until later Missourian or even Virgil time. Studies by Ham and Wilson (1967, p. 386-387) pointed out that some of the compressive forces responsible for the overthrust faulting produced strike-slip faulting in the Arbuckle Mountains during interval E (Virgil) time.

The compressive forces of later phases of the Ouachita orogeny created joint systems and belts of en echelon faults in rocks as young as Early Permian in the eastern half of Oklahoma. The joint systems radiate outward from the Ouachita region (Melton, 1929) and are attributed to compression (Friedman, 1964, p. 485-486). The en echelon faults are shown on the "Geologic Map of Oklahoma" (Miser, 1954) and are regarded by Friedman (1964, p. 486) as "near surface features related to wrench faults at depth, which would be left-lateral and would trend N. 10°-15° E."

In unfolded strata in eastern Oklahoma the joints that parallel folds of the Ouachita Mountains are regarded by Friedman (1964, p. 486) as "relaxation fractures—formed upon release of stored elastic strain energy."

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D is equivalent to the rocks of the Missouri Series of the midcontinent region. This unit includes the fusulinid genus *Kansamella*. Rock-stratigraphic units assigned to interval D are shown on plate 13 and in table 23.

The maximum number of rock-stratigraphic units in the interval occurs in the area between southern Muskogee County, Okla., and the Kansas border, where 2 groups and 11 formations are recognized. Between Muskogee County and the Arbuckle Mountains to the south, equivalent rocks are divided into four formations.

In the Ardmore basin area, interval D consists of the Hoxbar Formation. The same name is used for equivalent rock units in the subsurface west of the Arbuckle Mountains. In the Anadarko basin the *Pleasanton, Kansas City, and Lansing Groups* are differentiated (pl. 10D, sec. W-W'). In other parts of this region, formation names generally are not applied to rock units; the name Canyon Group or Missouri Series is used.

**UPPER BOUNDARY OF INTERVAL D**

In the area where rocks of interval D (Missouri) crop out in eastern Oklahoma, the top of the interval is placed

**Table 23.—Rock-stratigraphic units included in interval D in eastern Oklahoma**

<table>
<thead>
<tr>
<th>Platform</th>
<th>Formation</th>
<th>Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ochelata</td>
<td>Tallant Formation</td>
<td>Hilltop Formation</td>
</tr>
<tr>
<td></td>
<td>Barnsdall Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Wann Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Iola Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Chanute Formation</td>
<td></td>
</tr>
<tr>
<td>Skiatook</td>
<td>Dewey Limestone</td>
<td>Belle City Limestone</td>
</tr>
<tr>
<td></td>
<td>Nellie Bly Formation</td>
<td>Francis Formation</td>
</tr>
<tr>
<td></td>
<td>Hogshooter Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Coffeyville Formation</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Checkerboard Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Seminole Formation</td>
<td>Seminole Formation</td>
</tr>
</tbody>
</table>
at the base of the widespread Vamoosa Formation. The base of the Vamoosa has been mapped between the Arbuckle Mountains and the Kansas border by Oakes (1950). In the subsurface to the west, the equivalent of the Cheshawalla Sandstone Member at the base of the Vamoosa is the Tonkawa sand of subsurface usage. The Tonkawa is the basal unit of the Douglas Group of the Virgil Series (interval E) and, though locally discontinuous, is recognizable throughout most of the Anadarko basin (pl. 10D, sec. W-W'). Despite earlier disagreement (Pate, 1959, p. 45; Adkison, 1960), the Tonkawa sand is not regarded as equivalent to the Stalnaker sand of Kansas (Lukert, 1949, p. 145; Winchell, 1957, p. 137; W. L. Adkison, oral commun., 1967).

In the Ardmore basin and in the subsurface west of this basin, the Zuckerman Limestone Member is at the top of the Hoxbar Formation, and its top is used as the top of interval D.

In the Palo Duro, Hardeman, and Dalhart basins, Canyon (Missouri) rocks are very similar to those of the overlying Cisco (Virgil) rocks, and the interval boundary is difficult to recognize. Where carbonate rock constitutes the dominant facies, fusulinid identifications by the Paleontological Laboratory, at Midland, Tex., have been used for both series and interval boundaries. Where detrital rocks constitute the dominant facies, the boundary is placed in a dominantly gray mudstone sequence. The mudstone below the boundary is splinterly, contains traces of brown marl, and is only slightly arenaceous, whereas the mudstone above does not have a splinterly texture and very rarely has traces of brown marl; the higher unit is more arenaceous than the rock below and contains thin interbedded sandstone, limestone, and dolomite.

THICKNESS TRENDS

Thickness trends of interval D (Missouri) rocks, shown on plate 8A, parallel thickness trends of earlier intervals. An exception to this occurs in eastern Oklahoma, where trends are nearly normal to trends in older intervals. The original edge of the interval in eastern Oklahoma is believed to parallel the present zero line shown on plate 8A.

Interval D is thin along the north edge of the region but thickens southward to more than 3,455 feet in the Anadarko basin. Maximum thicknesses in other parts of the region are 2,450 feet in the panhandle areas and at least 600 feet in the Dalhart basin, 1,200 feet in the Palo Duro basin, and 3,600 feet in the Ardmore and Marietta basins.

Northeastward and southwestward from a line paralleling the Amarillo-Wichita axis, rocks thicken toward the Anadarko and Hardeman basins. Some thickening occurs both north and south of the Matador arch and in the axial part of the Dalhart basin. The Ardmore and Marietta basins seem to have been a single depositional basin, with thicknesses greatest in the Marietta, but the trends are much disturbed by faults.

LITHOFACIES TRENDS

Interval D (Missouri) rocks consist dominantly of fine detrital sediments (pl. 8B); the principal variations in facies result from variations in percentage of limestone. In most of the area north of the Amarillo-Wichita axis the interval contains 50 percent or more mudstone, but limestone becomes abundant northward toward and onto the shelf. Interval D in the Ardmore and Marietta basins consists mostly of fine detrital material that only locally includes more than 10 percent sandstone or limestone.

Coarser detrital rocks fringe uplifted areas that were undergoing active erosion. In western Oklahoma such rocks are limited to narrow bands along the north edge of the Wichita uplift in Washita and Beckham Counties; many of these rocks are feldspathic along the south edge of the uplift in Greer and Harmon Counties.

In the Texas Panhandle, coarse detrital rocks were deposited near the Amarillo uplift and the Bravo dome. The only interruption in the trend is in the intervening area of the shallow Dalhart basin.

Mudstones and other fine-grained detrital rocks are dominant in the deep parts of the Anadarko, Palo Duro, and Hardeman basins. The proportion of limestone increases in shallower parts of the basins in areas of thinning.

SOURCES AND ENvironments OF DEPOSITION

The Amarillo-Wichita uplift, the Ozark uplift, the Ouachita structural belt and the area of the present Arkoma basin were positive areas during Missouri time (interval D), and probably all were major source areas of relatively fine detritus deposited in northern and northeastern Oklahoma. The Bravo dome and a small crescent-shaped island area, on the Amarillo uplift west of the main uplift, were minor source areas. The coarse detritus in rocks of the Dalhart basin probably came from positive areas to the west in New Mexico. Most deposition seems to have been in normal shallow-marine environments, but in the Anadarko and the Ardmore and Marietta basins fine sediments apparently were deposited in deeper water during rapid subsidence.

PALEOTECTONIC IMPLICATIONS

During Missouri time the Amarillo-Wichita uplift continued to be a positive area and was the source of a great volume of sediment.

In the Anadarko, Ardmore, and Marietta basins active subsidence took place as deposition occurred. The eastern part of the region was raised epeirogenetically to produce a large landmass. The uplifted region in-
cluded the Ozark area, Arkoma basin area, Ouachita structural belt, and, according to Tomlinson and McBe (1959, fig. 10), the Arbuckle uplift.

**INTERVAL E**

**FORMATIONS INCLUDED**

Rocks of the Virgil Series of the midcontinent region include the *Tritites* fusulinid zone and make up interval E. The interval includes equivalents of the Douglas, Shawnee, and Wabaunsee Groups of Kansas (pl. 13). Between the Arbuckle Mountains and the Kansas border, outcropping rocks of this interval include the Vamoosa, Lecompton, Ada, and Vanoss Formations.

The Vamoosa Formation is equivalent to the Douglas and lower part of the Shawnee Group of Kansas; the Ada, Lecompton, and Vanoss Formations are equivalent to the upper part of the Shawnee and the Wabaunsee Groups of Kansas. The Vanoss Formation crops out around the west end of the Arbuckle Mountains and in the area between the Arbuckle Mountains and the Criner uplift.

In part of the Anadarko basin the three groups of the Kansas section can be differentiated (pl. 10D, sec. W-W’); elsewhere in the basin, the Virgil Series is undivided. In the southern part of the Texas Panhandle, the rocks are called the Cisco Group.

**UPPER BOUNDARY OF INTERVAL E**

The top of interval E coincides with the boundary between the Pennsylvanian and Permian Systems. In the northern part of the region, the top of the Brownville Limestone Member of the Wood Siding Formation is used as the boundary. The Brownville, or an approximately equivalent bed, is also present and marks the boundary in the panhandle areas of both Oklahoma and Texas.

In the eastern part of the Anadarko basin and east of this basin, the boundary is more difficult to place. In these areas the upper limit of fragments of chert in sandstone is used as an arbitrary boundary, on the premise that chert-bearing rocks were completely stripped from source areas by the end of Pennsylvanian time.

The top of interval E in the Dalhart basin is obscure because of the similarity between Lower Permian and Upper Pennsylvanian rocks and because of poor paleontologic control (Dixon, 1967, p. 68). Interval E contains more carbonate rocks than the overlying Permian, but the physical boundary thus determined does not correspond exactly with the time boundary.

For the Palo Duro basin and for the Hollis and Hardeman basins paleontologic data are moderately abundant and suggest a boundary near the bottom of a limestone overlying a sequence of mudstone and siltstone in most places (Dixon, 1967, p. 67). Locally, however, the boundary is obscured by limestone reefs of late Virgil age (Roth, 1955; Totten, 1956), and at some places the lithologic break does not coincide with the systemic boundary (Totten, 1956). The boundary is also obscure on the southwest flank of the Amarillo-Wichita uplift, where arkosic siltstone and sandstone are the dominant lithology in both Upper Pennsylvanian and lowest Permian rocks. In all these areas the top of interval E is placed at the top of rocks that are more calcareous than overlying rocks.

In the eastern parts of the Hollis and Hardeman basins, the top of interval E is arbitrarily placed at the base of the lowest arkosic, non-chert-bearing sandstone. This horizon seems to coincide very closely with that of the area to the west, where paleontologic data are used to determine the systemic boundary. Similar lithic criteria are used to determine the base of the Permian System in the Ardmore and Marietta basins and on the arch west of the Arbuckle Mountains.

**THICKNESS TRENDS**

The thickness trends of interval E (Virgil) rocks, shown on plate 9A, are about the same as those of interval D. Major differences occur in the Anadarko basin area, where thinning from basin to shelf is less than in other intervals (pl. 10D, sec. W-W’).

In the Anadarko basin the axis of maximum thickness lies much farther north than axes in earlier intervals (pl. 9A). From this axis of thickest deposits, the rate of thinning southward toward the area of faulting at the front of the Wichita uplift is about equal to the rate northward from the axis.

The maximum thickness of the interval in the eastern part of the Anadarko basin is about 3,500 feet, as shown in a well in northern Caddo County. In the panhandles of Oklahoma and Texas the maximum is 2,458 feet, in Wheeler County, Tex. In the Dalhart basin the thickness is as great as 950 feet but averages about 500-600 feet. In the Palo Duro basin the thickness is 700-900 feet in the eastern part and 300-400 feet in the western part. In the Hardeman basin the questionable maximum thickness is 1,783 feet, and the average is about 1,200-1,300 feet. A general thinning southeastward across the Ardmore and Marietta basins is probably the result of post-Pennsylvanian erosion rather than depositional thinning.

**LITHOFACIES TRENDS**

The lithofacies pattern for interval E is shown on plate 9B. The interval generally contains more fine-grained detrital rocks than do intervals C and D, and facies differences are less marked than earlier in the Pennsylvanian.

In the deep part of the Anadarko basin gray mudstone and siltstone are dominant. Around the south, west, and north edges of the basin sandstone and sandy or silty
mudstone are dominant. The deep part of the Palo Duro basin also contains mainly gray mudstone but also has minor amounts of fine sandstone and siltstone. In the Hardeman basin fine sandstone and siltstone are dominant. In the Dalhart basin medium- to fine-grained gray to red sandstone and siltstone are most abundant, and some coarse arkosic material is present along the west edge.

Coarse detrital rock is present in small areas in Washita, Greer, and Harmon Counties, Okla., and in Hutchinson, Moore, Potter, Carson, and Gray Counties, Tex. Coarse detrital rock is also present in a small lobe south of the Bravo dome and east of the Sierra Grande uplift in northeastern New Mexico.

In the eastern part of the Texas Panhandle and in western Oklahoma interval E immediately north of the Amarillo-Wichita uplift is somewhat more calcareous than underlying intervals, but the increase is too small to be apparent on plate 9B. Limestone is present in the Ardmore and Marietta basins area, but there, too, the amount is less than 20 percent of the total, so the increase is not apparent on the facies map.

Limestone and silty limestone are dominant at the margins of the Palo Duro and Hardeman basins; this is also the facies relation in intervals C and D, but the limestone facies is less extensive in interval E. Likewise, sedimentary rocks at the west end of the Nemaha highlands (central Oklahoma platform), on the adjoining Keyes dome, and on the Cimarron uplift are dominantly silty limestone and limestone in interval E as they are in the two underlying intervals.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Fine-grained detritus in interval E was derived from the Amarillo uplift, the Arbuckle uplift, a positive area in the east that included the Ozark and Ouachita uplifts, the Bravo dome, the Sierra Grande uplift in northeastern New Mexico, and the west end of the Matador arch. Absence of interval E rocks in two small areas in the Texas Panhandle suggests that some local islands may have been sources of sediments in areas where seas were shallow.

Sandstones of interval E in some areas contain fragments of both chert and feldspar. Sources of the chert were probably rocks of Cambrian and Ordovician age in the Arbuckle uplift and at the east end of the Wichita uplift. Feldspar may have come from the west end of the Wichita uplift and possibly also from the Amarillo uplift. Longshore currents probably carried feldspathic sand to the eastern part of the Anadarko basin and into the area east of the basin, as suggested by the eastward trend of some of the sandy facies in the basin north of the uplift.

Interval E accumulated in normal-marine, generally shallow water environments. The carbonate facies generally coincides with areas where the interval is thin, which indicates slower rates of sedimentation for the carbonate facies than for the less calcareous rocks.

PALEOTECTONIC IMPLICATIONS

Tectonic features established earlier in the Pennsylvanian were, with rare exception, less active in interval E (Virgil) time than before. Possibly, uplift and basin development was in a dying phase.

Appreciable renewed uplift in the Arbuckle area is indicated by conglomerates in the Vanoss Formation north of the uplift. The Collings Ranch Conglomerate, preserved in a synclinal downwarp within the uplift, is composed of boulder-size material. Erosion that resulted from the renewed uplift breached the Paleoozoic cover in the Arbuckle uplift area and exposed granitic rocks of the basement; conglomerates in both the Vanoss and the Collings Ranch contain granitic gravels. Feldspar detritus was also furnished to sediments north of the uplift.

The isopachs (pl. 9A) indicate the continued downwarping of the Anadarko basin. Deepening of the Anadarko basin accompanied uplift along the axis of the Amarillo and Wichita uplifts. Slight doming on the north side of the uplifts may have dragged upward the sedimentary rocks on the south flank of the basin. Probably this upward drag was lessened by faulting between the Wichita-Amarillo uplift and the basin sedimentary rocks but was sufficient to establish a northward slope on the floor of the depositional basin. Once such a slope had been established, normal depositional processes probably would have brought about greater accumulation in the deep parts of the basin. This mechanism seems to us to be the most probable cause of the northward shift of the depositional and structural axis in the Anadarko basin in Virgil time. The Hollis and Hardeman basins were incorporated in a larger, unnamed area of deposition that included the site of the former Red River arch (W. J. Mapel, chap. K). Isopachs also define the Dalhart basin, showing two centers of deposition separated by a saddle area between the Bravo dome and the Cimarron uplift in the western part of the Texas Panhandle. The Matador arch was only weakly positive according to available thickness data.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

THICKNESS TRENDS

In the Oklahoma and Texas Panhandle region Pennsylvanian rocks are thickest in the Anadarko basin, where they are more than 15,000 feet thick (pl. 11). They thin northward within the basin to less than 1,900 feet in a small area in Woods and Alfalfa Counties, Okla. The thickness decreases westward toward the Cimarron uplift. Thinning is also apparent along the south side of the
basin, but the trend is interrupted by faulting along the north side of the Wichita and Arbuckle uplifts.

More than 14,000 feet of Pennsylvanian rocks is present in the axial part of the Ardmore basin. The rate of thinning in all directions away from this maximum has probably been increased by postdepositional uplift and erosion of the bordering positive features—the Arbuckle uplift to the north and the Criner uplift to the south.

Pennsylvanian rocks in the Dalhart and Palo Duro basins are thinner than in nearby basins even though both basins were negative during most of Pennsylvanian time. The Dalhart basin is more clearly defined on plate 11 than on interval thickness maps. The basin axis lies along the west side of the basin, adjacent to the positive element that was the source of sediments during much of Pennsylvanian time.

In eastern Oklahoma, including the region of the Arkoma basin, isopachs tend to parallel the strike of outcropping beds and reflect, in large part, postdepositional truncation. The thickening trend changes to southwest in central Oklahoma, where rocks of Pennsylvanian age are covered by Permian rocks.

PALEOTECTONIC IMPLICATIONS

The pattern of total Pennsylvanian thickness in the eastern part of the region reflects modification of the Early Pennsylvanian structures by later Pennsylvanian uplift in the Ouachita and Arbuckle areas. In the southern Ouachita Mountains the north-south thickness trend is now nearly normal to the trend of the depositional trough of Early Pennsylvanian age. West of the Ozark area, uplift during later Pennsylvanian time was in part responsible for shifting the original east-trending strike of Lower Pennsylvanian rocks to a present northeasterly direction.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

UNITS OVERLYING PENNSYLVANIAN

Pennsylvanian rocks are widely exposed in the eastern third of Oklahoma and in a westward-projecting salient between the Arbuckle and Criner uplifts; they are overlain to the west by rocks of Wolfcamp (Permian) and younger age (pl. 12). In southeastern Oklahoma, Permian rocks are absent, and Pennsylvanian strata are overlain by the Trinity Group (Lower Cretaceous).

PALEOTECTONIC IMPLICATIONS

The distribution of geologic units above the Pennsylvanian System indicates periods of tectonic activity at the end of Pennsylvanian and Permian times. In the west the distribution, stratigraphic relations, and lithologies of Lower Permian rocks indicate that Late Pennsylvanian depositional conditions continued into the Permian without appreciable break. In the east, post-Pennsylvanian uplift resulted in westward movement of the shoreline prior to subsidence that permitted Permian deposition. An absence of conglomerates in the Lower Permian at most places indicates that post-Pennsylvanian source areas were relatively low lying.

At the end of Permian time, subsidence in the south permitted the deposition of Tertiary (Early Cretaceous) rocks on Pennsylvanian rocks in the southern part of the Ouachita Mountains.

REFERENCES


Cunningham, B. J., 1961, Stratigraphy, Oklahoma-Texas Panhandles, in Oil and gas fields of the Texas and Oklahoma Panhandles: Amarillo, Tex. Panhandle Geol. soc., p. 45-60.


Eddelman, M. W., 1961, Tectonics and geologic history of the Texas and Oklahoma Panhandles, in Oil and gas fields of the Texas and Oklahoma Panhandles: Amarillo, Tex., Panhandle Geol. Soc., p. 61-68.


Frezon, S. E., 1962, Correlation of Paleozoic rocks from Coal County, Oklahoma, to Sebastian County, Arkansas: Oklahoma Geol. Survey Circ. 58, 53 p.


Kirk, M. S., 1957, A subsurface section from Osage County to Okfuske County, Oklahoma: Shale Shaker, v. 7, no. 6, p. 2-4, 9, 11-21.


Central and West Texas

By ELEANOR J. CROSBY and WILLIAM J. MAPEL

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-K
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES.
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

CENTRAL AND WEST TEXAS

By ELEANOR J. CROSBY and WILLIAM J. MAPEL

ABSTRACT

During the transition from Mississippian to Pennsylvanian time, the central and west Texas region was a land area of low relief. Marine waters of the Ouachita geosyncline bordered the region on the east and south, and shallow seas partly covered areas to the southwest. From Early Pennsylvanian into Middle Pennsylvanian time, the sea advanced into the relatively stable central part of the region, reaching its maximum extent in early or middle Des Moines time. Slow sinking of platforms and weak development of northwest-trending basins within the central region were accompanied by uplift and inward compression in parts of the marginal geosyncline, by uplift of the westward-trending Muenster, Red River, and Matador arches in the north, and by maintenance of a shallowly submerged or partly emergent positive element northwest of the Llano area. The Fort Worth, Kerr, and Marathon basins deepened, and their axes moved slowly inward toward the central stable region in response to compressive forces within the geosynclinal belt.

Fairly thick fine-grained sandstone, mudstone, and local conglomerates were derived from uplifts in the Ouachita belt to the east and from rising land areas along the Oklahoma border and in the Texas Panhandle. These sediments graded westward and southward into thinner deposits of mostly argillaceous and limy mud. Uplifted areas south or southeast of the Marathon area supplied detrital sediment intermittently to the Marathon segment of the geosyncline.

From middle Des Moines time onward, increased tectonic activity deepened the interior basins (Midland, Delaware, and Val Verde). The interbasin structures (Diablo and Central Basin platforms and Bend arch) remained shallowly submerged or slightly elevated above sea level. Uplift accompanied by thrusting in Des Moines time in the eastern and southeastern segments of the Ouachita belt resulted in sedimentary filling and partial deformation of the Fort Worth basin and reversed the surface slope in the area of the Bend arch from east to west. Filling of the Marathon segment of the geosyncline was completed in later Pennsylvanian time and culminated with northwestward overthrusting south of the Val Verde basin very early in Permian time.

Most of the supply of detrital sediment in Middle and Late Pennsylvanian time was derived from the Ouachita belt; smaller amounts came from independently located sources in westernmost Texas. In the eastern part of the region nonmarine sediments, including coal deposits, intertongued westward with marine limestones and mudstones. Little or no sediment reached the deeper parts of the interior basins in the later Pennsylvanian. Thick banks or reefs of limestone grew on the platform margins and in the Horseshoe atoll within the north half of the Midland basin.

The passage from Pennsylvanian to Permian time was marked by an accentuation of the previously established structural features — downwarping of basins, faulting at basin margins, and local uplift of mountain sources of sediment continued. From uplifted source areas within central and west Texas and others to the northwest came detrital sediments that gradually filled the west Texas basins during Permian and earliest Triassic time.

REGION DEFINED

The central and west Texas region, as defined here (fig. 43), consists of the counties north of lat 29° N. and west of long 96°00' W., excluding the panhandle region of Texas. Pre-Pennsylvanian rocks are exposed in the Llano area in the southeastern part of the region, in the Marathon area in Brewster County, and in the Franklin and Hueco Mountains and Sierra Diablo of westernmost trans-Pecos Texas. Outcropping Pennsylvanian rocks form an almost continuous band as much as 60 miles wide from the east and north edges of the Llano area northeastward to within about 20 miles of the Oklahoma border. Pennsylvanian and older rocks crop out much less extensively in the Marathon area and in trans-Pecos Texas farther to the west and northwest.

In the early part of the Pennsylvanian, major structural elements that determined the distribution and thickness of the deposits were a fairly stable interior lowland and the bordering Ouachita geosyncline on the east and its continuation, the Marathon geosyncline, on the south (fig. 44). The Muenster, Red River, and Matador arches formed a structurally high band across northern Texas, and the Fort Worth basin occupied a broad embayment at the west edge of the Ouachita geosyncline south of the Muenster and Red River arches. Other structural features in the Texas interior region were the eastern part of the Hardeman basin in Hardeman and adjacent counties of Texas; the Marietta syncline, a segment of the Ardmore basin in northern Grayson County; a shallow unnamed basin in a broad embayment in southeastern New Mexico and adjacent Texas north of the Marathon geosyncline; and a northwest-trending...
Figure 43. — Geographic features in central and west Texas and adjacent areas mentioned in text.
structural high in Pecos and Crockett Counties, called by Galley (1958, p. 417) the Pecos arch. Other structural features may have existed within the interior region but are not definable in the absence of an Early Pennsylvanian depositional record.

In Middle and Late Pennsylvanian time, important structural features in central and west Texas were, from east to west, the Ouachita structural belt, the Fort Worth basin (in part of Middle Pennsylvanian only), the Bend arch, the Midland basin, the Central Basin platform, and the Pedernal uplift (fig. 45). Farther south in Texas, the Kerr, Val Verde, and Marfa basins also were active in the later part of the Pennsylvanian. Most of the features listed were within or framed the Permian basin of west Texas and New Mexico in Late Pennsylvanian and Permian time.

The Bend arch, between the Fort Worth and Midland basins, has a complex history. A broad north-trending arch is evident in lower Middle Pennsylvanian rocks and on the pre-Pennsylvanian surface, but, as Cheney (1929, p. 10; 1940) and others have pointed out, it differs importantly from the other major structural elements mentioned above. It is a composite feature — its east flank was formed by downfolding or tilting to the east in Late Mississippian to early Middle Pennsylvanian time, and its west flank was formed mainly by later Pennsylvanian and Permian tilting in the opposite direction, although it presumably had a slight westward slope in earlier Pennsylvanian time. The term “Bend flexure” (Cheney, 1929, 1940) has been applied to the principal hinge line of downwarping of the Fort Worth basin. In this publication the term “Bend arch” is used, in a morphologic rather than genetic sense, for the broad two-phase structure as a whole; in early Middle Pennsylvanian time it functioned mostly as the western shelf of the Fort Worth basin, whereas in Late Pennsylvanian and Permian time, after filling of the Fort Worth basin, it formed the eastern shelf of the Midland basin. To the south it intersects the older Paleozoic, northwest-trending Concho arch.
The Ouachita geosyncline, including the Marathon segment, of Early Pennsylvanian time was replaced later in the Pennsylvanian by the broadly sinuous Ouachita structural belt. Rocks in this belt are intensely deformed, locally metamorphosed, and mostly buried beneath Cretaceous rocks. A virtually complete Pennsylvanian sequence, folded and thrust northward in Late Pennsylvanian or earliest Permian time, is exposed in the Marathon salient, in Brewster and Pecos Counties; and farther south, deformed rocks of the structural belt crop out in the Solitario area, in Brewster and Presidio Counties. Elsewhere in the structural belt, data from scattered wells give some information on the deformed Pennsylvanian rocks, but the information is inadequate for any detailed stratigraphic analysis.

PALEOGEOLOGY UNITS UNDERLYING PENNSylvANIAN

Folded and faulted Ordovician, Cambrian, and Precambrian rocks underlie the Marathon area (pl. 2) on the crests of the Muenster and Red River arches, and Ordovician rocks of the Ellenburger Group lie beneath the Pennsylvanian west and northwest of the Llano area in a broad salient that extends from Irion, Tom Green, and Concho Counties northward into Jones County. The Ellenburger also underlies the Pennsylvanian at places within a narrow band along the south edge of the Llano uplift. (The term “uplift” where applied to the Llano area refers to the post-Pennsylvanian structural element.) Within the broad subcrop area of the Ellenburger, Cambrian and Precambrian rocks are directly below the Pennsylvanian west and northwest of the Llano area in a few fault blocks too small to be shown on plate 2 (Conselman, 1954). Pennsylvanian rocks overlie rocks of Mississippian age north, east, and south of the Llano area and in much of west Texas. The Mississippian consists of the Comyn Formation, Barnett Shale, and Chappel Limestone in the area of the Fort Worth basin, unnamed limestone units of Chester and other Mississippian age in the northern Midland basin area, rocks assigned to the Barnett Shale in the areas of the southern Midland and Delaware basins and Val Verde basin, part of the Tesnus Formation in the Marathon area, and the Helms Formation in westernmost trans-Pecos Texas.

At the south end of the Central Basin platform in Pecos County and in much of Crockett County and nearby areas, Ordovician (Ellenburger and Simpson Groups), Silurian, and Devonian rocks lie beneath the Pennsylvanian. Devonian rocks (in areas too small to be shown at the scale of pl. 2) are reported under Pennsylvanian locally along the northern part of the Central Basin platform; at a few places in the Marathon area; and elsewhere, locally, in trans-Pecos Texas.

LOWER BOUNDARY OF PENNSylvANIAN

Deposition continued without interruption from Mississippian into Pennsylvanian time on some marginal parts of the interior lowland (fig. 44); inward from the margin, the base of Pennsylvanian rocks is marked by an unconformity.

Sedimentation seems to have been continuous in the eastern Fort Worth basin in parts of Denton, Tarrant, and some nearby counties where mudstone and sandstone of Morrow age overlie, without any apparent lithologic break, mudstone, sandstone, and, locally, some thin limestone beds of the Barnett and Comyn sequence of Mississippian age.

Thick, continuously deposited sequences of interbedded dark mudstone and sandstone correlated with the Stanley Shale and overlying Jackfork Sandstone of Oklahoma and Arkansas have been found by drilling in the frontal zone of the Ouachita structural belt in Fannin County (Flawn, in Flawn and others, 1961, p. 73). The position of the Mississippian-Pennsylvanian boundary within these formations has long been debated. In this study, the Stanley-Jackfork contact is considered to be the contact between the Mississippian and Pennsylvanian Systems.

In the Marathon area, the lithologic and structural correlative of the Stanley and Jackfork sequence is the Tesnus Formation, a thick mudstone and sandstone unit containing few fossils. The uppermost part of the Tesnus is considered to be Early Pennsylvanian on the basis of plant fossils (David White, in King, 1937, p. 61) and a few Foraminifera (Bruce Harlin, in Powers, 1928, p. 1066). In the lower part of the Tesnus are Mississippian conodonts (Ellison, 1962). The Tesnus in the Solitario area may be partly Pennsylvanian and partly Mississippian (Flawn and others, 1961, p. 60) or Mississippian only (Wilson, 1954, p. 2460). Inferred lateral continuity of the Tesnus with the Stanley and Jackfork sequence along the Ouachita structural belt between Oklahoma and west Texas is the basis for the assumption that the Mississippian-Pennsylvanian boundary is fairly high within the Tesnus — a position analogous to its assumed position in Oklahoma.

In westernmost trans-Pecos Texas (El Paso and Hudspeth Counties), the lower boundary of the Pennsylvanian is taken to be a conformable contact of the Magdalena Formation, of Early to Late Pennsylvanian age, on the Helms Formation, of Mississippian age (Harbour, 1972; G. O. Bachman, oral commun., 1964). Possibly the boundary should be placed somewhat lower in the upper, apparently unfossiliferous part of the Helms. From northern Culberson County eastward to central Pecos County, identification of the upper and lower contacts of Pennsylvanian strata is uncertain. These contacts are in a dominantly mudstone unit
between Permian rocks and mudstone commonly assigned to the Mississippian Barnett Shale. In this area, deposition probably was continuous from Mississippian into Pennsylvanian time.

Successively younger rocks compose the base of the Pennsylvanian from the Fort Worth basin westward across the axis of the Bend arch and northward across the Muenster and Red River arches. The overlapping units, mapped in figure 46 as intervals A, B, C, and D, are approximately equivalent in age to the Morrow, Atoka, Des Moines, and Missouri Series of the midcontinent region of the United States. Locally, near the Oklahoma border in Montague, Clay, Wichita, and Foard Counties, Pennsylvanian rocks as young as Missouri are at the base of the system, and rocks of Missouri age locally lap onto Ordovician rocks in southernmost McCulloch County and eastern Menard County on the west side of the Llano region (Plummer, 1950, pl. 1).

In outcrops on the north and east sides of the Llano region in San Saba, Lampasas, Burnet, and Blanco Counties, cherty or siliceous limestone of Morrow age unconformably overlies brownish-black mudstone of the Mississippian Barnett Shale and, locally, overlies the Chappell Limestone or the Ellenburger Group. The basal Pennsylvanian limestone in this area is the Marble Falls Limestone, or, locally, Group (Plummer and Moore, 1922, p. 32; Cloud and Barnes, 1948, p. 59; Plummer, 1950, p. 47; Moore, 1944, p. 697; Stewart, 1957, p. 47; Bell, 1957, 1962, p. 94-95). Farther west, in most of McCulloch County, the lithologic relations at the Pennsylvanian-Mississippian contact are the same as in San Saba and Lampasas Counties; however, because of westward transgression, the oldest Pennsylvanian limestone in McCulloch County is mostly or wholly Atoka in age and is referred to by some geologists to the Big Saline Formation (Plummer, 1947, p. 196, 197; Thompson, 1947, p. 147; Stewart, 1957, p. 47; Turner, 1957a, p. 67). In the subsurface north of the Llano area, the Pennsylvanian-Mississippian boundary is picked at the
top of the Comyn Formation of Palo Pinto County or its equivalents as correlated in cross sections of the North Texas Geological Society (1954). A Mississippian (Chester) age is inferred for the Comyn because, as Turner stated (1957a, p. 63-64), the Comyn apparently intertongues at its base with the underlying Mississippian Barnett Shale, and at places along the western and northern flanks of the Fort Worth basin the Comyn is unconformably overlain by sandstone or chert conglomerate that is interpreted to be the basal deposit of the transgressing Pennsylvanian sea. Fossils necessary for unequivocally dating the Comyn are lacking.

On the southern part of the Bend arch in Concho, Tom Green, Schleicher, Sutton, and parts of adjacent counties, the Ordovician Ellenburger Group is separated from overlying limestone that contains Pennsylvanian fossils by a discontinuous deposit of red, green, or gray mudstone containing local sandstone lenses and abundant fragments of varicolored chert. The thickness of the "detrital zone," as it is called, commonly ranges from a few inches to about 20 feet, though locally it is reported to be much thicker (Cheney and Goss, 1952, p. 2244; Rall and Rall, 1958, p. 845, 846). The deposit is generally regarded as residual material derived from prolonged weathering of the Ellenburger; it is included in this study with the Pennsylvanian rocks.

A similar "detrital zone" is present locally at the base of the Pennsylvanian where it overlies Ordovician to Devonian rocks in Crockett and eastern Pecos Counties and at places on the Central Basin platform northwest of these counties. On and near the Central Basin platform, comparable units of variegated mudstone, fragmental chert, and sandstone finger out laterally between Pennsylvanian limestones, suggesting intermittent transportation in Pennsylvanian time of residual material from the structurally high areas into parts of the adjacent basins.

In the northwestern part of the Midland basin area, rocks of Morrow age are missing at the base of the Pennsylvanian, and dominantly detrital rocks of Atoka age unconformably overlie Upper Mississippian limestone (fig. 46). In the central and southern parts of the basin
the age relations at the contact are the same, but the proportion of carbonate rock in the Atoka is greater than farther north, and the underlying Mississippian is a mudstone called the Barnett Shale. Probably this unit, in its upper part, is somewhat younger than the Barnett of the Llano area. Lower Strawn limestone of Des Moines age lies unconformably on Mississippian limestone or mudstone and on older rocks on the east and south sides of the Midland basin area and on the south and east margins of the Central Basin platform; at a few places on the Central Basin platform, rocks of Missouri age are at the base of the Pennsylvanian.

PALEOTECTONIC IMPLICATIONS

The greater part of central and west Texas was a stable to slowly sinking shelf covered by a shallow epicontinental sea during much of Mississippian time (Eardley, 1949, fig. 4; Adams, 1962, p. 374). At about the end of Mississippian or the beginning of Pennsylvanian time, a regional uplift or lowering of sea level initiated the unconformity that separates Pennsylvanian from Mississippian and older rocks in most of central Texas and parts of west Texas. Probably most of this region was brought only slightly above sea level.

Uplift may have been greatest along the trend of the Red River arch and in the southern part of the Bend arch (Turner, 1957a, p. 75), inasmuch as Middle Pennsylvanian or younger rocks overlie rocks older than Mississippian on these structural elements. Beveled Mississippian rocks beneath the Pennsylvanian System on the crest of the east-trending Pecos arch, on the Central Basin platform, and in the northern part of the Bend arch, indicate that these areas may also have been slightly positive. Additional uplift occurred on all these structural features during Pennsylvanian time, however, and the amount and effect of any Late Mississippian movement cannot be evaluated separately. An embayment that extended into the northern part of the future Fort Worth basin sank fairly rapidly and continued to receive deposits of marine mud and calcium carbonate from Mississippian into Early Pennsylvanian time.

Little is known of Late Mississippian or earliest Pennsylvanian tectonic elements in trans-Pecos Texas. The southward-trending axis of the Pedernal uplift in New Mexico may have extended into Hudspeth County as a low uplift at this time (Lloyd, 1949, p. 14; Adams, 1962, p. 374), but subsequent uplift and erosion and sparsity of drilling make difficult the interpretation of any structural detail.

In Late Mississippian time, the stable or slowly rising interior lowland of central Texas was bordered on the east and south by the rapidly sinking Ouachita geosyncline (Sellards, 1933, p. 22; P. B. King, in Flawn and others, 1961, p. 184), including the Marathon segment in southwestern Texas. Facies and thickness relations in Upper Mississippian rocks in the Fort Worth basin indicate the presence of a hinge line at the west edge of the Ouachita geosyncline, somewhat east of the present edge of the Ouachita folded belt.

Rocks of the geosyncline can be traced in the subsurface from the Oklahoma-Texas border around the south side of the Llano area into Uvalde County; farther west, similar rocks have been recognized in drill holes from Terrell County northwest into the Tesnus outcrops in the Marathon salient of the folded belt in northern Brewster County (Flawn and others, 1961; Goldstein and Hendricks, 1962). From the Ouachita Mountains to southwestern Texas, the only discontinuity in this belt of thick Mississippian to Lower Pennsylvanian detrital rocks is in Kinney and Val Verde Counties near an abrupt bend in the structural trend from southwest to northwest. In that area, the Devils River uplift, a buried positive element of Precambrian and lower Paleozoic rocks, apparently lies along the northwest-trending segment of the structural belt. The geosyncline may have been deflected to the southwest side of the uplift; if so, rocks equivalent to the Tesnus may be present below metamorphosed older rocks that in Middle or Late Pennsylvanian time were thrust northward from the interior of the structural belt (Flawn and others, 1961, p. 172-173).

The Ouachita-Marathon geosyncline is believed to have extended from the Solitario area, Texas, west or southwest into eastern Chihuahua, Mexico. Silty or cherty limestone in the Placer de Guadalupe area (fig. 43) is interpreted as having been deposited on a shelf or platform at the north edge of the geosyncline or in a transitional zone between the geosyncline and the shelf (Flawn and others, 1961, p. 101; Bridges, 1964, p. 93). The axial position of the geosyncline in Chihuahua has not been determined.

INTERVAL A
FORMATIONS INCLUDED

Interval A (pl. 3A) consists of the Marble Falls Limestone in the eastern part of the Llano area (table 1) and in the central part of the Fort Worth basin, and it consists of unnamed rocks presumed to be of Morrow age in the Hardeman basin. Along the east side of the Fort Worth basin (McLennan, Bosque, and Johnson Counties), no lithologic or paleontologic basis for separating intervals A and B has been recognized, and the two are shown combined on the thickness map for interval B (pl. 6A).

The Marble Falls Limestone, as the name is applied here, is the limestone sequence that occurs between the base of the Smithwick Shale and the top of the Barnett Shale in outcrops along the north and east sides of the Llano area in San Saba, Lampasas, Burnet, and Blanco Counties. This limestone is here assigned entirely to the
Morrow Series, and hence to interval A, following the usage of Thompson (1947, p. 148), Stewart (1957, p. 49), Branson (1957, p. 83), and Turner (1957a, p. 65).

Rocks mapped as interval A in other parts of Texas include the uppermost part of the Tesnus Formation in the Marathon region; the La Tuna Member of the Magdalen Formation in the Franklin Mountains, El Paso County (Harbour, 1972); part of the lower division of the Magdalen in the Hueco Mountains (King and others, 1945); and limestone and mudstone believed to be of Morrow age from wells in Culberson and Loving Counties (Paleont. Lab., Midland, Tex.). Elsewhere in west Texas, interval A is absent or cannot be distinguished from overlying interval B.

In chapters of this publication dealing with central and eastern regions of the United States, a lower part of interval A has been separately described as subinterval A1. In Texas, subinterval A1 is assumed to be present in the Tesnus Formation and is mapped on plate 4 because deposition in the Marathon area is believed to have been continuous from Late Mississippian into Early Pennsylvanian time; no specific thicknesses can be assigned to this unit.

**UPPER BOUNDARY OF INTERVAL A**

In the Fort Worth basin and in the Llano area, the top of interval A is defined as the top of the Marble Falls Limestone. Mudstone and interbedded sandstone conformably overlie limestone at this contact in most places. Near the edge of interval A (pl. 3A), however, limestone and argillaceous limestone that are correlated with the Big Saline Formation of McCulloch County directly overlie limestone of the Marble Falls. Where this relation is found, the contact is drawn arbitrarily within the limestone sequence.

In the eastern part of the Hardeman basin, the top of interval A is the top of a glauconitic limestone overlain by interbedded limestone, sandstone, mudstone, and conglomerate.

Fusulinid evidence suggests that the top of the Morrow Series may be within the lower part of the Dimple Limestone in the Marathon area (Sanderson and King, 1964), but such evidence provides no mappable boundary for a rock-stratigraphic interval. Therefore, the upper boundary of interval A is placed at the base of the Dimple Limestone as defined by King (1937, p. 62), which is below the lowermost limestone above the mudstone-sandstone sequence of the Tesnus Formation. In the subsurface to the northwest, in Culberson and Loving Counties, where paleontological control is adequate, the upper boundary for interval A corresponds to the boundary of Morrow and Atoka rocks. Where the control is less complete, the base of a sandstone unit thought to lie at about the top of Morrow rocks serves as the top of the interval. In the Hueco Mountains of westernmost trans-Pecos Texas, the top of interval A is the Morrow-Atoka boundary as tentatively identified by King, King, and Knight (1945) within thick-bedded coralline limestone of the lower division of the Magdalena Limestone. In the Franklin Mountains, the boundary lies at the top of the La Tuna Member of the Magdalena (Harbor, 1972).

**THICKNESS TRENDS**

Rocks assigned to interval A are as much as 500 feet thick in the Hardeman basin and are at least 600 feet thick in Denton County in the northern part of the Fort Worth basin (pl. 3A). The zero isopach parallels the Bend flexure and the south edge of the Muenster arch in the Fort Worth basin, and it outlines the south edge of part of the Hardeman basin in Hardeman and adjacent counties. Discontinuities in the isopachs in Cooke and Denton Counties result from faulting and erosion of interval A rocks on the Muenster arch later in Pennsylvanian time.

Interval A in areas of outcropping Marble Falls Limestone on the east side of the Llano uplift is approximately 300-400 feet thick. Limestone identified as Marble Falls in the subsurface east and south of the uplift is generally less than 100 feet thick, but in this area an undetermined thickness of overlying dark calcareous mudstone that has been placed in interval B probably is equivalent to the upper part of the limestone on the outcrop.

The thickness of interval A is not more than 300 feet in the northwestern part of the Marathon area. Rocks of Morrow age probably increase in thickness to the east and southeast, where they constitute an undetermined part of the more than 6,500-foot total thickness of the Tesnus (King, 1937, p. 55). In the Franklin and Hueco Mountains, less than 500 feet of rock is placed in interval A. In northeastern trans-Pecos Texas, the interval thickens eastward from less than 500 feet in Culberson County to at least 1,000 feet in Loving County.

**LITHOFAENCIES TRENDS**

Interval A, where recognized in the Fort Worth basin (fig. 45), is dark-gray to black mudstone interbedded with grayish-black to dark-grayish-brown limestone that is locally glauconitic, oolitic, or cherty. Limestone of the Marble Falls predominates in most parts of the basin where the interval can be recognized (fig. 47); mudstone is more abundant than limestone at the north end and along the east margin of the basin in McLennan, Bosque, and Johnson Counties (pl. 3B). The limestone-mudstone sequence grades laterally eastward into mudstone and sandstone that contain no limestone. The mudstone-sandstone facies is indistinguishable from overlying rocks of interval B and is combined with the lithofacies for interval B (pl. 6B). The lower part of interval A
is slightly sandy in Wise County and, locally, in Parker County.

In the eastern end of the Hardeman basin, limestone and mudstone occur in roughly equal amounts; sandstone is generally present but mostly in amounts of less than 10 percent. The limestone is glauconitic and locally oolitic. No systematic trends in lithofacies are evident.

The Marble Falls Limestone, which makes up interval A on the outcrop along the east side of the Llano uplift, consists of “dark gray and black, siliceous, fossiliferous limestone ledges, generally thin-bedded, containing some layers of black shale and grading eastward into a shale facies * * * east of the area of outcrop” (Plummer, 1950, p. 48). In the subsurface of Williamson County, it is a dark calcareous spicule-bearing mudstone; in well sections to the south, in Kendall County, it consists of limestone, dolomitic limestone, and black mudstone (Flawn and others, 1961, app.).

Interval A in the Marathon area consists of terrigenous detritus of the upper, Morrow part of the Tesnus Formation (King, 1937, p. 55-63; McBride and Thomson, 1964, p. 17-18). In the northwestern part of the Marathon area, where only the thin, upper part of the Tesnus is present, the rocks consist of green to black mudstone and subordinate interbedded cherty mudstone and chert conglomerate. To the east and south, these rocks grade into the upper part of a sequence of several thousand feet of clayey fine-grained sandstone and mudstone; farther to the southeast, fairly coarse sandstone that is notably arkosic constitutes 30-80 percent of the upper two-thirds of the Tesnus, and a few thick beds of white quartzite are present.

Rocks of Tesnus lithology have been recognized in the subsurface east and southwest of the outcrops of the Marathon salient, but an overlying equivalent of the Dimple has not been differentiated from the Tesnus in the subsurface (Flawn and others, 1961, p. 74).

In El Paso and western Hudspeth Counties, interval A consists of dark-gray thick-bedded to massive dense cherty limestone that is crinoidal in the Franklin Mountains (Harbour, 1972) and coralline in the Hueco Mountains (King and others, 1945). In northern Culberson County, subsurface rocks of possible Morrow age assigned to interval A are mainly limestone and mudstone; a wedge or tongue of sandstone thickens and is conglomeratic toward the south.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

Sediments of interval A in north-central Texas were deposited in a sea that slowly encroached on low-lying land areas from the north and east. In the southern part of the Fort Worth basin, the present western limit of interval A probably represents approximately the original depositional limit. Along the south edge of the Hardeman basin and in the northern part of the Fort Worth basin, the interval is truncated beneath rocks of overlying interval B. Interval A, therefore, originally extended some unknown distance onto, and probably across, the areas of the later formed Muenster and Red River arches.

In the Hardeman basin and the central part of the Fort Worth basin, glauconitic spicular limestone characterizes much of interval A and is believed to indicate deposition in quiet, shallow to moderately deep, marine water of normal salinity (Cloud, 1955, p. 490-491). Glauconite seems to require for its formation stagnant, slightly reducing bottom conditions (Cloud, 1955, p. 490). In the Llano area, however, the limestone is oolitic and contains corals, algae, and crinoids (Plummer, 1950, p. 52-54), which indicate that in this area the water was at least intermittently aerated and somewhat turbulent.

In the Fort Worth basin and along the east side of the Llano uplift, the water may have become deeper rather abruptly at about the east edge of the limestone facies of interval A. Detritus in this area was probably supplied from rising land areas farther to the east or southeast.

The detrital zone at the base of Pennsylvanian rocks in Tom Green, Schleicher, Sutton, and other nearby counties indicates deep weathering and the formation of a thick soil. Although the oldest dated Pennsylvanian rocks in this area are of Atoka or Des Moines age and
rocks of Morrow age have not been recognized, formation of the detrital zone may have been in progress during the time of interval A. The irregular thickness of the deposits in this detrital zone has been ascribed to the development of sink holes and other karst features (Cheney and Goss, 1952, p. 2244; Adams, 1954, p. 72; Rall and Rall, 1958, p. 848). Such features would be compatible with a humid climate in an area of fairly low relief.

In the Marathon area, a highland source in crystalline rocks south or southeast of the site of Tesnus deposition is indicated by a thickening of the formation in that direction and an increase in grain size, proportion of sandstone to mudstone, and abundance of arkosic material in the same direction (Waterschoot van der Gracht, 1931, p. 1030; King, 1937, p. 87); flute-cast orientation seems to confirm a dominant northward or northwestward direction of sediment transport (Johnson, 1962, p. 790). A few chert conglomerates near the northwestern limits of the Tesnus, where the formation is mainly mudstone, may have been derived from nearby outcrops of older Paleozoic rocks.

In early investigations, the depositional environment of the Tesnus was variously interpreted as deltaic, shallow marine, or partly continental (Waterschoot van der Gracht, 1931, p. 1035; King, 1937, p. 87; Fan and Shaw, 1956, p. 266). More recently, deep water, into which sand was carried from an adjacent shelf by turbidity currents, has been suggested to account for the several thousand feet of interbedded thin sandstone and mudstone strata in the central and southern parts of the Marathon area (P. B. King, in Flawn and others, 1961, p. 184; Johnson, 1962, p. 791; McBride and Thomson, 1964, p. 20). A few thick sandstone units are massive, but the more abundant thin beds of sandstone contain internal structures and gradation of grain sizes commonly associated with deposition by turbidity currents.

Little or no record of life has been found in most of the upper part of the Tesnus, but Foraminifera near the top indicate a marine environment for at least that part of the interval, and water-worn plant fragments in some sandstones presumably reflect a nearby land area.

In westernmost Texas and adjacent New Mexico, the lowest part of the Magdalena, now exposed in the Franklin and Hueco Mountains, was produced in a warm clear shelf sea. Crinoids and corals grew abundantly. Near the border of the two States, limited amounts of mud were deposited, and sand and coarser detritus were transported into northern Culberson County, apparently from a southern source.

PALEOTECTONIC IMPLICATIONS

The central part of Texas was a stable, slightly emergent lowland during interval A time. Slowly and irregularly sinking shelves bordered this platform on the east and south, and these shelves merged outward into the rapidly sinking Ouachita geosyncline, the position of which can now be located only indefinitely. Detritus supplied to the geosynclinal trough came from rising land areas presumably still farther east and south. Decrease in sandstone in the uppermost part of the Tesnus Formation in the Marathon area suggests a slowing or cessation of uplift and a leveling of high areas of the southern source during late Morrow time.

The Hardeman basin along the Texas-Oklahoma border in Hardeman County was part of a larger, moderately downwarped area that extended northward and northward into Oklahoma and the panhandle of Texas. Fairly rapid downwarping occurred along the northwest-trending Marietta syncline, part of which was in north Texas and part in Oklahoma. This linear basin probably opened out southeastward into the Ouachita geosyncline across Grayson and Fannin Counties.

Thickness trends and lithofacies patterns in the Hardeman basin and the northern part of the Fort Worth basin suggest that the area of the present Red River and Muenster arches was gently downwarped and that a marine connection existed between the two basins across that area.

Northern trans-Pecos Texas subsided slowly and was covered by a shallow sea during the time of interval A. The thickness of mudstone and limestone assigned to this interval suggests slightly greater downwarping in the Delaware basin area than on adjacent parts of the platform (pl. 34; Galley, 1958, p. 420). In northern Culberson County, gradation northward from conglomerate into fine-grained sandstone seems to indicate local uplift in the central part of the county. Elsewhere in trans-Pecos Texas, the record of Early Pennsylvanian tectonic history is masked by thick younger rocks or has been destroyed by postburial uplift and erosion.

INTERVAL B
FORMATIONS INCLUDED

Formations assigned to interval B in the western part of the Llano area and in the western part of the Fort Worth basin are the Big Saline Formation (fig. 47), predominantly a light- to dark-gray crystalline cherty limestone, and the overlying Smithwick Shale, typically dark-gray to black claystone containing some interbedded limestone, siltstone, and sandstone (pl. 13). Both formations change facies eastward; the Big Saline grades into dark mudstone that is indistinguishable from the Smithwick; farther east the Smithwick, including the part equivalent to the Big Saline, grades into sandstone, mudstone, and conglomerate of Atoka age (Cheney and Goss, 1952, p. 2255; Turner, 1957a, p. 67-69). Interval B may locally be missing from outcrops along the Colorado River in eastern San Saba County, for the Strawn Group is mapped directly on the Marble Falls Lime-
stone (Plummer, 1950, pl. 1; Bell, 1957, fig. 4); however, rocks in this area assigned by Plummer (1950, pl. 1) and Bell (1957, fig. 4) to the lower part of the Strawn are regarded by some other geologists (Turner, 1957a, p. 69) as Atoka in age. The Smithwick Shale is present in outcrops along the east side of the Llano uplift in Burnet and Blanco Counties. In the subsurface east and south of the Llano area, interval B thickens abruptly and grades laterally into mudstone and sandstone considered to be of Atoka age.

West of the Bend flexure, interbedded limestone, mudstone, and sandstone near the base of and beneath the Caddo lime, which is at the base of the Strawn Group (table 24), contain fusulinids of Atoka age and are assigned to interval B.

Interval B is missing or indistinguishable from overlying rocks on the northern and central part of the Bend arch, but it reappears to the west along an irregular and poorly determined line beneath the Midland basin. No formal names are applied to interval B there or in the subsurface elsewhere in west Texas.

In outcrops in the Marathon salient of the Ouachita structural belt, the interval consists of the Dimple Limestone of Atoka and probably late Morrow age (Sanderson and King, 1964, p. 32-33). In the Franklin Mountains of westernmost trans-Pecos Texas, it consists of the Berino Member (Atoka and early Des Moines age) of the Magdalena Formation; and in the nearby Hueco Mountains, it is composed of parts of the lower and middle divisions of the Magdalena Limestone. It consists of the lower part of an unnamed sequence of Pennsylvanian rocks in the Sierra Diablo, Culberson County (King, 1965, p. 43-45).

The middle part of the Dornick Hills Formation makes up interval B in the Marietta syncline in Cooke and Grayson Counties, and the Atoka Formation makes up the interval in the Hardeman basin.

UPPER BOUNDARY OF INTERVAL B

In outcrops on the north side of the Llano area, the top of interval B is the conformable contact of the dark-gray Smithwick Shale and the overlying, generally lighter colored mudstone, sandstone, and local conglomerate that are assigned to the Strawn Group or Series (Plummer, 1950, p. 86). Farther north, on the west side of the Bend flexure and along the west and north sides of the Fort Worth basin, the top of interval B lies within or at the base of the Caddo lime. In the deeper, eastern part of the basin, the Caddo lime pinches out within interbedded mudstone and sandstone into which the upper boundary of the interval is projected arbitrarily. Where the Caddo lime is missing in wells of Erath County and parts of nearby counties (Turner, 1957a, p. 69), a zone of brown micaceous mudstone in the upper part of interval B gives local control for picking the top of the interval. Cretaceous rocks unconformably overlie rocks of interval B farther east in the deepest part of the Fort Worth basin and in a narrow belt that extends along the east and south margins of the Llano area and from there to the eastern and central parts of the Kerr basin (fig. 45).

On the west side of the Kerr basin the upper boundary of interval B is placed at the top of limestone containing fusulinids of early or middle Atoka age. The overlying unfossiliferous detrital rocks may include strata of Atoka age, but these rocks are here assigned to higher Pennsylvanian intervals on the basis of regional stratigraphic relations.

Within the Val Verde basin, at the southern end of the Bend arch, and in the western part of the Midland basin, the boundary between intervals B and C is generally conformable and is at or near the Morrow-Atoka contact as determined locally from fusulinids. Where fusulinids are absent, the contact is located at the top of rocks that are dominantly mudstone and sandstone; these rocks are in contrast with rocks at the base of interval C that are dominantly limestone. This lithologic distinction becomes less clear cut toward the north, where the base of the Strawn Group of interval C contains appreciable detrital material.

In the subsurface of Culberson County, the boundary is drawn to include in interval B rocks that contain fusulinids of Atoka age. The contact is indicated by an increase in the proportion of limestone above the boundary as compared with that below.

Within the Magdalena Limestone, the boundary is the upper limit of rocks of Derry (approximately Atoka) age. This horizon falls within the middle division of the Magdalena (King and others, 1945) in the Hueco Mountains. The top of the Berino Member (Harbour, 1972) in the Franklin Mountains is used as the top of interval B.

The top of interval B in the Marathon area is the top of the Dimple Limestone, which is overlain conformably by the dominantly detrital Dorrondale Formation in interval C. The upper boundary of the Dimple is within rocks of probable Atoka age and therefore is stratigraphically somewhat lower than the Atoka-Strawn boundary picked as the top of interval B to the north and east.

In the Hardeman basin, the upper boundary of interval B is arbitrarily picked in the lower part of a sequence several hundred feet thick of interbedded limestone, mudstone, sandstone, and conglomerate. The boundary conforms, in general, to that used for separating rocks of Des Moines age from those of Atoka age in the adjacent panhandle region of Texas (G. H. Dixon, chap. J); however, no independent lithologic or faunal evidence was established to show that the boundary is mapped consistently in the Hardeman basin at the Atoka-Des Moines contact.
THICKNESS TRENDS

Interval B thickens eastward from an indefinitely located wedge edge west of the Bend flexure to an average of about 500 feet along the flexure, and from there it thickens at a much increased rate to more than 6,000 feet in Dallas County in the deepest part of the Fort Worth basin (pl. 6A). Interval B is truncated and covered by Cretaceous rocks in Dallas County and in nearby counties to the southeast; its maximum original thickness in the Fort Worth basin was probably substantially greater than 6,000 feet.

A northwest-trending fault along the southwest side of the Muenster arch and lesser northeast-trending faults in Llampasas, Coryell, and Hamilton Counties break the continuity of isopach lines at the north and south ends of the Fort Worth basin. Faults at the south end of the basin were active in post-Atoka time, and the thickness changes there are due to uneven truncation of interval B beneath Cretaceous beds in the displaced blocks.

East and south of the Llano uplift, interval B increases from a thin eroded edge to thicknesses of several hundred to several thousand feet in horizontal distances of 5-15 miles. In the southern part of the Kerr basin, more than 7,000 feet of Pennsylvanian may be mainly or wholly Atoka in age; no paleontological control, however, is available to justify showing such a thickness on plate 6A.

On the southern end of the Bend arch, interval B is irregular in thickness and distribution to a degree that cannot be shown on the scale of plate 6A. Thickness of the interval is generally less than 500 feet and commonly less than 50 feet; the interval thickens locally and abruptly in grabens, erosional depressions, and sinkholes (Rall and Rall, 1958, p. 848).

In the Marathon area, measurements on the outcropping edges of the folded and faulted Dimple Limestone suggest an originally lenticular mass having a maximum thickness of more than 1,000 feet. The formation thins northwestward to about 250 feet at the point where it disappears under the Glass Mountains, and it is less than 500 feet thick on the most southeasterly outcrops.

Few specific thicknesses of rock have been assigned to interval B in the western Val Verde and southern Delaware basin areas, and age ranges of these few are in doubt; but maximum thickness assignable to interval B in these areas and also in Culberson, Hudspeth, and El Paso Counties probably does not exceed a few hundred feet.

Thickness of interval B exceeds 500 feet on the southwest side of the Central Basin platform and in a narrow zone parallel to the east edge of the platform, but the interval thins rapidly eastward from these areas and is locally and perhaps entirely absent in the eastern part of Midland basin.

Rocks assigned to interval B thicken abruptly northward from near zero on the Red River arch to about 600 feet in the Hardeman basin, and they are at least 2,000 feet thick in the Marietta syncline northeast of the Muenster arch.

LITHOFACIES TRENDS

Interval B is dominantly sandstone and mudstone in the eastern part of the Fort Worth basin; it grades westward into rocks that show many local variations, but the rocks are dominantly limestone west of the Bend flexure (pl. 6B). The change in facies is due in large part to a westward transgression of limestone of the Big Saline Formation and its lithogenetic equivalents across the Bend flexure and onto the adjacent shelf, and also to a rapid westward thinning and virtual disappearance of the overlying detrital rocks of the Smithwick Shale and its lithogenetic equivalents in the same area.

Similar lithofacies trends are evident elsewhere in the region. Aprons of dominantly fine-grained detrital rocks that grade outward into carbonate rock form a border along the Ouachita belt east and south of the Llano area, adjacent to an ill-defined zero isopach in the central trans-Pecos area, south of the Matador arch in the southern part of the Texas Panhandle, and along parts of the Central Basin platform and Pecos arch.

The noncarbonate rock that is so well developed in many parts of a belt bordering the eastern and southern parts of the Ouachita geosyncline is absent in the Marathon area. Here the interval consists of limestone, dark terrigenous mudstone, and thin interbeds of black chert in the southern part and grades northward into clastic and oolitic limestone, limy mudstone, and conglomerates of chert and limestone.

No clear lithofacies pattern is evident in the Hardeman basin, except perhaps that caused by a somewhat greater concentration of limestone in the central part of the basin than along the edges. As shown by figure 48, conglomerate occurs in rocks of interval B in the Hardeman basin (fig. 45), and conglomerate is common in the northern parts of the Fort Worth basin. A few thin seams of coal are reported in interval B (fig. 48), mainly in the northern part of the Fort Worth basin adjacent to the Muenster arch and in the Hardeman basin.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Lithofacies patterns show that the great eastward-thickening wedge of detrital rock of interval B in the Fort Worth basin coarsens southeastward and thus probably came from the southeast from land areas within what is now the Ouachita structural belt. Pebby conglomerates and other detrital materials that fringe the Muenster and Red River arches suggest fairly vigorous erosion in local source areas of at least moderate relief on these
The Bend arch apparently was above sea level during much of the deposition of the interval, as shown by a gradual onlap of interval B across it from the east and the advance of B beyond A toward the west side of the arch. The area was too low topographically, however, to supply any appreciable amount of detritus to the nearby seas.

Coarse arkosic material in Yoakum and adjacent counties probably was carried southward in shallow water from island areas along the Matador arch (fig. 44), which stretched across the south end of the Texas Panhandle.

Conglomerate restricted to the northwestern and northern parts of the Dimple Limestone in the Marathon area may have been derived from a northern source raised within the Atoka Sea.

At the beginning of deposition of the interval, a shallow-water marine environment existed along the west side of the Fort Worth basin, as represented by bioclastic limestone and interbedded detrital rocks of the Big Saline Formation. It has been suggested that fossils and textures of the Big Saline in the Llano area indicate deposition of at least part of the Big Saline in the intertidal zone (Freeman, 1964). The shelf area was bordered on the east by deeper water in which the sediment that formed the Smithwick Shale and Atoka Series was deposited, perhaps partly by turbidity currents that flowed down the axis of the basin from the northeast (P. B. King, in Flawn and others, 1961, p. 184; Goldstein and Hendricks, 1962, p. 427; McBride and Kimberly, 1963). These environments gradually shifted westward during deposition of the interval; dark mud of the Smithwick buried the previously deposited calcium carbonate at the shelf and was in turn buried by mud and sand of the Atoka Series as the Fort Worth basin filled (Turner, 1957a, p. 67-69; Adams, 1962, p. 376). Most of the sediment coming in from the east was trapped east of the Bend flexure. Where the sea was present on the Bend arch and in the Hardeman basin, its waters remained clear, shallow, and warm throughout accumulation of rocks of the interval.

Sandstone, mudstone, and local coal beds in the northeastern part of the Fort Worth basin probably represent fluvial-deltaic deposits derived in part from the Muenster and Red River arches.

Depositional conditions similar to those in the eastern part of the Fort Worth basin existed south of the Llano area and in the nearby Kerr basin. Clastic limestone accumulated in shallow water on the south end of the Bend arch. In parts of this area, erosional depressions and active grabens (Rall and Rall, 1958, p. 847-848) trapped pockets of sediment including carbonate deposits, terrigenous mud, sand, and small amounts of coaly material.
In the Marathon area, the Dimple Limestone apparently was deposited during a temporary reduction in a long-continued influx of detrital sediment from southern highlands. Clastic limestone, locally crossbedded on a large scale, and chert conglomerate accumulated in shallow agitated water in the northern part of the area underlain by the Dimple. Deeper water to the south is indicated by the presence of pelagic mudstone interbedded with limestone that has graded bedding and internal structures of types believed to be produced by turbidity currents. Paleocurrent patterns deduced from directional structures suggest that the greater part of the turbidite carbonate sediment moved downslope from the northern shelf toward the basin axis, though some evidence of northward currents has been found on the south side of the area (Thompson and Thomasson, 1964, p. 24-29). Reworked shallow-water Foraminifera are present in the basin facies (Sanderson and King, 1964, p. 33).

Limy mud and argillaceous mud were deposited in a shallow foreland sea that stretched from the Midland basin westward across trans-Pecos Texas.

**PALEOTECTONIC IMPLICATIONS**

Compressive forces, directed northward and northwestward along the arc of the Ouachita-Marathon structural system, were active during interval B time. Deformation of sedimentary rocks along much of the Ouachita geosyncline in eastern Texas began early in Atoka time (Goldstein and Hendricks, 1962, p. 427; P.B. King, in Flawn and others, 1961, p. 187). The Fort Worth basin was one of a chain of rapidly sinking foredeeps that formed between the orogenic lands and the stable continental interior (Galley, 1958, p. 419; P. B. King, in Flawn and others, 1961, p. 187; Adams, 1962, p. 377). The Bend flexure formed a fairly abrupt hinge line at the edge of the relatively stable area. The Muenster, Red River, and Matador arches probably began to rise as transverse positive elements early in the interval (McBee and Vaughan, 1956, p. 360; Adams, 1962, p. 376), and folding or faulting along these elements may have produced island areas that had fairly abrupt local relief.

A northward-trending system of normal faults, called by Conselman (1954) the Fort Chadbourne system, was active during the time of interval B in the southern part of the Bend arch in Nolan, Coke, Tom Green, Schleicher, and Sutton Counties. Outliers of Atoka age occupy grabens along the northern part of this fault system (Conselman, 1954; Rall and Rall, 1958, p. 848-850); one such outlier is shown by the closed thickness contour in Coke County on plate 6A. A similar feature, possibly part of another graben system, is shown in western Sutton County. Stratigraphic displacement of as much as 1,000 feet was reported by Conselman (1954, p. 17) for some faults in the Fort Chadbourne system. The graben deposits were interpreted by Conselman (1954) as remnants of a once extensive blanket of Atoka rocks that were removed from the surrounding area by post-Atoka-pre-Des Moines erosion. An alternative suggestion (Rall and Rall, 1958, p. 850), however, is that despite the large displacements, the area never had much relief. According to this concept, faulting continued intermittently over a long period during which the sinking grabens were kept filled. Thus, the grabens are interpreted as narrow embayments or local basins in an area that was nearly level otherwise and that, except for the grabens, received little or no sediment in its northern part and only a thin cover in the southern section during the time of interval B.

South and west of the Fort Worth basin, the depositional zone in the Ouachita belt was compressed into a narrow downwarp south of resistant cratonic masses in the Llano area and in Edwards and Real Counties. The structurally deep Kerr basin formed in a reentrant between these buttresses. The Devils River positive element (fig. 45) in Val Verde and Terrell Counties may also have functioned as a buttress against which the Ouachita geosyncline was compressed (Flawn and others, 1961, p. 144-145); if Atoka rocks now exist south of the positive element, they have not yet been reached in drilling.

In the Marathon salient of the structural belt, the scarcity of terrigenous material in the Dimple Limestone suggests that tectonic activity was slight within the belt or south of it during the time of interval B. Local uplift north of the area of Dimple deposition may be indicated by the presence of chert conglomerate in the northern part of the formation.

Moderate downwarping took place along the trends of the Delaware and Midland basins in the broad platform north of the Marathon area.

**INTERVAL C FORMATIONS INCLUDED**

Interval C includes all rocks assigned to the Strawn Group in outcrops in the Colorado River drainage (San Saba, Mills, southern Brown, and western Lampasas Counties, fig. 43). The interval consists of, from bottom to top, the Kickapoo Creek, Millsap Lake, and Lone Camp Groups in the Brazos River drainage (Parker, Palo Pinto, and Erath Counties). These groups make up the Strawn Series as used by most Texas geologists, a term generally used synonymously with the Des Moines Series of the midcontinent area (Turner, 1957b, p. 89; Brown, 1959, p. 2867). Many formations and members have been named in both the Colorado and the Brazos River areas, as shown by table 24. With a few exceptions, individual beds are lenticular and discontinuous (Plummer, 1950, p. 86; Turner, 1957a, p. 71), and formational units are difficult to trace on a lithologic basis for any great distance from outcrops into the subsurface.
On the Muenster arch and in the Marietta syncline in Cooke and Grayson Counties, rocks of interval C comprise the Deese Formation and the upper part of the underlying Dornick Hills Formation (Bradfield, 1957a, p. 34-40). In the Hardeman basin, interval C consists of unnamed rocks considered to be Des Moines in age.

Interval C in the Marathon area consists of the Haymond Formation of late Atoka and possible Des Moines age (King, 1937, p. 72; Skinner and Wilde, 1954, p. 796; McBride, 1966, p. 15-16) and the lower (Des Moines) part of the Gaptank Formation (King, 1937, p. 76-77). In the Hueco Mountains, the interval is that part of the Magdalena Limestone which is of Des Moines age; in the Franklin Mountains, it is the Bishop Cap Member (Des Moines and younger) of the Magdalena Formation. Elsewhere in western Texas, rocks assigned to the Strawn Group or the Des Moines Series make up the interval. The term lower Strawn limestone is used by some geologists for a widespread limestone that formed in much of the central part of the central and west Texas regions early in Des Moines time. In the Midland basin and in some adjacent areas, this unit composes most of the interval, the remainder being an upper thin mudstone unit.

**UPPER BOUNDARY OF INTERVAL C**

The upper boundary of interval C in most of the Colorado River drainage basin is placed at the top of the Capps Limestone Lentil, and in the Brazos River drainage basin, at the base of the Lake Pinto Sandstone Member of the Salesville Formation (interval D, table 25). Where the Lake Pinto Sandstone Member is so thin and inconspicuous that it cannot be identified, the upper boundary is the top of the Village Bend Limestone Member of the East Mountain Shale (table 24). South of the Colorado River, in parts of McCulloch County, the Capps is missing, and the base of the Rochelle conglomerate, a probable channel-fill unit at the base of the Brownwood Shale Member of the Graford Formation (table 25), or equivalent sandy beds is the top of the interval. The boundary as thus drawn corresponds closely to the boundary between the Des Moines and Missouri

---

**Table 24. — Some named stratigraphic units in interval C, north-central Texas**

<table>
<thead>
<tr>
<th>Colorado River drainage (Eagle, 1909)</th>
<th>Brazos River drainage (Abilene Geol. Soc., 1949)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Strawn</strong></td>
<td><strong>Strawn</strong></td>
</tr>
<tr>
<td>Capps Limestone Lentil of Plummer and Moore (1922)</td>
<td>Lone Mountain</td>
</tr>
<tr>
<td>Shale</td>
<td>East Mountain</td>
</tr>
<tr>
<td>Ricker Sandstone Member of Nickell (in Lee and others, 1938)</td>
<td><strong>Lone Camp</strong></td>
</tr>
<tr>
<td>Shale and sandstone</td>
<td>Village Bend Limestone</td>
</tr>
<tr>
<td>Ricker Station Limestone of Cheney (1949)</td>
<td>Hog Mountain Sandstone</td>
</tr>
<tr>
<td>Sandstone and shale undivided</td>
<td>Brazos River Sandstone</td>
</tr>
<tr>
<td></td>
<td>Garner</td>
</tr>
<tr>
<td></td>
<td>Mingus Shale (includes Thurber coal bed)</td>
</tr>
<tr>
<td><strong>Grindstone Creek</strong></td>
<td><strong>Lazy Bend</strong></td>
</tr>
<tr>
<td><strong>Millsap Lake</strong></td>
<td><strong>Lazy Bend</strong></td>
</tr>
<tr>
<td>Goen Limestone</td>
<td>Brannon Bridge Ls.</td>
</tr>
<tr>
<td>Santo Limestone</td>
<td>Steussy</td>
</tr>
<tr>
<td><strong>Kickapoo Falls Ls.</strong></td>
<td>Meek Bend Ls.</td>
</tr>
<tr>
<td></td>
<td>Hill Creek</td>
</tr>
<tr>
<td></td>
<td>Dennis Bridge Ls.</td>
</tr>
<tr>
<td><strong>Kickapoo Creek</strong></td>
<td><strong>Kickapoo Creek</strong></td>
</tr>
<tr>
<td>Rayville</td>
<td>Rayville</td>
</tr>
<tr>
<td>Parks</td>
<td>Parks</td>
</tr>
<tr>
<td>Caddo Pool</td>
<td>Caddo Limestone</td>
</tr>
</tbody>
</table>

Correlations with units in Brazos River drainage not established.
Series of the midcontinent region (Cheney, 1940, p. 66; Plummer, 1950, p. 88; Shelton, 1958, p. 1518). Beds at the boundary are virtually conformable (Shelton, 1958, p. 1518; Adams, 1962, p. 377), except in McCulloch County where the *Rochelle conglom erate* or equivalent rocks overlap the Strawn Group and older beds and locally rest on the Ellenburger Group of Ordovician age (Plummer, 1950, p. 94; Cloud and Barnes, 1948, p. 60).

In the subsurface in northern Cooke and Grayson Counties, the top of the interval is the top of the *Deese Formation*, which is overlain conformably by the *Confederate Limestone Member* of the *Hoxbar Formation*.

In the Hardeman basin, the upper boundary is within a conformable sequence of limestone and mudstone and is picked at the top of an unnamed limestone bed correlated westward with a limestone at the top of the zone of *Fusulina*, according to reports of the Paleontological Laboratory of Midland, Tex. The top of the zone of *Fusulina* is also the top of the interval along the east side of the Midland basin, where paleontological control is available.

Cretaceous rocks unconformably truncate and overlies rocks of interval C in a band averaging about 40 miles wide that crosses the Fort Worth basin from San Saba County northeastward to northwestern Grayson County, and from there northward into Oklahoma.

Limestone of Des Moines age is overlain conformably by limestone of Missouri age in much of central west Texas and trans-Pecos Texas. A similar relation occurs, also, in many limestone banks and reefs in the vicinity of the Central Basin platform and in the Horseshoe atoll in the northern Midland basin. In all these areas, the boundary between intervals C and D is the boundary between the Des Moines and Missouri Series, as determined from fusulinids or by the tracing of lithologic markers for short distances away from wells in which fossils have been reported.

Locally along the west side of the Midland basin and on part of the Pecos arch, rocks of known or probable Wolfcamp age directly overlie limestone of Des Moines age. In the central Midland basin and in part of the Val Verde basin area to the south and southwest, the upper limit of interval C is at the estimated top of Des Moines rocks within a thin dark mudstone unit, probably of Middle or Late Pennsylvanian age, that rests on the *lower Strawn limestone*. This mudstone, less than 100 feet thick in much of the Midland basin to a possible few hundred feet thick in the Val Verde, was included with overlying rocks of Wolfcamp age in maps of the Permian System (McKee, Oriel, and others, 1967a, pls. 3, 7). On the isopach map for Pennsylvanian interval C (pl. 7A), thicknesses in the starved area of the Midland and Val Verde basins are shown by a specific figure for the *lower Strawn* and a plus sign for the unknown thickness of overlying mudstone of Strawn age (for example, 250 ft.+). In cross section *CC-C'C'* (pl. 10E), however, the starved-facies mudstone of probable Pennsylvanian age has been divided arbitrarily, and upper boundaries of intervals C, D, and E are shown in the mudstone.

The Des Moines-Missouri boundary is at an undetermined position in the lower part of a sandstone and mudstone sequence in the Gaptank Formation in the Marathon area. To the east in Edwards and adjacent counties, the boundary is within a detrital sequence that overlies fossiliferous *lower Strawn limestone* in most wells and is assigned to the Middle and Upper Pennsylvanian on the basis of regional relations and because of Missouri or Virgil fusulinids in a few wells. No interval boundaries were designated for these areas; however, the rocks are shown as present on maps of intervals C, D, and E.

**THICKNESS TRENDS**

Interval C has a maximum thickness of nearly 4,500 feet in the northeastern part of the Fort Worth basin in northeastern Wise County and the adjacent part of Denton County (pl. 7A). It thins rapidly from this area westward and southwestward across the Fort Worth basin to about 1,500 feet along the east side of the Bend arch, and from there it thins at a reduced rate to less than 100 feet at places on the east side of the Midland basin. Interval C is missing at places on the Red River arch in northern Montague, Wichita, and Clay Counties, where it is overlapped by rocks of interval D. Interval C has a maximum thickness of about 1,600 feet in the Texas part of the Hardeman basin. Its lower part is as much as 9,300 feet thick north of the Muenster arch in the Marietta syncline along the Texas-Oklahoma border; the upper part has been removed by pre-Cretaceous erosion along the axis of the Marietta syncline, and the maximum original thickness, therefore, was greater.

Abrupt thickness changes along the north side and at the south end of the Fort Worth basin are due to faults that had several hundred to several thousand feet of movement during deposition of interval C.

Rocks in excess of 1,500 feet in thickness are probably assignable to interval C in the deepest part of the Kerr basin. North and west of the Kerr basin, the interval thins to less than 500 feet in Kimble, Sutton, and Edwards Counties; it is thicker farther west, in Val Verde County. In the Marathon area, interval C may have a present maximum thickness of between 4,000 and 5,000 feet, as estimated from several outcrop sections (Texas locs. 2679, 2680, and elsewhere).

In much of the Midland basin and in areas southwest and west of the basin, the interval is less than 500 feet thick. Figures shown in the central part of the Midland basin (pl. 7A) are for only the *lower Strawn limestone*; an undetermined part of an overlying Middle and Upper Pennsylvanian mudstone, commonly less than 100 feet
thick, may also be of Des Moines age. On the margins of the Central Basin platform and in the Horseshoe atoll, thicknesses between 500 and 1,000 feet include reef or bank limestone representing a late part of the Des Moines Series, as well as the blanket limestone of earlier Des Moines age.

LITHOSERIES TRENDS

Patterns of sedimentation established during intervals A and B were maintained during the early part of interval C. Sandstone and mudstone predominate in much of the Fort Worth basin, but the interval is represented westward in the Midland basin by rocks that are mostly limestone (pl. 7B; figs. 45, 49). This facies distribution is due in part to a very abrupt westward thinning of the upper, detrital part of the interval at the east edge of the Midland basin. Complex facies patterns characterize the area of the Bend arch, between the two areas of more uniform lithology, and the Hardeman basin farther north, where beds of limestone, sandstone, and mudstone interfinger complexly in the part of the interval above the Caddo lime. A discontinuous belt consisting almost entirely of limestone in Wichita, Wilbarger, and southern Foard Counties corresponds in general to areas on the Red River arch where interval C is thin and where it represents only late Des Moines time.

Interval C contains many conglomerates in the region bordering the Red River and Muenster arches and also in a few patchy areas in the central part of the Fort Worth basin (figs. 45, 50). These conglomerates are most continuous and widespread in the basal part of the interval, but they also occur in the upper part. Red beds form an irregular band, widest to the northeast, that extends from Concho and McCulloch Counties northward to Cooke and Grayson Counties (fig. 50).

Coal occurs at several horizons in the northern part of the Fort Worth basin and adjacent parts of the Muenster arch and in an isolated area farther south near the junction of Eastland, Brown, and Callahan Counties (figs. 43, 50).

Limestone and smaller amounts of mudstone are widespread, not only throughout the Midland basin but also on the adjoining shelves, on the Central Basin platform, and in much of the area where interval C is present in westernmost Texas. The amount of mudstone and sandstone exceeds that of limestone along the southern margin of the region, and mudstone composes most of the interval in the southern part of the Delaware basin and probably along part of the Diablo platform. Correlations in the Diablo area are uncertain.

SOURCES AND ENVIRONMENTS OF DEPOSITION

A coarsening in the grain size of rocks eastward in interval C and the presence of conglomerates, coal, and red beds in the eastern or northeastern parts of the region show that land areas east or northeast of the Fort Worth basin were the principal contributors of sediment to the Fort Worth basin, Marietta syncline, and nearby areas to the west. In early Des Moines time, the Muenster and eastern part of the Red River arches also provided some sediment to adjacent areas, as shown by conglomerates that fan out southward from these structural features and by a gradual onlap of interval C southwestward from the Marietta syncline onto the Muenster arch (Bradfield, 1957a, fig. 1; 1957b, figs. 6, 7). The types of chert in gravels of the Brazos River Sandstone Member of the Garner Formation and other conglomerates of interval C are similar to, if not identical with, those in strata of Ordovician and older rocks in the Ouachita Mountains of southern Oklahoma (Bay, 1932, p. 182-184). Thus, lower Paleozoic rocks probably were broadly exposed by erosion in the Ouachita structural belt and were the source of the chert.

In central Texas, environments of deposition during the early part of the interval differed only slightly from those at the end of Atoka time. Unfossiliferous dark-gray mudstone and interbedded siltstone and sandstone in the lower part of interval C in the southeastern part of the Fort Worth basin resemble rocks of interval B to the north. Their presence suggests that deposition was in moderately deep water along the east side of the basin in early Des Moines time. The sea apparently shallowed westward, and lime-secreting organisms, which were responsible for producing the Caddo lime (fig. 47), probably flourished in clear warm aerated water that must have characterized most of the rest of the region. The deep-water part of the Fort Worth basin filled gradually; sand and clay spread westward from the basin area and covered limestone on the adjacent shelf. By mid-Des Moines time, lagoonal and deltaic environments, characterized by thin coal beds, red beds, and lenticular channel sandstones, had been established nearly across the Fort Worth basin. Shallow-water neritic conditions probably persisted in the Bend arch areas, as shown by the local presence of limestone banks and reefs in the predominantly sandstone-mudstone sequence lying above the Caddo lime.

Rocks assigned to interval C in the central part of the Kerr basin suggest continuous building up of a muddy and, locally, a sandy floor. The western and northern margins of the basin in early Des Moines time were sites of carbonate deposition which was terminated by the spreading of mud and sand into these areas in later Des Moines time.

In interval C of the Marathon area, directional indicators in sedimentary structures show that most of the sediment was transported from an eastern or southeastern source and a smaller amount from the west (McBride, 1964a, p. 39). Well-rounded pebbles and
cobbles in the upper third of the Haymond Formation were derived in part from igneous and metamorphic rocks not known to be present in the Marathon area. Other coarse constituents, ranging in size from small fragments to massive blocks tens of feet long, came from the Dimple, Tesnus, and older Paleozoic formations. The large blocks have been explained by King (1958, p. 1734) as derived from unstable, probably faulted margins of a deepwater trough by "subaqueous landslips which developed proximally into turbid flows." Other interpretations of the Haymond boulder beds were reviewed by King (1958).

The Haymond Formation, excepting beds transitional into the overlying Gaptank Formation, is considered to have been deposited in deep water of normal salinity. The lower two-thirds accumulated mainly as alternating thin layers of mud of probable pelagic origin and carbonaceous sand that may have been brought in by turbid flow from shallower, marginal waters (P. B. King, in Flawn and others, 1961; McBride, 1964a, p. 38); a few thick coarse-grained arkosic sand units interrupt the rhythmic sequence of fine-grained deposits. The upper third of the formation, including the exotic pebbles and cobbles and the boulder-bed blocks and fragments previously noted, is dominantly coarse grained. The textural change presumably reflects both subaqueous faulting nearby and uplift in subaerial source areas. Also, it may indicate an increase in subaqueous slumping and turbid flow from shallow into nearby deep water.

In beds transitional between the Haymond and overlying Gaptank Formations, deepwater poorly fossiliferous mudstones typical of the Haymond are succeeded by limestone, conglomerate, and mudstone containing shallow-water corals, brachiopods, and fusulinids (King, 1937, p. 76-78).

Much of the Texas interior beyond the detrital apron that bordered the Ouachita belt probably was covered by a warm shallow sea in early Des Moines time, as in-
Figure 50. — Distribution of conglomerate, red beds, and coal in interval C, north-central Texas.
PALEOTECTONIC IMPLICATIONS

Compressional deformation in the Ouachita belt adjacent to the Fort Worth basin culminated during interval C time (Barnes, 1956, p. 6; Tomlinson and McBee, 1959, p. 40; P. B. King, in Flawn and others, 1961, p. 188). Ordovician rocks overlie rocks of interval C in Grayson County; westward thrusting, therefore, was no earlier than the latter part of De Moine time in that area. Subsequently, major movements in the Ouachita structural belt east of the Fort Worth basin seemingly consisted of simple uplift (Cheney and Goss, 1952, p. 2258).

Present knowledge of geology along much of the southern border of central and west Texas provides little definitive information about the extent and degree of tectonic activity during the deposition of interval C. Detrital sediments in the Kerr basin and Marathon area indicate at least moderate uplift south of the geosyncline. In the Marathon area, downwarping of the Marathon-Ouachita geosynclinal belt culminated during deposition of the Haymond Formation. Major faulting along the margins of the deepening Marathon geosyncline is inferred from the massive blocks and fragments of pre-Haymond Pennsylvanian and older Paleozoic formations in the Haymond boulder beds.

The Fort Worth basin apparently sank rapidly during early Des Moines time. The rate of sinking decreased later, and the deeper parts of the basin gradually filled. Sediment produced by uplift and erosion in the structural belt was largely trapped in the basin or on the adjacent Bend arch (Adams and others, 1951). A reconstruction of interval C, from the Bend arch westward onto the Central Basin platform at the end of C time (pl. 10E, sec. CC-C'C'), suggests that subsidence occurred across the entire area but that it was notably less for the Central Basin platform than for the Midland basin and the Bend arch. The development of the Midland basin in later Des Moines time seems to have resulted from folding on the west side; but on the east side it was due largely to a great buildup of sediment on the basin margin as compared with almost no deposition of sediment within the basin itself. A marked topographic relief had developed between the eastern shelf and the adjacent starved basin by the end of deposition of interval C.

During the latter part of interval C time, the Val Verde basin began to form as a new foredeep along the northern margin of the Marathon area and from there eastward into Terrell and Val Verde Counties. The initiation of the Val Verde basin thus approximately coincided with the beginning of downwarping along the west side of the Midland basin.

The limited data available indicate that the Delaware basin was mildly negative in Des Moines time.

No great degree of uplift during interval C time is evident in the interbasin areas of west Texas, although the southern part of the Diablo platform supplied some sediment to the Haymond Formation and therefore must have been slightly elevated.

The Muenster and Red River arches were intermittently raised during interval C time. Uplifting of the Muenster arch, accompanied by faulting along the flanks, was coupled with accelerated downwarping in the Marietta syncline to the north. The Red River and Muenster arches were beveled by erosion early in the interval and were largely buried by the end of the interval.

In the northeastern part of the Llano area and extending from there northeastward into the Fort Worth basin, several roughly parallel faults produced a set of narrow grabens and horsts in a zone more than 100 miles long and as much as 60 miles wide. The time of faulting was early in the Des Moines, for rocks of the Kickapoo Creek Group locally are displaced several hundred feet, whereas younger rocks of Des Moines age either are not faulted or are affected only by minor fault movements (Cheney, 1940, p. 105; Abilene Geol. Soc., 1949). Movement during Des Moines time is indicated, also, by the preservation, among rocks of late Des Moines age, of fissiliferous conglomerate that contains pebbles of early Des Moines age, presumably derived from one of the
fault scarps at a locality in McCulloch County (Freeman and Wilde, 1964).

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D consists of the Canyon Group (Canyon Series of Texas geologists), and it corresponds closely to the Missouri Series of the midcontinent region. Formations included in interval D in outcrops of north-central Texas are shown in table 25. The distribution of four widespread limestone units in this area is shown by figure 51.

On the Muenster arch and in the Marietta syncline, interval D consists of the Hoxbar Formation, which has the *Confederate Limestone Member* at its base (Tomlinson, 1934, p. 1085; McBee and Vaughan, 1956, p. 361; Tomlinson and McBee, 1959, p. 40). The base of the *Confederate Limestone Member* is selected as a suitable marker for the base of interval D in the subsurface in Cooke, Montague, and Grayson Counties, even though some authors place the base of the Missouri Series somewhat lower stratigraphically (Bradfield, 1957a, p. 67).

Through much of south-central and west Texas no formation names have been applied within the Canyon Group. Rocks of Canyon age constitute the middle part of the Gaptank Formation in the Marathon area. The Canyon part of the upper member of the Magdalena Formation is assigned to interval D in the Franklin Mountains (Harbour, 1972), and part of the upper division of the Magdalena Limestone is assigned to interval D in the Hueco Mountains (King and others, 1945).

**UPPER BOUNDARY OF INTERVAL D**

The top of interval D is the top of the Home Creek Limestone Member of the Caddo Creek Formation in much of north-central Texas. The contact with overlying rocks is conformable except where the Home Creek is cut out by local sandstone channels.

The top of interval D is a conformable contact between interbedded sandstone, mudstone, and thin limestone beds at the top of the Hoxbar Formation and coarse sandstone and conglomerate of the Pontotoc Formation in northern Cooke and Grayson Counties (Bradfield, 1957b, p. 85).

Interval D unconformably underlies Cretaceous rocks in parts of Cooke and Grayson Counties north of the Fort Worth basin and at a few places farther south, and interval D is exposed at places where Cretaceous rocks are stripped away along the west side of the Fort Worth basin.

Along the east side of the Midland basin, the upper boundary of the interval is the top of a limestone se-

---

**Table 25. Some named stratigraphic units in interval D, north-central Texas**

<table>
<thead>
<tr>
<th>Group</th>
<th>Formation</th>
<th>Member</th>
<th>Group</th>
<th>Formation</th>
<th>Member</th>
</tr>
</thead>
<tbody>
<tr>
<td>Caddo Creek</td>
<td>Home Creek Limestone</td>
<td></td>
<td>Caddo Creek</td>
<td>Home Creek Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Colony Creek Shale</td>
<td></td>
<td></td>
<td>Colony Creek Shale</td>
<td></td>
</tr>
<tr>
<td>Brad</td>
<td>Ranger Limestone</td>
<td>Placid Shale</td>
<td>Brad</td>
<td>Ranger Limestone</td>
<td>Placid Shale</td>
</tr>
<tr>
<td>Winchell Limestone</td>
<td>Upper</td>
<td></td>
<td></td>
<td>Winchell Limestone or Merriman Limestone</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Upper</td>
<td></td>
<td></td>
<td>Wolf Mountain Shale</td>
<td>Adams Branch Limestone</td>
</tr>
<tr>
<td></td>
<td>Lower</td>
<td></td>
<td></td>
<td>Palo Pinto Limestone</td>
<td>Wiles Limestone, Posideon, and Wynn Limestone</td>
</tr>
<tr>
<td>Canyon</td>
<td>Cedarton Shale</td>
<td>Adams Branch Limestone</td>
<td></td>
<td>Keechi Creek</td>
<td>Turkey Creek Sandstone</td>
</tr>
<tr>
<td></td>
<td>Brownwood Shale (includes <em>Rochelle Conglomerate</em> at base)</td>
<td></td>
<td></td>
<td>Salesville</td>
<td>Dog Bend Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Lake Pinto Sandstone</td>
</tr>
</tbody>
</table>
Figure 51. — Distribution of some limestone units in interval D, north-central Texas.
sequence that is reported to contain fusulinids of Missouri age (Paleont. Lab., Midland, Tex.). Unfossiliferous mudstone overlies this limestone with apparent conformity at most places. Elsewhere in much of the Midland basin, all of Missouri time is believed to be represented within the lower few feet or few tens of feet of unfossiliferous mudstone that overlies the lower Straun limestone of early Des Moines age. In the thick limestone sequence of the Horseshoe atoll at the north end of the basin (Stafford, 1959, p. 8) and in the banks or reefs of about the same age along the west side of the basin and on the Central Basin platform (Paleont. Lab., Midland, Tex.), the top of interval D was determined paleontologically to conform with the Missouri-Virgil Series boundary of the midcontinent region.

In the Val Verde basin and the southern part of the Delaware basin, the top of the interval is within unfossiliferous mudstone probably equivalent to that overlying the lower Straun limestone in the central part of the Midland basin. The position of the Missouri-Virgil boundary has not been determined accurately within the wedge of Upper Pennsylvanian detrital rocks in the Kerr basin, in the Gaptank Formation in the Marathon area, or in the Magdalena Formation in westernmost Texas. In these areas, either the interval boundary is picked arbitrarily or rocks of the interval are indicated as present but have no assigned thickness.

THICKNESS TRENDS

Interval D has a maximum thickness of about 2,300 feet near the Texas-Oklahoma border in Montague and Cooke Counties (pl. 84). It thins gradually from this area westward and southwestward to between 600 and 800 feet at the west edge of the Bend arch. A narrow northeast-trending zone in which thinning is abrupt marks the west edge of the shelf and the east edge of the starved Midland basin. Interval D is generally less than 200 feet thick west of this zone in Sterling and Mitchell Counties and less than 100 feet thick in the western part of the basin. Zero isopachs tentatively drawn near the west edge of the Midland basin (pl. 8A) indicate areas of little or no accumulation of sediment, perhaps in current-swept areas at the foot of limestone banks that rimmed the Central Basin platform (compare pl. 11) or on the faces of the banks.

Thickness of interval D locally approaches or exceeds 500 feet in the western part of the Horseshoe atoll and along the east edge of the Central Basin platform; limestone thickness reaches 700 feet on the south side of the platform. Elsewhere in west Texas, except, possibly, areas of deposition marginal to the Ouachita structural belt, the interval is generally less than 500 feet thick. In extensive areas on the Central Basin platform and the Pecos arch and in trans-Pecos Texas, interval D was eroded in latest Pennsylvanian or earliest Permian time.

LITHOFACIES TRENDS

Interval D is mostly limestone where it occurs on parts of the Red River arch and in the two discontinuous northeast-trending bands in the central and southern parts of the Bend arch (figs. 45, 52; pl. 8B). These areas are partly surrounded by other areas that have a higher percentage of clastic rocks, and these clastic rocks are coarsest, in general, to the northeast. Red beds occur discontinuously along the west margin of the Fort Worth basin, and conglomerates and a few thin coal beds occur mostly in the northeastern parts of the region (figs. 45, 53).

Limestone is the principal component of interval D in the Horseshoe atoll, on and near the Central Basin platform, and in El Paso and western Hudspeth Counties. In El Paso County, gypsum is also present. Limestone is abundant in the Gaptank Formation in the Marathon area and in areas to the north in Culberson County. In contrast, throughout the greater part of the Midland, southern Delaware, and Val Verde basins, interval D is more than 80 percent mudstone and commonly contains no limestone. Sandstone exceeds mudstone southeast of the Val Verde basin in Edwards County.

SOURCES AND ENVIRONMENTS OF DEPOSITION

A northeastward increase in the proportion of sandstone and the occurrence of conglomerate along the north and east sides of the Fort Worth basin indicate that highlands existed in the region of the Ouachita structural belt in east Texas and in the Wichita Mountains area in southern Oklahoma (Tomlinson and McBea, 1959, p. 44) and furnished debris to northcentral Texas during interval D time. Sediment flooding out onto the eastern shelf of the Midland basin was trapped there in shallow lagoons and on delta fronts. Only minor amounts of detritus reached deep water of the Midland basin (Adams and others, 1951; Van Siclen, 1958; Jackson, 1964, p. 322). Comparison of sediment thicknesses of intervals C and D, combined, on the platform and in the Midland basin to the west suggests that by the end of Missouri time the west side of the Bend arch was a west-facing submarine slope having topographic relief on the order of 1,000 feet in 20 miles in western Coke and Tom Green Counties and a slope having somewhat less relief in areas to the north and south.

A rhythmic pattern of deposition on the eastern shelf of the Midland basin is apparent in interval D in the repetition of thick units of mudstone and sandstone interstratified with four main units of limestone (Plummer and Moore, 1922, p. 90). The Palo Pinto (or Adams Branch Limestone Member of Wolf Mountain Shale), Winchell, Ranger, and Home Creek Limestones each appear to represent the end of a regional regressive-transgressive cycle. Two cycles of deposition have been recognized in Canyon rocks to the south in Schleicher
and Sutton Counties (Rall and Rall, 1958, p. 857). At least two periods in which sea level was lowered during deposition of interval D have been inferred by Burnside (1959, p. 31) and Stafford (1959, p. 14-15) from leached zones and wave-cut benches within the Horseshoe atoll. Burnside (1959, p. 30) concluded from the shape and orientation of the Horseshoe atoll and from its fossils that the sea in the northern part of the Midland basin was clear and warm and of normal salinity and that a fairly constant wind and current came from the south or southwest.

Limestone accumulated on the Central Basin platform under environmental conditions that must have been similar to those inferred for the Horseshoe atoll. Island sources of sediment on the Central Basin platform at times supplied material that formed tongues of sandstone and of red and green mudstone. Fluctuations in sea level periodically interrupted growth of lime-secreting organisms on the limestone banks.

A muddy bottom similar to that of the deeper, starved parts of the Midland basin existed throughout Canyon time in much of the Val Verde basin and the southern Delaware basin. Moderately deep, stagnant water is inferred for these areas. Sand as well as mud in large quantity, presumably from a source within the Ouachita belt to the south, entered both the Kerr basin and the easternmost part of the Val Verde basin.

In the Marathon area, conglomerate beds that wedge out in short distances northward suggest a nearby southern source for Gaptank components in interval D (King, 1931, p. 44). The conglomerate is interbedded with lenticular deposits of mudstone, sandstone, and clastic limestone, in part crossbedded and containing shallow-water fossils (Ross and Oana, 1961, p. 243; McBride, 1964b, p. 44).

Seas in westernmost Texas (El Paso and Hudspeth Counties) probably were shallow, agitated, and well aerated, as shown by locally conglomeratic and biostromal limestone exposed in the Hueco Mountains (King and others, 1945). Interbedded limestone, mud-

---

**Figure 52.** Percentage of limestone in interval D, north-central and part of west Texas. Contour interval 20 percent. Shaded where interval D is absent.
Figure 53. — Distribution of red beds, conglomerate, and coal in interval D, north-central Texas.
stone, and, locally, conglomeratic sandstone that crop out in the Franklin Mountains (Harbour, 1972) indicate periodic incursions of muddy water in seas that otherwise were clear. Gypsum in limited amounts in the Franklin Mountains suggests that parts of the region at times were cut off from the open sea by reefs or other barriers.

PALEOTECTONIC IMPLICATIONS

The Muenster arch was buried during Missouri time and ceased thereafter to have any effect on sedimentation. Local thinning of interval D on the crest of the former Red River arch and growth of limestone reefs along this trend indicate that subsidence was somewhat retarded and that this area probably remained near sea level during the early part of the interval.

Low anticlinal folds in Canyon and older Pennsylvanian rocks along the trend of the Fort Chadbourne fault system suggest local reactivation of deep-seated faults along the system during Missouri time (Conselman, 1954, p. 20).

The broad region from the Bend arch westward into trans-Pecos Texas continued to sink intermittently. This movement, together with uplift in the Ouachita belt to the east, produced a regional westward slope, opposite to the direction of previous eastward tilting along the west side of the Fort Worth basin (Cheney, 1929, p. 10). Sedimentation generally kept pace with downwarping on the Bend arch (now functioning as an eastern shelf of the Midland basin), on the limestone banks and reefs of the Horseshoe atoll, and on the Central Basin platform but not with downwarping on the basin floors.

Conglomeratic deposits in the Franklin Mountains and in the Gaptank Formation in the Marathon area indicate local episodic uplifts. The presence in the upper part of the Gaptank of limestone cobbles from the Dimple and probably from the lowermost part of the Gaptank itself shows a northward advance of uplift within the Marathon segment of the Ouachita belt from the interior zone of the structural belt into the zone of late geosynclinal deposits.

INTERVAL E
FORMATIONS INCLUDED

Interval E consists of the Cisco Group in most of north-central Texas. The lower part of the Pontotoc Formation makes up the interval along the Texas-Oklahoma border in northern Cooke and Grayson Counties.

Some of the named stratigraphic units in interval E in the area of the Bend arch (fig. 45) are listed in table 26, and the distribution of four of the more persistent and widespread limestone units is shown in figure 54. Rocks of Cisco age are present but not divided into formations in the basin and platform areas that lie west and south of the Bend arch. The Gaptank Formation in the Marathon area and the upper member (or upper division) of the Magdalena Formation (or Limestone) in westernmost trans-Pecos Texas include strata of Cisco age that are assigned to interval D.

UPPER BOUNDARY OF INTERVAL E

At most places in the area of the Bend arch, the top of the Chaffin Limestone Member of the Thrifty Formation or the top of its correlative, the Crystal Falls Limestone Member of the Thrifty Formation, is the top of interval E. The systemic boundary between Pennsylvanian and Permian rocks is placed by Cheney (1940, p. 91) and L. G. Henbest (in Eargle, 1960, p. 75) a few feet above the Chaffin, within a continuously deposited sequence of mudstone assigned to the Waldrip Shale Member of the Pueblo Formation by Eargle (1960, p. 75-76); however, for convenience in mapping and correlation for this report the top of the limestone is picked as the top of the interval.

Limestone beds in the stratigraphic position of the Chaffin or Crystal Falls pinch out northward in counties that border Oklahoma. The top of interval E in these areas is projected from nearby areas at a consistent stratigraphic position with reference to correlated beds in the overlying or underlying rocks.

The position of the Pennsylvanian-Permian boundary in the Midland basin is known only within broad limits. On the east side of the basin, several hundred feet of unfossiliferous dark-gray mudstone and local sandstone separate limestone units containing Missouri fusulinids from those containing Wolfcamp (Permian) fusulinids (Paleont. Lab., Midland, Tex.). Tracing of limestone and sandstone beds along the east side of the basin has led several workers (Van Siclen, 1958; Rall and Rall, 1958; Jackson, 1964) to conclude that most, if not all, of the mudstone sequence is Permian in age. The same interpretation is made in the present investigation. The top of interval E accordingly is placed within the mudstone sequence, near its base. This boundary differs locally from that chosen for the base of the Permian System by D. A. Myers (in McKee, Oriel, and others, 1967b), who considered the mudstone and interbedded sandstone at most places near the east margin of the basin to be mostly Late Pennsylvanian (Missouri and Virgil) in age.

In most of the central and western parts of the Midland basin, the Pennsylvanian-Permian boundary is drawn at the top of a thin mudstone unit that overlies limestone of early Des Moines age and that is overlain by thick mudstone and sandstone of Permian age. Locally on the western side of the basin, Wolfcamp fusulinids have been identified from samples taken only a few feet or a few tens of feet above the lower Strawn limestone (Paleont. Lab., Midland, Tex.).
Chaffin Limestone Member or Crystal Falls Limestone Member of Thrifty Formation and approximately equivalent limestone beds

Gunsight Limestone Member of Graham Formation

Ivan Limestone Member of Graham Formation

Bunger Limestone Member of Graham Formation

Figure 54. — Distribution of some limestone units in interval E, north-central Texas.
In the Horseshoe atoll and in the limestone banks on the Central Basin platform, the upper boundary of interval E is placed at the top of rocks of Virgil age, as determined from fusulinids (Stafford, 1959; Paleont. Lab., Midland, Tex.).

By analogy with the Midland basin, the boundary between Pennsylvanian and Permian rocks in the central part of the Val Verde basin is considered to be at the top of a mudstone unit, a few hundred feet thick at most, that was deposited between the lower Strawn limestone and several thousand feet of mudstone and sandstone commonly assigned a Wolfcamp age. Fossils are generally lacking. The mudstone unit was not separated from rocks mapped as Wolfcamp on Permian paleotectonic maps (McKee, Oriel, and others, 1967a, pl. 3); its presence is indicated in maps of the present publication, but no specific thicknesses are assigned for interval E (pl. 9A).

An upper boundary for interval E has been placed approximately at the systemic boundary in the easternmost Val Verde basin and on the southwestern slope of the Bend arch. Rocks above and below the boundary are most unfossiliferous mudstone and sandstone; however, the sequence contains a few tongues of fusulinid-bearing limestone in eastern Sutton and Schleicher Counties.

In the southern part of the Delaware basin, the upper boundary for interval E is arbitrarily placed at the base of a thick unfossiliferous mudstone, sandstone, and conglomerate sequence generally considered to be Permian in age. The exact position of the top of Pennsylvanian rocks in this area is much in doubt.

In the Marathon area, the top of the interval is the top of the Gaptank Formation; and in westernmost trans-Pecos Texas, the upper boundary is the top of the Magdalena Limestone. The top of these formations,

| Table 26. — Some named stratigraphic units in interval E, north-central Texas |
|-----------------------------------------------|-----------------------------------------------|-----------------------------------------------|-----------------------------------------------|
| **Group**                                      | **Colorado River drainage**                   | **Brown River drainage**                      | **North Texas (subsurface)**                  |
| **Formation**                                  | **Member**                                    | **Formation**                                 | **Stratigraphic Unit**                        |
| **Chaffin Limestone**                          | **Crystal Falls (or Chaffin) Limestone**       | **Quinn Clay equivalent**                     | **Megargel limestone**                        |
| **Thrifty**                                    | **Parks Mountain Sandstone**                   | **Thrift**                                    | **Ivan Limestone**                            |
| **Brekenridge Limestone**                      | **Brekenridge Limestone**                      | **Blach Ranch (or Speck Mountain) Limestone** | **Avis Sandstone**                            |
| **Speck Mountain Limestone**                   | **Ivan Limestone**                             | **Gunsight Limestone**                        | **Wayland Shale**                             |
| **Cisco**                                      | **Wayland Shale**                              | **Gunsight Limestone**                        | **Gunsight Limestone Member**                 |
| **Graham**                                     | **Graham**                                    | **Necessity Shale**                           | **Bunger Limestone**                          |
| **Bluff Creek Shale (includes several discontinuous limestone beds)** | **North Leon Limestone**                       | **Noble limestone**                           | **Gonzales Limestone**                        |
|                                               |                                               |                                               | **Salem School Limestone**                    |
|                                               |                                               |                                               | **Salem School Limestone Member**             |

*Formation boundaries from Myers, 1965.
however, may not coincide with the time boundary between the Pennsylvanian and the Permian. The fusulindid *Pseudoschwagerina*, which is considered to be an indicator of Early Permian (Wolfcamp) age, has been reported from rocks mapped as Gaptank Formation (West Texas Geol. Soc., 1952, p. 26); fossils present in the uppermost Magdalena cannot definitely be assigned either to Pennsylvanian or to Permian (Harbour, 1972).

**THICKNESS TRENDS**

Interval E has a maximum thickness of about 1,200 feet in Wilbarger County near the Texas-Oklahoma border (pl. 9A). The interval thins gradually from this region southward across the Bend arch. It is truncated abruptly on the east because of Cretaceous and more recent erosion. A north-trending zone of conspicuous westward thinning marks a transition from the Bend arch to the Midland basin; in Mitchell and Sterling Counties at the east edge of the basin, the interval is assigned a thickness of 50-150 feet. Rocks of Virgil age assigned to interval E range in thickness from zero to about 500 feet on the Horseshoe atoll in Kent and Scurry Counties, according to Stafford (1959, p. 8). Elsewhere in the Midland basin, thickness of interval E ranges from less than 100 to a few hundred feet; at a few places near the west edge of the basin, rocks of Virgil age seem to be absent. Interval E is absent from structurally high areas within and marginal to the Central Basin platform and on part of the Pecos arch, owing partly to nondeposition and partly to pre-Permian or Early Permian erosion.

Within the Val Verde and southern Delaware basins, rocks of Virgil age probably are less than 500 feet thick. They may be missing in part of the Delaware basin (pl. 9A). The estimated thickness of rocks of Virgil age in the Marathon area and on the south side of the western Val Verde basin is between 500 and 1,000 feet.

In Presidio County and in northwestern Culberson County, the estimated thickness of rocks of Virgil age is less than 500 feet. In El Paso and Hudspeth Counties, the interval may be somewhat thicker than 500 feet.

**LITHOFACIES TRENDS**

Interval E is nearly 100 percent limestone on the Horseshoe atoll and ranges from more than 50 to nearly 100 percent limestone on much of the Central Basin platform. Limestone comprises 20-50 percent of the interval in a very irregular band along the western and southern parts of the Bend arch (pl. 9B). It makes up about 25 percent of interval E in the Marathon area, and it exceeds in amount both mudstone and sandstone in parts of westernmost trans-Pecos Texas. A mudstone or a sandstone and mudstone facies containing less than 20 percent limestone lies adjacent to areas of high-carbonate rock in the Midland, southern Delaware, and Val Verde basins, and similar facies are present across wide areas on the eastern and northern parts of the Bend arch. The proportion of sandstone increases southward in Edwards County, in the Marathon area, and in Presidio County, and it increases northeastward on the Bend arch. Conglomerate and thin coal beds occur in the northeastern part of the region, and red beds occur along the east side in a wide band that broadens northeastward (fig. 55). Gypsum is associated with mudstone and limestone in El Paso County.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

Sources for interval E in central Texas were land areas in the Wichita-Arbuttle region of southern Oklahoma (Tomlinson and McBee, 1959, p. 46, 47) and in the region of the Ouachita structural belt to the east. At least moderate relief in the source areas is suggested by chert-pebble conglomerates at several horizons in interval E in the northeastern part of the region.

The presence of land areas in the Ouachita belt along the southern margin of central and west Texas is implied by a southward increase of sandstone in and near Edwards County and by probable shallow-water deposits in the upper part of the Gaptank Formation in the Marathon area.

A shallow-water marine environment probably prevailed across most of the Bend arch area during deposition of interval E. Limestone reefs and banks formed offshore along the west edge, whereas mudstone, sandstone in channels, and thin coal beds, prevalent to the east and northeast, are believed to have been deposited in shallow lagoons, in estuaries, and on delta plains. Alternating beds of mudstone, sandstone, and limestone indicate a fairly rapid lateral shifting of the environments of deposition, brought about partly by oscillation of the shoreline. A cyclic sequence of mudstone, channel-fill sandstone, and limestone is repeated five times in the Graham Formation in Stephens and Eastland Counties (Myers, 1965, p. C14). Nine cycles occur between the Bunker Limestone and Wayland Shale Members of the Graham Formation (Lee and others, 1938, p. 25-45); during each cycle, erosion and locally deep channeling were followed by deposition of channel sandstone, fossiliferous marine mudstone, and limestone. Red detritus that interferes from the east with dark-gray marine mudstone suggests fairly rapid deposition on the shelf and the erosion of oxidized soils at the source.

A marked west-facing slope, well established by the end of Des Moines time, that separated shallow-water environments on the shelf from a much deeper water environment in the Midland basin, persisted into and perhaps through Virgil time (Adams and others, 1951; Van Siclen, 1958). The presence of dark-gray and black mudstone and absence of limestone that characterize in-
Figure 55. — Distribution of red beds, conglomerate, and coal in interval E, north-central Texas.
interval E in most of the Midland basin suggest stagnant bottom conditions unsuited for the growth and preservation of lime-secreting organisms. Maximum depth of water over the subsiding basin floor in Late Pennsylvanian and Early Permian time may have been 1,500-2000 feet (Burnside, 1959, p. 31; Adams, 1962, p. 382). The upper parts of the Horseshoe atoll, which grew throughout Virgil time with only temporary cessations (Stafford, 1959), stood above surrounding parts of the basin, at or near the surface of the sea. Reef-building organisms continued to flourish in this better aerated environment.

Aerated shallow-water environments that favored the growth of reef-building organisms prevailed adjacent to and on much of the Central Basin platform and probably much of the Diablo platform. Stagnant bottom conditions such as characterized the Midland basin were duplicated in the deepwater areas of the Val Verde and southern Delaware basins.

In the Franklin Mountains, El Paso County, the presence of gypsum associated with mudstone and limestone of interval E (Harbour, 1972) suggests some restriction of the open-shelf sea of earlier Pennsylvanian time in westernmost Texas. Evaporite deposits also occur in interval E to the north in New Mexico.

PALEOTECTONIC IMPLICATIONS

Much of north-central and west Texas sank slowly during deposition of interval E. Sedimentation kept about even with downwarping on the shelves but lagged behind in the Midland and Val Verde basins and in the southern part of the Delaware basin. The former Red River arch, which had ceased to exist as a positive element by the end of Missouri time, was the site of maximum downwarping north-central Texas during deposition of interval E.

Intermittent uplift occurred in the Ouachita structural belt, which probably remained above sea level throughout Virgil time. The Central Basin and Diablo platforms were maintained as positive elements which were wholly, or in large part, below sea level. They received sediment during much of Virgil time; but in the transition from latest Pennsylvanian to earliest Permian time, they were beveled by erosion that in places cut as deep as Precambrian rocks. Faulting within and marginal to these positive elements, well defined by Permian time, may have begun late in Pennsylvanian time.

In the Marathon area, Virgil time was one of relative calm between stronger tectonic pulses during Missouri time and in the earliest Permian.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

THICKNESS TRENDS

Pennsylvanian rocks have a maximum thickness of about 13,000 feet in the Marietta syncline and about 6,700 feet in the northern part of the Fort Worth basin (pl. 11). In these areas and elsewhere in the eastern part of the region, the upper part of the system is truncated beneath unconformably overlying rocks. In the Hardeman basin, where Pennsylvanian rocks apparently retain their originally deposited thickness, the maximum is about 4,400 feet.

A broadly curving arc, concave southwestward, in which Pennsylvanian rocks are generally thin, marks the position of the Red River and Muenster arches along the north side of the Fort Worth basin. Pennsylvanian rocks are thin or missing, also, on the crest of the Central Basin platform and on the Pecos arch.

The east side of the Midland basin is sharply outlined by a marked westward thinning of Middle and Upper Pennsylvanian rocks in Tom Greene, Coke, and other counties farther north; within the basin, the system is generally less than 500 feet thick on the east and increases by a few hundred feet toward the west margin. Middle and Upper Pennsylvanian carbonate banks and reefs in the Horseshoe atoll and along the east margin of the Central Basin platform locally increase the total thickness of the system to more than 2,000 feet. A narrow zone of thick Pennsylvanian rocks marks a line of reefs that follows the west side of the Central Basin platform and crosses its southern end.

In the Kerr basin, Pennsylvanian rocks may be 7,000-8,000 feet thick. To the west, in the Val Verde and southern Delaware basins, the probable total thickness is on the order of 1,500-2,000 feet. In the Marathon area, maximum thickness of the preserved Pennsylvanian rocks is at least twice as great as in the Val Verde basin. In these areas, neither the upper nor the lower boundary of the Pennsylvania is precisely located, and thicknesses (pl. 11), therefore, are approximate.

Pennsylvanian rocks in westernmost trans-Pecos Texas may be 3,000 feet or more in maximum thickness.

PALEOTECTONIC IMPLICATIONS

The total thickness of Pennsylvanian rocks now present in central and west Texas resulted from a gradual shift of the loci of deposition during Pennsylvania time from geosynclinal areas on the eastern and southern margins of the region to a system of cratonic basins and platforms in the interior of the region. Areas that were tectonically most negative early in Pennsylvania time — the Ouachita geosyncline and adjoining basins — were the sites of pronounced uplift and deep erosion in Late Pennsylvanian time, and uplift and erosion continued in peripheral parts of the region into the Permian and later.

As one consequence of the changing tectonic pattern, rates of sedimentation varied widely in some areas for different parts of the system. Lower Pennsylvanian rocks are thin or absent on the Bend arch, whereas later Penn-
Pennsylvanian rocks are thick there. Conversely, the thickest deposits of Early Pennsylvanian age occur in the eastern Fort Worth basin and the adjacent Ouachita belt, and in these places later Pennsylvanian rocks are thin or missing. A similar interchange in sites of maximum and minimum deposition took place in the Marathon area, where downwarping and uplift migrated across a much narrower zone; probably the same changes occurred elsewhere along the southern margin of the region.

In contrast, in the Midland basin, rates of detrital sedimentation remained slow throughout Pennsylvanian time despite pronounced downwarping during the later part of the period. This basin presumably existed as a submarine depression during later Pennsylvanian time, cut off from sources of sediment by broad surrounding regions of shallow water (Adams and others, 1951). Within the northern part of the Midland basin, an arcuate band of thicker rocks marks the position of the Horseshoe atoll. This feature resulted from an extraordinary production of biostomal limestone in a favorable sedimentary environment; the thickening has no tectonic significance except its value as a measure of subsidence of the basin floor. Irregular thickening in reefs or banks along the west edge of the basin is aligned along the hinge zone between the basin and the relatively stable Central Basin platform.

The structural and depositional history of the Val Verde and southern Delaware basins was similar to that of the Midland basin, but downwarping was greater in the early stages of development. When calcium carbonate of early Des Moines age was deposited in shallow waters of the slowly subsiding Midland and easternmost Val Verde basins, depths of the southern Delaware and western Val Verde basins had already exceeded the limits for precipitation and preservation of calcium carbonate, and these areas may have been starved deep-water basins from earliest Pennsylvanian time (Adams, 1965, p. 2144).

In Texas west of the Delaware basin, the Pennsylvanian System was dominated by two poorly delineated structural elements. The southern tip of the Pedernal uplift, principally developed as the western flank of the Delaware and earlier Tobosa basins in New Mexico, probably stood above sea level in trans-Pecos Texas in Early Pennsylvanian time but was submerged as the period advanced. The Diablo platform extended east-southeast from the approximate south end of the Pedernal uplift and formed the southwestern flank of the Delaware and Val Verde basins. Intermittent positive tendencies along the trend of the Diablo platform, from Precambrian time onward, culminated in Late Pennsylvanian folding, faulting, and erosion which is most clearly demonstrable in the Sierra Diablo region (King, 1965). Little is known of the boundaries of the platform in Pennsylvanian time. Previously published maps (Adams, 1962) and the present interpretation (pl. 11), based on very limited information for this area, suggest a possible structural sag near the east end of the platform which may have been a part of the Marfa basin of Pennsylvanian and Permian time (fig. 45).

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSylvanian SYSTEM**

**UNITS OVERLYING PENNSylvanian SYSTEM**

Permian rocks of Wolfcamp age overlie Pennsylvanian rocks in the greater part of central and west Texas (pl. 12). In the subsurface in much of this region, the basal Permian rocks commonly are referred to simply as rocks of Wolfcamp age.

The Waldrip Shale Member of the Pueblo Formation is the basal Permian unit on the Bend arch north of the Colorado River area. In northern trans-Pecos Texas, the Hueco Limestone is the lowermost Permian formation. The Permian Wolfcamp Formation overlies the Pennsylvanian Gaptank Formation in the northeastern outcrops of the Marathon area; to the southwest, if the base of the Permian is defined by the straigraphically lowest occurrence of the fusulinid *Pseudoschwagerina*, the contact with the Pennsylvanian in part of the area may be within rocks mapped as Gaptank Formation.

The contact between Pennsylvanian and Permian rocks is generally gradational in the basins and on parts of the shelves and platforms. Locally on the Bend arch, channel sandstone of earliest Permian age cuts down into Pennsylvanian rocks. On and near the Central Basin platform and on parts of the Diablo platform, rocks of Wolfcamp age overlap Pennsylvanian rocks of Virgil to Atoka age; at a few places on the Central Basin platform, Permian rocks of Leonard age are directly above Pennsylvanian rocks.

Fluvialite and shallow-water marine deposits of the Trinity Group of Early Cretaceous age unconformably overlie Pennsylvanian rocks in most of the eastern part of the region. In small patches northwest of the Llano area, rocks of Fredericksburg age extend beyond rocks of Trinity age and rest directly on the Pennsylvanian. The surface of unconformity — the Wichita paleoplain of Hill (1901, p. 363-367) — was a rolling surface of low relief that now has an average southeast slope in the Fort Worth basin of about 50 feet per mile (P. T. Flawn, in Flawn and others, 1961, p. 19). Cretaceous rocks also rest on Pennsylvanian rocks in the Marathon area of late Paleozoic folding beyond the limit of Permian deposition.

Mudstone and sandstone of the Schuler Formation of Late Jurassic age may be present unconformably above Pennsylvanian rocks in the central part of Collin County.
sources of the Triassic rocks are believed to have been elevated areas in southern Colorado, southern New Mexico, and southern west Texas (McKee and others, 1959, p. 24, pl. 9).

In Jurassic time the Gulf Coast region sagged downward and sediments were deposited in southern and eastern Texas in a northward-advancing sea (R. W. Imlay, in McKee and others, 1956, pl. 8). The Upper Jurassic Malone Formation in southern Hudspeth County was deposited near the north limit of a central Mexican seaway (Albritton and Smith, 1965, p. 37). Other parts of Texas stood above sea level throughout the Jurassic Period.

Uplift, partial withdrawal of the sea, and planation in latest Jurassic and earliest Cretaceous time were followed by renewed downwarping and deposition across the entire region of Lower Cretaceous deltaic, fluvial, and shallow-water marine rocks. Marine submergence continued in Late Cretaceous time.

Cenozoic regional uplift caused retreat of the sea toward the Gulf of Mexico, and subsequent erosion has resulted in the present-day exposure of Pennsylvanian rocks on the Bend arch and west side of the Fort Worth basin, around the margins of the Llano uplift, and in the Marathon area. Faulting in the basin ranges of trans-Pecos Texas exposed Pennsylvanian rocks in areas where they otherwise might have remained deeply buried beneath younger rocks. Tertiary intrusive and extrusive rocks further modified and in part concealed the older rocks of southern and westernmost west Texas.

REFERENCES

Abilene Geological Society, 1949, Cross sections, Texas, Pennsylvanian — Coke County to Hamilton County; Stonewall County to Hood County; McCulloch County to Young County; Scurry County to Parker County, in Abilene Geol. Soc. Guidebook, Field trip to type localities. Canyon-Strain Series, Pennsylvanian System, 1949: 4 sheets.

1950, Cross sections, Texas, Schleicher County to Brown County; Coke County to Lampasas County, in Abilene Geol. Soc. Guidebook, 1950.

Adams, J. E., 1954, Mid-Paleozoic paleogeography of central Texas, in San Angelo Geol. Soc. Guidebook, Cambrian field trip — Llano area, 1954: p. 70-73. [Also, Shale Shaker, v. 4, no. 6, p. 4-5, 8,9.]


---1957b, Subsurface geology of Cooke County, Texas, in The geology and geophysics of Cooke and Grayson Counties, Texas: Dallas Geol. Soc., p. 75-98.


Lee, Wallace, Nickell, C. O., Williams, J. S., and Henbest, L. G., 1938, Stratigraphic and paleontologic studies of the Pennsylvanian and...


North Texas Geological Society, 1954, West-east cross section, King County to Grayson County, Texas: West-east cross section, Stonewall County to Fannin County, Texas; North-south cross section, Coton County, Oklahoma, to Young County, Texas: Wichita Falls, Tex., North Texas Geol. Soc., 3 sheets.


Turner, G. L., 1957a, Paleozoic stratigraphy of the Fort Worth basin [Texas], in Abilene and Fort Worth Geol. Soc. Joint Field Trip


Waterschoot van der Gracht, W. A. J. M. van, 1931, Permo-


New Mexico

By GEORGE O. BACHMAN

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-L
CONTENTS

Abstract ........................................ 233
Region defined .................................. 233
Paleogeology ...................................... 233
   Units underlying Pennsylvanian ............ 233
   Lower boundary of Pennsylvanian and paleotectonic implications .... 233
Interval A ........................................ 234
   Formations included ......................... 234
   Upper boundary of interval A .............. 234
   Stratigraphic relations .................... 234
   Thickness trends ................................ 234
   Lithofacies trends ........................... 234
   Sources and environments of deposition .... 234
   Paleotectonic implications ............... 235
Interval B ........................................ 235
   Formations included ......................... 235
   Upper boundary of interval B .............. 235
   Thickness trends ................................ 235
   Lithofacies trends ........................... 235
   Sources and environments of deposition .... 235
   Paleotectonic implications ............... 236
Interval C ........................................ 236
   Formations included ......................... 236
   Upper boundary of interval C .............. 236
   Thickness trends ................................ 236
   Lithofacies trends ........................... 236

<table>
<thead>
<tr>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>233</td>
</tr>
<tr>
<td>233</td>
</tr>
<tr>
<td>233</td>
</tr>
<tr>
<td>233</td>
</tr>
<tr>
<td>234</td>
</tr>
<tr>
<td>234</td>
</tr>
<tr>
<td>234</td>
</tr>
<tr>
<td>234</td>
</tr>
<tr>
<td>235</td>
</tr>
<tr>
<td>235</td>
</tr>
<tr>
<td>236</td>
</tr>
<tr>
<td>236</td>
</tr>
<tr>
<td>236</td>
</tr>
<tr>
<td>236</td>
</tr>
<tr>
<td>237</td>
</tr>
<tr>
<td>237</td>
</tr>
<tr>
<td>237</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Interval C — Continued</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sources and environments of deposition</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Intervals D and E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Formations included</td>
</tr>
<tr>
<td>Upper boundary of interval D</td>
</tr>
<tr>
<td>Thickness trends</td>
</tr>
<tr>
<td>Lithofacies trends</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Interval D</th>
</tr>
</thead>
<tbody>
<tr>
<td>Formations included</td>
</tr>
<tr>
<td>Upper boundary of interval E</td>
</tr>
<tr>
<td>Thickness trends</td>
</tr>
<tr>
<td>Lithofacies trends</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Interval E</th>
</tr>
</thead>
<tbody>
<tr>
<td>Formations included</td>
</tr>
<tr>
<td>Upper boundary of interval E</td>
</tr>
<tr>
<td>Thickness trends</td>
</tr>
<tr>
<td>Lithofacies trends</td>
</tr>
<tr>
<td>Sources and environments of deposition</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
<tr>
<td>Total thickness of Pennsylvanian rocks</td>
</tr>
<tr>
<td>Thickness trends</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
<tr>
<td>Geologic units directly above Pennsylvanian System</td>
</tr>
<tr>
<td>Units overlying Pennsylvanian</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
</tr>
<tr>
<td>References</td>
</tr>
</tbody>
</table>

ILLUSTRATIONS

[For listing of plates (in separate case) see volume “Contents”]

<table>
<thead>
<tr>
<th>FIGURE</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>56</td>
<td>Map showing geographic features in New Mexico mentioned in text</td>
</tr>
<tr>
<td>57</td>
<td>Map showing major tectonic elements in New Mexico during Pennsylvanian and Early Permian time</td>
</tr>
<tr>
<td>58</td>
<td>Diagram of stratigraphic relations of Pennsylvanian rocks in the Sacramento Mountains, illustrating typically abrupt facies changes as shown in the Gobbler Formation</td>
</tr>
</tbody>
</table>

TABLE

<table>
<thead>
<tr>
<th>TABLE</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>27</td>
<td>Divisions of the Pennsylvanian System in central and southern New Mexico as established by M. L. Thompson (1942, p. 27)</td>
</tr>
</tbody>
</table>
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES. PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

NEW MEXICO

By GEORGE O. BACHMAN

ABSTRACT

Rocks of Pennsylvanian age, which were originally deposited over most of New Mexico, are now widespread in the State. Their present distribution has been determined in part by post-Pennsylvanian erosion. Pennsylvanian rocks are exposed in most of the mountain ranges throughout central New Mexico, in some outlying ranges, and in the subsurface in other areas of the State.

The rocks of Pennsylvanian age have a complex nomenclature in New Mexico because facies changes make correlations difficult and workers in different areas have devised nomenclature to fit local needs in mapping. Pennsylvanian rocks consist of equivalents to the Morrow, Atoka, Des Moines, Missouri, and Virgil Series of the midcontinent region, although not all these series may be represented at any single locality. These rocks range in thickness from zero to at least 10,000 feet. Lithologies are conglomerate, arkose, sandstone, shale, limestone, and some gypsum.

Pennsylvanian time was characterized by broad shelf areas and generally elongate depositional basins that received sediments from adjacent highlands. The major active tectonic features trended northerly, but numerous minor features had local trends in other directions. During Middle Pennsylvanian time the environment was dominantly marine, and the rocks deposited were mainly limestone and shale. During Late Pennsylvanian time basins were filled, and facies changes were more abrupt; sandstone, arkose, and conglomerates were dominant rock types near highlands, and these intertongued with marine limestone and shale in the basin areas.

The contact of the Pennsylvanian with underlying rocks is a regional unconformity. The contact with overlying Permian rocks at some places is gradational, at other places is disconformable, and at a few places is an angular unconformity.

REGION DEFINED

The area discussed in this section consists of all of the State of New Mexico except a part in Union County in the northeastern part of the State that is included under the discussion of the Texas Panhandle by G. H. Dixon (chap. J).

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

Rocks that underlie the Pennsylvanian in New Mexico range in age from Precambrian to Mississippian (pl. 2). Over much of central and northern New Mexico, Precambrian schist, gneiss, quartzite, and granite and the associated metamorphic and igneous rocks are directly beneath the Pennsylvanian. In south-central New Mexico belts of Paleozoic rocks that are successively younger toward the south underlie the Pennsylvanian. The Montoya Dolomite of Middle and Late Ordovician age underlies the Pennsylvanian in the northern part of the Caballo Mountains (Kelley and Silver, 1952, p. 256) and in the northern San Andres and southern Oscura Mountains (Bachman, 1968, pl. 2) (fig. 56). Shale and dolomite of Devonian age underlie the Pennsylvanian in the Derry Hills (Thompson, 1942, p. 33), in part of the Caballo Mountains (Kelley and Silver, 1952, p. 72), and in the subsurface along the east border of New Mexico. Some of these subsurface dolomitic rocks may be of Silurian age (Lloyd, 1949, p. 53). At many places in northern New Mexico and in most of southern New Mexico, limestones of Mississippian age underlie those of Pennsylvanian age (Baltz and Read, 1960). These Mississippian rocks are generally younger toward the south: Meramec and older Mississippian rocks lie beneath the Pennsylvanian in northern and central New Mexico, whereas rocks of Chester age are present in southern New Mexico (Laudon and Bowsher, 1949, p. 19-20; Armstrong, 1962, p. 24).

LOWER BOUNDARY OF PENNSYLVANIAN AND PALEOTECTONIC IMPLICATIONS

The base of the Pennsylvanian in New Mexico is everywhere marked by an unconformity, below which are rocks of widely different ages (pl. 2). In northern and central New Mexico the unconformity is commonly one of sharp angularity across igneous and metamorphic rocks of Precambrian age. At places where the Pennsylvanian rests on earlier Paleozoic sediments the boundary is less angular. Rocks beneath the unconformity were very broadly folded and were not severely faulted during Paleozoic time, which indicates that tectonism immediately preceding the Pennsylvanian was epeirogenic and did not include mountain building.

In central and northern New Mexico, basal Pennsylvanian strata consist of sandstone and some conglomeratic
interval a

formations included

interval a consists of unnamed shale and limestone of morrow age (pl. 13) that generally are designated "rocks of morrow age." these rocks are preserved in the orogrande, delaware, and pedregosa basins in southern new mexico and possibly in the rowe-mora basin and farther west in northern new mexico (fig. 57), though rocks of the northern areas could not be mapped on plate 3.

in the sacramento mountains, rocks of probable morrow age have been included in the gobbler formation (pray, 1961). in the southern san andres mountains they are part of the lead camp limestone (bachman and myers, 1969). in the sangre de cristo mountains in north-central new mexico, rocks of morrow age have been included in the la pasada and flechado formations (miller and others, 1963, p. 34-45).

numerous classifications of pennsylvanian rocks have been developed in new mexico. most of these classifications, including the one used in this report (pl. 13), are based on local rock facies. attempts have been made also to devise classifications based on paleontology as well as lithology. one such classification is that of thompson (1942), which is based largely on fusulinid zones. thompson's terminology is accepted by many workers (table 27), though it is not used in this publication.

upper boundary of interval a

the contact at the top of interval a is not consistently recognizable on a lithostratigraphic basis. on the basis of fusulinids it is in the zone of millerella and below the lowest occurrence of fusulinella.

stratigraphic relations

many problems in stratigraphic correlation stem from complex facies relations that are results of concurrent development of local mountain uplifts and intermontane basins. individual stratigraphic units are notably discontinuous. in south-central new mexico a sequence composed dominantly of marine limestone that ranges from atoka through missouri in age is correlative with strata composed of limestone, shale, sandstone, and
arkose in central New Mexico. In northern New Mexico rocks of the same age include marine limestone, black shale, arkose, conglomerate, and red beds. Characteristic transitions from continental to marine deposits are exemplified in the present Sacramento Mountains, where in Pennsylvanian time marine carbonate sediments of the Orogrande basin interfingered with terrestrial sediments derived from the Pedernal uplift to the east (fig. 58). In other places in New Mexico such interfingered transitions occur within a fairly short distance and encourages the use of local formation names for local facies.

THICKNESS TRENDS

In the Orogrande basin in south-central New Mexico interval A thickens generally southward from 0 feet in the northern, northwestern, and eastern parts to more than 400 feet in the southern part. In the central part of the Pedregosa basin in southwestern New Mexico interval A attains a thickness of about 300 feet. In the Delaware basin of southeastern New Mexico interval A is locally more than 1,000 feet thick.

On the basis of certain brachiopods, rocks of Morrow age are also believed to be present in the Rowe-Mora basin of north-central New Mexico (Miller and others, 1963). The brachiopod *Schizophrora* cf. *oklahomae*, which occurs in rocks of Morrow age in Oklahoma, has been found in strata in northwestern New Mexico (Northrop and Wood, 1945; Wood and others, 1946), and fusulinids of possible Morrow age have been reported from two wells in northwestern New Mexico (Bass, 1944). Data on thickness are at present inadequate to contour interval A in these areas, however, and in this publication these rocks are included with Atoka rocks in interval B.

LITHOFACIES TRENDS

Interval A in New Mexico consists largely of limestone and shaly limestone (pl. 3B) and is characterized by fine-grained sediments. Some interbeds of calcareous mudstone and thin beds of sandstone are present in this interval in the southern San Andres Mountains (Bachman and Myers, 1963). Black shale of Morrow age is preserved on the Central Basin platform in southeastern New Mexico (Meyer, 1966, p. 31). Argillaceous limestone, mudstone, and argillaceous sandstone make up most of the interval in the Delaware basin east of the Pedernal uplift. Fine-grained detrital sediments trend southeasterly from the Pedernal uplift into the western Delaware basin. Some sandstone occurs in this interval “over essentially all of the Permian [Delaware basin in New Mexico] and Orogrande basins” (Meyer, 1966, p. 34).

SOURCES AND ENVIRONMENTS OF DEPOSITION

Interval A in most of New Mexico lapped onto adjacent land areas as a result of northward transgression of a sea. A tongue of the sea may have invaded the Northwest shelf in the southeastern part of the State (fig. 57) from the east (Meyer, 1966, p. 31). Interval A probably was deposited over much or all of southern New Mexico and was subsequently stripped off the Pedernal
and Central Basin uplifts by later Pennsylvanian and Permian erosion. Sandy beds locally present on both sides of the Pedernal uplift indicate that areas on the crest may have been undergoing some erosion during Morrow time. The distribution of the sandstone also suggests that strandlines along the margins of the uplift during Morrow time may have been near the present limit of Morrow rocks in that area; likewise, the northern part of the zero isopach probably represents a strandline. Fine-grained argillaceous sediments were probably derived from local nearby sources.

Seas during Morrow time in southern New Mexico were relatively shallow and turbid, and land areas were low lying. Some calcium carbonate probably was deposited in clear water, but the widespread occurrence of argillaceous sediments indicates that clear-water conditions were limited in extent.

PALEOTECTONIC IMPLICATIONS

No major orogenic uplift occurred in New Mexico during Morrow time. Some regional warping allowed seas to invade local shallow basins.

INTERVAL B

FORMATIONS INCLUDED

Interval B in New Mexico consists of rocks of Atoka age and, for convenience in mapping, some lithologically similar older and younger rocks that are not set apart by any obvious stratigraphic break (pl. 13). In southeastern and southern New Mexico the interval comprises unnamed rocks of Atoka age in the Delaware basin (fig. 57), the lower part of the Gobbler Formation (Pray, 1961, p. 80) in the Sacramento Mountains (fig. 56), part of the Lead Camp Limestone in the San Andres Mountains, the Red House Formation (Kelley and Silver, 1952, p. 91) in the Caballo Mountains, and rocks equivalent to the Derry Series (Thompson, 1942, p. 26-30) in other areas. In southwestern New Mexico, interval B consists of the lower part of the Horquilla Limestone. In central New Mexico, including the Sandia, Manzano, and Oscura Mountains, it consists of the Sandia Formation. Most of the Sandia Formation in the Rowe-Mora basin in north-central New Mexico is Atoka in age, but part may be of Morrow age. In northwestern New Mexico in the subsurface, interval B consists of the Molas Formation only. Atoka time may not be completely represented by any of these rock units.

UPPER BOUNDARY OF INTERVAL B

Rocks assigned to interval B are transitional and gradational with overlying strata assigned to interval C at all places observed in New Mexico. Throughout much of central New Mexico the top of interval B has been placed at the change from dominantly terrigenous detrital rocks below to dominantly carbonate rocks above. Locally, in the Oscura and San Andres Mountains, these two lithologies interfinger.

THICKNESS TRENDS

Interval B is thickest in north-central New Mexico in the Rowe-Mora basin, where it exceeds 4,000 feet (pl. 64). The interval is generally less than 500 feet thick in a belt that extends from northwestern New Mexico southward into the Orogrande basin of south-central New Mexico, but it exceeds 600 feet within this belt in the Lucero basin. In southeastern New Mexico it is more than 500 feet thick at several places east of the Pedernal landmass and west of the Central Basin platform. It is absent from east-central New Mexico where the Pedernal landmass may have connected with the Roosevelt uplift (fig. 57; Milnesand dome of Adams, 1962; part of Matador arch of others) during this time. In southwestern New Mexico, interval B in the Pedregosa basin is more than 700 feet thick.

LITHOFACIES TRENDS

In most of New Mexico, interval B is composed largely of detrital rocks that range from sandstone through sandy mudstone to mudstone (pl. 6B). In the Sandia and Manzano Mountains pebble to cobble conglomerate derived from the underlying Precambrian is commonly found near the base of the interval; in the San Andres Mountains chert cobbles derived from underlying Mississippian rocks are common at the base of the interval (Bachman and Myers, 1969, p. 26). Conglomeratic sandstone is present in the interval in the Sangre de Cristo Mountains (Miller and others, 1963, p. 30). Shaly limestone and limestone are dominant in the Orogrande basin. Shaly mudstone is present in the central part of the Rowe-Mora basin. In east-central New Mexico and in the Delaware basin, shaly limestone is a major constituent.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Over much of New Mexico, interval B was deposited in a transgressive sea partly on a peneplain that had developed in pre-Pennsylvanian or earliest Pennsylvanian time. Where interval B directly overlies pre-Pennsylvanian rocks, it consists of reworked sediments derived by weathering of the underlying rocks. Many of the thin interbeds of limestone are sandy, silty, or bioclastic, which suggests that periodically the seas were turbulent. Locally, the environment was swampy, and coal beds formed. Sources of sediments were local, and little evidence of extensive deltaic deposits is available to indicate that major stream systems flowed from the highlands.

In north-central New Mexico the Rowe-Mora basin began to develop during Atoka time. Downwarping con-
continued through much of Pennsylvanian time. Mud and silt were deposited during the early stages of development, presumably during times of poor circulation. Bordering land areas were not high above sea level, and coarse sediments were not produced. Ostracodes collected from shaly mudstone beds in the central part of the basin near Holman, N. Mex., are not indicative of age but do indicate a marine and possibly lagoonal environment (I. G. Sohn, U.S. Geol. Survey, written commun., July 18, 1962). Coal along the west edge of the basin near Pecos, N. Mex., indicates a swampy coastal lagoonal environment in that area.

In the basins in southern New Mexico, lime muds were deposited in shallow open seas. These seas were precursors of widespread seas that were characteristic of the time of deposition of the interval C. These seas were warm and generally clear and contained a normal marine fauna of corals, brachiopods, crinoids, and fusulinids. In the Delaware basin, detrital sediments suggest that streams flowed westward from parts of the Central Basin platform.

PALEOTECTONIC IMPLICATIONS
Regional downwarping allowed the sea to transgress across areas of low relief in much of New Mexico during the time of interval B; there is no evidence of major orogenic activity at that time. The Rowe-Mora basin began to develop as a major depression, but warping in that area was gentle until later in the Pennsylvanian Period.

INTERVAL C
FORMATIONS INCLUDED
The Paradox Member at the base of the Hermosa Formation and, in areas where the age has been determined by fusulinids, a lower part of the Hermosa Formation in the subsurface of northwestern New Mexico are included in interval C (pl. 13). Also included are the lower gray limestone member of the Madera Formation in central New Mexico, the Nakaye Formation (Kelley and Silver, 1952, p. 92) in the Caballo Mountains, the upper part of the Gobbler Formation (Pray, 1961, p. 80) in the Sacramento Mountains, rocks of Des Moines or Strawn age in the subsurface of southeastern New Mexico, and that part of the Horquilla Limestone which is Des Moines in age in southwestern New Mexico. This interval also includes unnamed rocks of Des Moines age in many places in New Mexico.

UPPER BOUNDARY OF INTERVAL C
The upper boundary of interval C is the top of the stratigraphic range of the fusulinid genus *Fusulina* where this can be determined. This fossil is not present at all localities, and commonly the upper boundary must be drawn on a tenuous lithologic correlation. In central New Mexico it is the top of a dominantly limestone sequence and is stratigraphically below a sequence of interbedded arkose, sandstone, mudstone, and limestone. This boundary is gradational, and rocks characteristic of interval C grade into and interfinger with strata that are typical of overlying interval D.

THICKNESS TRENDS
Owing to uncertainties of correlation and to local variations in thickness, 500-foot isopachs for interval C in New Mexico have been drawn in most of the State, but 100-foot isopachs are shown locally in the west (pl. 7A). These isopachs indicate the general magnitude of thickness and regional trends throughout the State. In northwestern New Mexico the interval is more than 1,000 feet thick adjacent to the Paradox basin of southwestern Colorado; it thins from this area southward to a wedge edge along the Zuni uplift. Shelf areas and sedimentary basins are well separated elsewhere in New Mexico by the 500-foot isopach. Thus, the Lucero, Orogrande, Delaware, and Pedregosa basins are delineated by the 500-foot isopach. Locally, in the Lucero and Delaware basins, thickness of interval C exceeds 1,000 feet. In the Pedregosa basin, interval C is over 900 feet thick. It is less than 500 feet thick over most of the rest of the State. The apparent thickening of rocks of this interval in sedimentary basins may be the result, in part, of erosion of strata from highlands during later Pennsylvanian and Permian uplift. Actual deposition of these rocks was probably more widespread than present occurrence indicates.

LITHOFACIES TRENDS
Throughout most of New Mexico interval C is dominantly gray limestone (pl. 7B). Some of the limestone is argillaceous near the Zuni uplift, in parts of the Delaware basin, and at the south end of the Rowe-Mora basin. Dark-gray nodular chert is conspicuous in many limestone beds of this interval. Calcareous mudstone is a major part of the interval in the Rowe-Mora basin and along the southwest edge of the Uncompahgre uplift, where some arkose is also present. A small area of sandy mudstone is on the northeast edge of the Pedernal uplift, and conglomeratic sandstone and siltstone are present in the northwestern part of the Rowe-Mora basin and in the bordering Sangre de Cristo Mountains. In the Sangre de Cristo Mountains an abrupt facies change occurs from a dominantly carbonate sequence in the southern part of the mountains to a thicker sequence composed of coarse detrital rock in the northern part (Miller and others, 1963, p. 33).

SOURCES AND ENVIRONMENTS OF DEPOSITION
Seas were widespread in New Mexico during the time represented by interval C. Because deposits of the inter-
val have been partly removed by subsequent erosion, the seas may once have been much more widespread across the State than can now be determined. The dominance of limestone indicates that over much of the area the seas were clear and that any source areas for terrigenous sediment either were far away or had very low topographic relief. Corals, fusulinids, and brachiopods are well represented in the rocks of this interval and indicate that marine life was abundant and that the seas were shallow and relatively warm.

**PALEOTECTONIC IMPLICATIONS**

Significant vertical movement along the southern part of the Uncompahgre uplift during interval C time is indicated by conglomeratic sandstone and siltstone in the western part of the Rowe-Mora basin. This is the earliest indication of the building of the Ancestral Rocky Mountains in New Mexico during Pennsylvanian time.

Thinning of interval C toward many of the Pennsylvanian uplifts may have been due partly to local warping of low amplitude. The absence of rocks of this interval from many of the uplifts resulted from post-Pennsylvanian erosion of deposits that were at one time more extensive, as shown by the fact that carbonate rocks typical of offshore deposits adjoin many areas of uplift.

**INTERVALS D AND E**

Intervals D and E include the sediments of Pennsylvanian age deposited during the time span of the fusulinid *Triticoptes*. They represent the Missouri and Virgil Series of the midcontinent region. In New Mexico, lithology of these two intervals is similar, and throughout wide areas it is not practical to separate them, owing to the absence of faunal control; the problem is especially difficult in the northern part of the State. Rocks of interval D are widespread and probably are present at all places in which post-Des Moines Pennsylvanian rocks are known. On the accompanying maps, strata of interval D are shown to be present in many areas on the basis of lithologic correlation alone, although specific thicknesses of interval D could not be determined in parts of the region.

Significant Pennsylvanian uplift in New Mexico began in interval C time in the Uncompahgre area. During the time represented by intervals D and E, orogenic activity increased and spread southward. Parts of the Ancestral Rocky Mountains developed in the region of both the Uncompahgre and the Pedernal uplifts. Large amounts of detrital sediments were eroded from these uplifts and carried into adjacent basins. Precambrian rocks that were exposed in parts of central New Mexico provided feldspars and associated minerals that accumulated as arkosic sediments. In some areas, cobbles and boulders were deposited close to the uplifts. Pulses of uplift have not been closely dated within these intervals.

The Ancestral Rocky Mountains were first mentioned in the literature by Lee (1918, p. 5). Later he (1923, p. 286) stated that "in Late Carboniferous time mountains which have been called the Ancestral Rocky Mountains, comparable in size to the present Rocky Mountains, occupied the site of the present Rockies and sediments were shed from them in all directions." Evidence for existence of these mountains was summarized by Melton (1925), Ver Wiebe (1930), and Heaton (1933). Reference to these uplifts as the "Ancestral Rocky Mountains" is less frequent in more recent literature, though the name is used occasionally as a collective term for all Pennsylvanian uplifts in the southern Rocky Mountain region (Thompson, 1942, p. 11; Wengert and Matheny, 1958, p. 2063). In New Mexico these mountains began to rise in Missouri time and reached maximum development in Virgil and Early Permian time.

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D consists of all the strata in New Mexico known to be equivalent to the Missouri Series in the midcontinent region. It comprises the middle part of the Hermosa Formation in northwestern New Mexico, the lower part of the arkosic limestone member of the Madera Formation in central New Mexico, the lower part of the Sangre de Cristo Formation in parts of north-central New Mexico, the lower part of the Atrasado Member of the Madera Limestone in the Lucero basin, the lower part of the Beeman Formation (Pray, 1961, p. 84) in the Sacramento Mountains, and unnamed rocks of Missouri age in southern and southeastern New Mexico (pl. 13). It includes rocks of Missouri age in the Horquilla Limestone of southwestern New Mexico.

**UPPER BOUNDARY OF INTERVAL D**

Interval D at most places is gradational with overlying interval E, and characteristic facies of the two intervals interfinger laterally. No widespread disconformity has been recognized at the boundary, and only locally do marked changes in lithology occur at or near this boundary in New Mexico. In the Sacramento Mountains the Beeman Formation is mapped as interval D because it is lithologically distinguishable from the overlying rocks of interval E, although it may not represent all of interval D time. The Beeman Formation consists largely of interbedded gray limestone and calcareous shale but also contains thin beds of feldspathic sandstone and red shale. Overlying Pennsylvanian strata of interval E contain a diversity of rock types that are largely terrestrial in origin. Thus, in the Sacramento Mountains, interval D represents a transition between limestone deposited in

**THE PENNSYLVANIAN SYSTEM, PART I**
the regionally uniform offshore environment of interval C and terrestrial detritus deposited in the more diverse nearshore environments of interval E. In parts of the Sacramento Mountains, in the Joyita uplift, and at places in southwestern New Mexico interval E was eroded before deposition of Permian strata, and the upper boundary of interval D is an unconformity.

THICKNESS TRENDS
In the northern half of New Mexico, faunal control is not adequate to permit detailed delineation of thickness trends of interval D (pl. 84). Thickness of rocks of Missouri age that ordinarily would be assigned to the interval probably does not exceed 500 feet at any place in northern New Mexico except on the southern shelf of the Paradox basin. The interval is present in the Rowe-Mora basin, but very little information that would permit determining its thickness is available. Generally, this interval is thicker in the southern part of the State than in the northern part. In southwestern New Mexico in the Pedregosa basin, rocks assigned to interval D are 500 feet thick; in the northern part of the Orogande basin they are over 700 feet thick; and at some places in the Delaware basin they exceed 900 feet in thickness.

LITHOFACIES TRENDS
Throughout much of northern New Mexico interval D consists of interbedded arkose, arkosic sandstone, mudstone, and limestone, probably in more complex relationship than is shown by the limited data available for plate 8B. Although sandstone and other coarse detrital rocks are conspicuous, calcareous mudstone and shaly limestone generally form the major part of the interval. In the vicinity of the Joyita uplift, interval D contains much mudstone. Although not indicated on the lithofacies map, the oldest Pennsylvanian red shales observed in the Orogande basin are in interval D (Pray, 1961, p. 87). Along the east side of the Pedernal uplift shaly sandstone is a major constituent of the interval. In the southern part of the Orogande basin and in the Pedregosa and Delaware basins of southern New Mexico, limestone is the dominant rock type.

SOURCES AND ENVIRONMENTS OF DEPOSITION
The areal extent of seas in New Mexico during the time of interval D was more restricted than during the time of interval C. Most of the area of deposition initially was marine, but the rapid influx of detritus caused continental sediments to encroach on marine areas. Some marine limestones in north-central New Mexico contain fresh grains of feldspar that indicate uplift of nearby land areas followed by rapid transport and deposition. Later in interval D time, seas began to withdraw from the region as a result of both regional uplift and filling of the basins with terrestrial sediments. Local moun-

tainous uplift, continuing in northern New Mexico and extending into the southern part of the State, produced coarse detrital sediment. Basins filled quickly, especially in the northern half of the State. Land areas along the margins of basins were periodically flooded by the sea. Conglomerate, as well as red arkose, was deposited in the southern Rowe-Mora basin while some thin beds of red mud were deposited in the northern part of the Orogande basin, indicating vigorous erosion of nearby land. Seas were shallow, and local bioherms and calcareous banks developed. The marine fauna was similar to that of interval C, but bryozoa may have been more abundant at various times during interval D time.

The Uncompahgre uplift was a major source of sediments in the Rowe-Mora basin, and the Pedernal uplift was the major contributor in basins farther south. On a much smaller scale the Zuni uplift contributed sediments to the shallow sea in northwestern New Mexico.

PALEOTECTONIC IMPLICATIONS
Uplift of the Ancestral Rocky Mountains was accelerated and extended southward in New Mexico during interval D time. Land areas higher than those existing during the time of interval C were reflected in deposition of quartz sand, arkosic sand, gravel, and some nearshore or continental red beds. In particular, "the Beeman Formation is interpreted as indicating an increasing degree of tectonic activity in the nearby area and the relative emergence of an area to the east compared with one to the west" (Pray, 1961, p. 87). The Rowe-Mora, Orogande, and Delaware basins were appreciably downwarped, and other smaller basins in the State also were actively sinking but to a lesser degree.

Missouri time in New Mexico was marked by a change from transgression of seas to extensive mountain building and downwarping of basins. The rate of basin sinking only slightly exceeded the rate of sedimentary filling.

INTERVAL E
FORMATIONS INCLUDED
Interval E consists of all strata in New Mexico that are equivalent to the Virgil Series in the midcontinent region. It includes much of the Sangre de Cristo Formation in north-central New Mexico, the upper part of the arkosic limestone member of the Madera Formation in central New Mexico, and the upper part of the Hermosa Formation in northwestern New Mexico. The upper part of the Atrasado Member and the Red Tanks Member of the Madera Limestone in the Lucero basin are in interval E. The upper part of the Bar B Formation (Kelley and Silver, 1952, p. 93) in the Caballo Mountains and all of the Panther Seep Formation and the Holder Forma-
tion (Pray, 1961, p. 87) in the San Andres and Sacramento Mountains, respectively, are included. The upper part of the Horquilla Limestone and the lower part of the Earp Formation in southwestern New Mexico, as well as unnamed rocks of Virgil age in southeastern New Mexico, are in this interval.

UPPER BOUNDARY OF INTERVAL E

In New Mexico the upper boundary of interval E, which represents the Pennsylvanian-Permian boundary, varies from a marked unconformity to an arbitrarily selected plane within a conformable sequence. In north-central New Mexico the boundary is within a sequence of arkosic conglomerate and sandstone in the Sangre de Cristo Formation. Where the boundary is the top of the arkosic limestone member of the Madera Formation or the Bar B Formation, it is usually easily picked, as these formations are normally overlain by distinctive red beds of the Permian Abo Formation.

The upper contact is not readily apparent in central New Mexico where the arkosic limestone member of the Madera Formation and also the Panther Seep Formation in the northern San Andres Mountains are overlain by the Lower Permian Bursum Formation. The Bursum consists of red Abo-like detrital sediments interbedded with marine limestone. Similar interbedded units that compose the upper part of the Pennsylvanian in parts of central New Mexico are differentiated from the Permian largely on the basis of fossils.

In the Delaware basin the upper boundary apparently is conformable in the center of the basin but is disconformable along the margins.

THICKNESS TRENDS

Although interval E is probably present throughout New Mexico at most places where interval D is present, these intervals are not easily differentiated. For this reason, thickness trends of interval E are not readily apparent over much of the State. In general, there is uniform thickening northward from the Zuni uplift. In the Rowe-Mora basin the interval is thickest in a northerly trend parallel to the axis of the Uncompahgre uplift. In much of central New Mexico the interval shows little apparent trend except for thickening in the Lucero basin and Estancia trough. In the Orogrande basin interval E is more than 2,000 feet thick, and in the Delaware basin it is locally more than 1,000 feet thick.

LITHOFACIES TRENDS

In north-central and central New Mexico interval E is similar lithologically to interval D; however, interval E locally contains a larger percentage of coarse detritus close to uplifts. In the northern Orogrande basin, interval E is dominantly mudstone and sandstone but has some conglomerate in an area west of the Pedernal uplift. In the southern Orogrande basin, dolomitic limestone and at least two beds of gypsum, 10 feet to more than 40 feet thick, are present in the upper part of this interval. In much of southeastern New Mexico, calcareous sediments are dominant.

SOURCES AND ENVIRONMENTS OF DEPOSITION

During interval E time in New Mexico, local orogenic uplifts (fig. 57) were recurrent and provided the sources for most of the sediment; the Uncompahgre and Pedernal uplifts provided the largest portion. Seas were near the uplifts, and much deposition was in shallow open-sea and lagoonal environments. These seas were frequently turbid. Detrital sediments were shed into basins so rapidly in north-central and central New Mexico that fresh grains of feldspar were deposited in calcareous marine environments. Basins were filled and many seaways were closed. Shorelines at this time had a complex history of advance and retreat, and it is probable that late in interval E time small enclosed basins were formed in southern New Mexico and that gypsum was deposited in these basins. In the central part of the Orogrande basin, an area of interfingering marine-shelf and fluviatile deposits developed (Pray, 1961, p. 90; Bachman, 1968, p. 23).

PALEOTECTONIC IMPLICATIONS

In New Mexico, uplift of the Ancestral Rocky Mountains reached a maximum during interval E time. Major faulting occurred along the west side of the Pedernal uplift and the east side of the Uncompahgre uplift and probably also along the borders of uplifts of other areas. The Orogrande and Delaware basins were important negative tectonic elements in the State.

TOTAL THICKNESS OF PENNSylvANIAN ROCKS

THICKNESS TRENDS

The major depositional basins in New Mexico during Pennsylvanian time are readily apparent on the isopach map showing total rocks preserved (pl. 11). These are the Rowe-Mora basin in the north, the Lucero basin, Estancia trough, and Tucumcari basin in central New Mexico, and the Orogrande, Pedregosa, and Delaware basins in the south (fig. 57). During Pennsylvanian time, northwestern New Mexico was a shelf extending southward from the tectonically active Paradox basin to the north in Colorado and Utah.

Pennsylvanian rocks exceed 10,000 feet in thickness in the Rowe-Mora basin, an asymmetrical zeugogeancline (Kay, 1951, p. 24) which complements the elongate Uncompahgre uplift adjacent on the west. Downwarping of the basin persisted through Pennsylvanian into Permian time. The great thickness in this area is due to
deposition of marine lagoonal black muds in Early Pennsylvanian time and to later nonmarine deposition as the basin filled and the sea withdrew from the basin and shelf areas.

In northwestern New Mexico, in contrast to the thick sediments that accumulated in the Rowe-Mora basin, Pennsylvanian sediments northward from the Zuni uplift accumulated to a thickness of only 2,500-3,000 feet along the north boundary of the State. Though detrital debris is common in the Pennsylvanian of this area, it is not as coarse, extensive, or thick as in the Rowe-Mora basin.

The Lucero basin contains over 2,000 feet of marine Pennsylvanian strata that include detrital rocks of terrestrial origin probably derived from the Zuni uplift to the west as well as from the Pedernal uplift to the east. In the Estancia trough Pennsylvanian rocks are over 3,000 feet thick and are largely detrital material of terrestrial origin derived from the nearby Pedernal uplift. The Tucumcari basin in east-central New Mexico contains more than 2,000 feet of Pennsylvanian rocks that consist of detrital arkosic debris and fine-grained interbeds of marine strata.

The northward-trending Orogrande basin in south-central New Mexico contains a large variety of Pennsylvanian sedimentary rocks that attain a thickness of over 3,000 feet. The basin was bounded on the east by the Pedernal uplift and on the north and west by shelf areas. Rocks deposited in this basin were largely marine, and the central part of the area received sediments throughout Pennsylvanian time. Some of the marginal thinning is due to Late Pennsylvanian or Early Permian erosion or to nondeposition. Pennsylvanian faulting along the east side of the basin at the margin of the Pedernal uplift probably influenced both the type and the thickness of sediment deposited.

Pennsylvanian rocks in the Pedregosa basin in southwestern New Mexico attain a thickness of over 2,000 feet. The limited data available suggest that this thickness represents dominantly carbonate deposition, including biohermal masses, throughout Pennsylvanian time. Detrital sediments from a northern source increased in that area during Late Pennsylvanian time but did not notably alter thickness trends in the basin. The Florida Island (Burro Mountain) uplift was a source of some sediments that were shed southward into the Pedregosa basin, but deposition resulting from erosion of that uplift was more noticeable northward on the Robledo shelf.

The Delaware basin (part of the larger Permian basin of New Mexico and west Texas) is an economically important area where petroleum and natural gas are produced from Pennsylvanian rocks. The Pennsylvanian is not exposed in that area in New Mexico and is known only from drill-hole data. The system is more than 2,500 feet thick in some parts of the Delaware basin; but, owing to orogenic activity late in the Pennsylvanian along the Pedernal uplift to the west and on the Central Basin and Roosevelt uplifts to the east, stratigraphic relations are complex. The Central Basin platform probably contributed only minor sediment to the Delaware basin, and during the Early and Middle Pennsylvanian the platform itself may have received a thin deposit of sediments that was later stripped away. Limestone and mudstone are the dominant rock types, though coarser detrital sediments are common throughout the Pennsylvanian sequence. The environment varied from open marine to stagnant lagoonal.

**PALEOTECTONIC IMPLICATIONS**

Major tectonic elements that influenced sedimentation throughout Pennsylvanian time but that were increasingly active in Missouri and Virgil time are apparent on plate 11 and are identified in figure 57. The most active and the largest of the uplifts were the Uncompaghre and Pedernal uplifts that trend north through central New Mexico. Late Pennsylvanian faulting between the Uncompaghre uplift and the Rowe-Mora basin is suggested by the asymmetry of the basin (pl. 11) and by a thick accumulation of coarse conglomeratic rocks near the west edge of the basin. Faulting also occurred at the west side of the Pedernal uplift between the uplift and the Orogrande basin. In the northeast, the Sierra Grande uplift was a major feature. In the west, the Zuni uplift was less spectacular but remained positive throughout most, if not all, of the Pennsylvanian Period.

The Joyita uplift was active near the close of Pennsylvanian time, when some, but not all, Pennsylvanian strata were eroded (Kottlowski and Stewart, 1970). The Florida Island uplift and Central Basin platform were mildly active early in the Pennsylvanian and strongly active late in the Pennsylvanian. Basinal areas were not well defined early in the period, but they became active in late Atoka or Des Moines time and continued to develop during the rest of the Pennsylvanian.

In general, the focus of orogenic activity moved southward in New Mexico through Pennsylvanian time.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

**UNITS OVERLYING PENNSYLVANIAN**

Rocks of Early Permian age overlie the Pennsylvanian System at most places (pl. 12). The contact at places is gradational but at other places is an angular unconformity. In central and southeastern New Mexico, rocks of earliest Permian age rest conformably on the Pennsylvanian. Along the margins of depositional basins the con-
contact between the two systems is commonly disconformable or unconformable, and the surface of unconformity is channeled as a result of stream erosion. In parts of southwestern New Mexico, north of the Florida Island uplift, a hiatus is apparent as a contact of erosional disconformity. In the eastern part of the Orogena basin the contact is locally an angular unconformity. In the central part of the Orogena basin, in the present-day southern Oscura Mountains, the contact is apparently conformable and gradational, and separation of the systems on the basis of contrasting rock types is difficult. In parts of southern New Mexico where the Hueco Limestone rests on the Pennsylvanian, a disconformity, or hiatus, between the two units represents nondeposition.

Units of Early Permian age overlying the Pennsylvanian include the Abo and Bursum Formations in central New Mexico, the Hueco Limestone in southern New Mexico, the Cutler Formation in northwestern New Mexico, and the Wolfcamp Formation in much of eastern New Mexico. In the northeastern and north-central parts of the State, the Pennsylvanian strata of the lower part of the Sangre de Cristo Formation grade upward into the Permian strata of the upper part of the formation.

Rocks of Tertiary and Quaternary ages, consisting chiefly of gravel, rest on the Pennsylvanian in some small areas (not shown on pl. 12) where Tertiary erosion has occurred.

PALEOTECTONIC IMPLICATIONS

In northern and central New Mexico, antclinal or faulted uplifting that had begun during Des Moines and Mississippian time continued into Early Permian time. Indications of uplift are disconformities and angular unconformities that are present along the margins of some depositional basins between rocks of Pennsylvanian and Permian ages.

REFERENCES


Wilpolt, R. H., and Wanek, A. A., 1951, Geology of the region from Socorro and San Antonio east to Chupadera Mesa, Socorro County, New Mexico: U.S. Geol. Survey Oil and Gas Inv. Map OM-121.


Eastern Colorado

By RICHARD F. WILSON

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-M
CONTENTS

Abstract
Region defined
Paleogeology
  Units underlying Pennsylvanian
  Lower boundary of Pennsylvanian
  Paleotectonic implications
Interval A
  Formations included
  Upper boundary of interval A
  Thickness trends
  Lithofacies trends
  Sources and environments of deposition
  Paleotectonic implications
Interval B
  Formations included
  Upper boundary of interval B
  Thickness trends
  Lithofacies trends
  Sources and environments of deposition
  Paleotectonic implications
Interval C — Continued
  Sources and environments of deposition
  Paleotectonic implications
Interval D
  Formations included
  Upper boundary of interval D
  Thickness trends
  Lithofacies trends
  Sources and environments of deposition
  Paleotectonic implications
Interval E
  Formations included
  Upper boundary of interval E
  Thickness trends
  Lithofacies trends
  Sources and environments of deposition
  Paleotectonic implications
  Total thickness of Pennsylvanian rocks
  Thickness trends
  Paleotectonic implications
Geologic units directly above Pennsylvanian System
  Units overlying Pennsylvanian
  Paleotectonic implications
References

ILLUSTRATIONS

For listing of plates (in separate case) see volume “Contents”

Figure 59. Map showing geographic features in eastern Colorado mentioned in text
Figure 60. Map showing Pennsylvanian tectonic features of eastern Colorado mentioned in text
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES.
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

EASTERN COLORADO

By RICHARD F. WILSON

ABSTRACT

During the Pennsylvanian Period, eastern Colorado was the site of both marine and nonmarine deposition controlled by two prominent uplifts — the Frontrange and the Apishapa–Sierra Grande — in the western and southern parts of the region, respectively, and by a seaway that occupied most of the northern, eastern, and southeastern parts of the region. Pennsylvanian rocks reached their maximum thicknesses in several subsidiary troughs and basins within a negative area adjacent to the Apishapa–Sierra Grande uplift and the south end of the Frontrange uplift. In some of these basins, 3,000-4,400 feet of Pennsylvanian strata accumulated. Pennsylvanian rocks thin northward and northeastward from these centers of accumulation to between 600 and 1,000 feet in northeastern Colorado.

In eastern Colorado, Pennsylvanian strata rest upon erosional unconformity on rocks ranging in age from Precambrian to Late Mississippian (Chester). Rocks of Chester age are confined to the southeastern corner of the State. North and west of the area of Chester rocks, Pennsylvanian strata progressively overlap older formations until they rest on Precambrian rocks over broad areas in northern Colorado and across the western part of the Apishapa uplift in southern Colorado.

Lower Pennsylvanian sediments (interval A) were characterized by dominantly detrital deposits, largely mudstone, derived from several sources including Nebraska and northwestern Kansas to the northeast, the northern part of the Frontrange uplift, and the western part of the Apishapa uplift. The deposits are interpreted as largely shallow-water marine and marginal marine, laid down during a period of transgression of a sea that entered the State in the southeast corner and spread northward and westward.

Middle Pennsylvanian sediments (intervals B and C) were derived from the rising Frontrange and Apishapa–Sierra Grande uplifts, which shed arkosic detritus eastward and northward from their flanks in the form of alluvial fans or aprons. These uplifts seem to have been most active during interval C. Pronounced basins developed during interval C, and more than 1,500 feet of detrital strata accumulated in a linear trough adjacent to and parallel with the north flank of the Apishapa uplift. North and east of the uplifts, carbonate deposits were dominant in both euxinic and normal marine waters.

Upper Pennsylvanian sediments (intervals D and E) suggest a decrease in the amount of uplift of the Frontrange and Apishapa–Sierra Grande positive areas and a corresponding decrease in downwarping of the basins. As a result, Upper Pennsylvanian strata are blanketlike deposits that thin gradually north and east. Carbonate deposits were dominant in most places that were distant from the positive areas, whereas detrital deposits bordered the flanks of uplifts. The southern part of the Frontrange uplift seems to have been the principal source of sand-sized detritus.

Lower Permian (Wolfcamp) rocks rest upon Pennsylvanian strata everywhere in eastern Colorado, with possibly some local exceptions in the Wet Mountains where Jurassic seems to rest on Pennsylvanian. In northern Colorado, the contact between the Permian and the Pennsylvanian is an erosional unconformity, beneath which Pennsylvanian rocks are progressively beveled in northward and westward directions. Locally along the Front Range in northernmost Colorado, all of interval E apparently was removed by pre-Permian erosion. In the southern part of eastern Colorado, evidence of a hiatus is not clear; Permian rocks rest on Pennsylvanian with apparent conformity.

REGION DEFINED

Pennsylvanian rocks are present in both outcrops and subcrops east of the crest of the Front Range from the northern border of Colorado to the termination of the range south of Colorado Springs; they are also present east of the Wet Mountains south of Canon City (fig. 59). In addition, rocks of Pennsylvanian age occur north and east of the buried Apishapa–Sierra Grande uplift in southeastern Colorado.

Within parts of this region the stratigraphic boundaries of the Pennsylvanian System are difficult to determine. Most of the Fountain Formation is considered to be Pennsylvanian in outcrop areas along the east flank of the Front Range and in the Canon City embayment. The entire formation is Pennsylvanian north of Dry Creek (loc. 754, pl. 1), but to the south, part of it is believed to be Permian and was mapped with other rocks of that age in the Permian paleotectonic publications (McKee, Oriel, and others, 1967a, b).

In most of the subsurface of eastern Colorado, Pennsylvanian rocks have not been divided into formations or other formal units, although, locally, group names used for intervals C through E in Kansas have been extended with varied success. Mostly, the subsurface rocks are subdivided into units on the basis of their inferred age rather than according to a distinctive lithology.

Throughout this chapter the term “Frontrange uplift” is used to distinguish the north-northwest-trending Pennsylvanian positive element from the present north-trending Front Range.
Figure 59. — Geographic features in eastern Colorado mentioned in text. Numbered localities are those in “Index to Localities and Sources,” part II of this publication.
PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVIANIAN

In eastern Colorado, rocks of Pennsylvanian age overlie with erosional unconformity rocks ranging in age from Precambrian to Late Mississippian (Chester) (pl. 2). Rocks of Chester age, composed of an unnamed sequence of interstratified sandstones, limestones, and red and green mudstones, are directly beneath the Pennsylvanian in the southeastern part of the region. Their distribution coincides with the approximate position of the western part of the Hugoton embayment of the Anadarko basin, a structural feature that extended into southeastern Colorado from Kansas during Pennsylvanian time (fig. 60).

North from the area of Chester rocks in southeastern Colorado, Pennsylvanian rocks progressively overlap older formations and so, in northern Colorado, Pennsylvanian rocks rest upon Precambrian granite. These Precambrian rocks mark the crest of an arch prominent in this region earlier in Paleozoic time but buried by Pennsylvanian sediments.

West of the area of Chester rocks in southeastern Colorado, Pennsylvanian strata overlap older rocks and rest upon Ordovician and Precambrian rocks in Las Animas, Huerfano, and southern Otero and Pueblo Counties (fig. 59). The pre-Pennsylvanian rocks outline the Apishapa-Sierra Grande uplift, a positive element that developed in this region in Pennsylvanian time. Part of the boundary of Ordovician with younger rocks is the Freezeout Creek fault, active during Pennsylvanian time (Buehler, 1947).

Between areas of Precambrian rocks in northern Colorado and of Precambrian and Ordovician rocks in southern Colorado, Mississippian strata nearly everywhere directly underlie the Pennsylvanian. The youngest Mississippian rocks throughout this area are of Meramec age.

Rocks of Meramec age in the subsurface are assigned, in ascending order, to the Warsaw, Salem, St. Louis, and the Ste. Genevieve Limestones. The Ste. Genevieve Limestone, bordering rocks of Chester age in southeastern Colorado and extending northward to the northern border of Kiowa County, is characterized by sandy oolitic limestone. Farther west, at two localities on the east flank of the Wet Mountains and at the southeast edge of the Canon City embayment (locs. 769, 776, pl. 1), the Beulah Limestone, which is lithologically similar to, and considered the equivalent of, the Ste. Genevieve, crops out beneath the Pennsylvanian (Maher, 1958, p. 70). To the north and west of the pinch-out of the Ste. Genevieve Limestone, the St. Louis Limestone, which underlies the Pennsylvanian, is characterized by oolitic limestone, minor dolomite, and chert. Correlatives of the St. Louis to the west are believed to be the Hardscrabble Limestone in the Canon City embayment (loc. 776) and the Madison Limestone, of some authors, in the Colorado Springs area (loc. 767) (Maher, 1958, table 1, figs. 8, 9, p. 70), both of which are of similar lithologic type.

The undifferentiated Warsaw and Salem Limestones, which form the most northerly band of rocks of Meramec age directly under the Pennsylvanian, consist primarily of interstratified crystalline limestone and dolomite containing brown and gray chert. The Williams Canyon Limestone, farther west, crops out beneath the Pennsylvanian in parts of the Canon City embayment (loc. 773) and north of Colorado Springs (loc. 775) and is considered to be a probable correlative of the Salem and Warsaw Limestones (Maher, 1950; 1958, p. 69).

Mississippian rocks of Osage age, assigned to the Burlington and Keokuk Limestones, underlie the Pennsylvanian in some drill-hole localities in a narrow southwest-trending band in Yuma and southern Washington Counties, in a north-trending lobe in western Washington and southeastern Morgan Counties, and in a southeast-trending lobe in northern Weld Coun-
ty. Osage rocks in the southwest-trending band consist of interstratified cherty dolomite and dolomitic limestone and contain some glauconite. Rocks in the north-trending and southeast-trending lobes are composed of limestone, lesser amounts of dolomite, and some chert; but they may not represent the same stratigraphic units as the Osage rocks farther east.

The southeast-trending lobe of Osage rocks in Weld County should extend farther east and south than is shown on the paleogeologic map (pl. 2), according to Maughan (1963, fig. 66.2). In eastern Weld, western Logan, and northeastern Morgan Counties (locs. 373, 208, 214, 751, 114), some places mapped as Precambrian rocks directly underlying Pennsylvanian may actually represent rocks of Osage age. A limestone unit that occurs at such localities is separated from Precambrian granite by a thin sequence of sandstone, conglomerate, and shale. Although this limestone and the underlying detrital rocks may be of Osage age, the foraminifer *Millerella*, which is most characteristic of Morrow age although it has a longer range, has been reported (Am. Strat. Co., unpub. data) from the detrital sequence below the limestone. On this basis these rocks have been assigned in this publication to interval A of the Pennsylvanian.

Locally, rocks of Cambrian and Ordovician age directly underlie the Pennsylvanian. This relationship is well illustrated in eastern Las Animas County, in the southern portion of the State, where sandy, glauconitic, cherty dolomite assigned to the upper part (Ordovician) of the Arbuckle Group occurs in the eastern part of the Apishapa–Sierra Grande uplift (fig. 60). In northern Colorado, Ordovician rocks lithologically similar to those in the south and also considered Arbuckle are present beneath the Pennsylvanian of southern Washington County.

Ordovician rocks are present directly beneath the Pennsylvanian in western Pueblo and eastern Fremont Counties on the northeastern side of the Canon City embayment. These rocks consist of the Manitou Limestone (locs. 143, 768, 778), Harding Sandstone (locs. 761, 762), and Fremont Limestone (locs. 771, 772). The Manitou is a cherty dolomite to finely crystalline limestone (Maher, 1958, p. 66); the Harding is a fine to very fine grained sandstone (Maher, 1958, p. 68); and the Fremont is a “finely granular to medium crystalline dolomite” (Maher, 1958, p. 68). The distribution of Ordovician rocks in this area delimits the southern end of the Front-range uplift, the axis of which was northwest of the crest of the present Front Range. A small area of Precambrian rocks surrounded by Ordovician rocks marks the axis of the Red Creek arch, a Pennsylvanian arch at the southern end of the Front Range (Glockzin and Roy, 1945).

Cambrian or questionable Cambrian rocks underlie the Pennsylvanian in two places: (1) an area northeast of Denver (loc. 853) where the strata have been penetrated by a single drill hole, and (2) a larger area in northwestern Yuma County. At the Denver locality, an unnamed sequence 55 feet thick of purple shaly mudstone, pink dolomite, and orange quartz conglomerate beneath the Pennsylvanian Fountain Formation has been questionably referred to the Cambrian or Ordovician (Scopel, 1964, p. 40-41). In the Yuma County area, Cambrian rocks consist of fine- to coarse-grained quartzose sandstone, locally containing glauconite and referred to the Reagan Sandstone.

Precambrian rocks beneath the Pennsylvanian in eastern Colorado consist of a complex of igneous and metamorphic rocks. Rocks of granitic composition are predominant, but metamorphosed sediments have been penetrated by some drill holes in Las Animas and El Paso Counties (Maher, 1958, p. 63).}

**LOWER BOUNDARY OF PENNSYLVANIAN**

The base of the Pennsylvanian throughout eastern Colorado is marked by an erosional unconformity separating Pennsylvanian strata, generally Morrow in age, from older rocks beneath. In most places, the lower boundary is easily recognized by abrupt lithologic change.

In outcrops along the east flank of the Front Range, except near Colorado Springs and in the Canon City embayment and northern Wet Mountains areas, the base of the Pennsylvanian is placed at the base of arkosic rocks of the Fountain Formation, which rest either on Precambrian granitic rock or on Ordovician and Mississippian carbonate rocks. The Fountain can be traced eastward in the subsurface several miles from its outcrop area, and contact relations similar to those in surface exposures prevail.

In the Colorado Springs area, the base of the Pennsylvanian is placed at the base of a mudstone unit, the Glen Eyrie, considered the basal member of the Fountain Formation by some workers but a separate formation by others (Maher, 1958, table 1). The Glen Eyrie is a sequence of varicolored mudstone, glauconitic limestone, and sandstone that overlies Mississippian carbonate rocks.

In most of the subsurface of eastern Colorado, the base of the Pennsylvanian is placed at the base of an unnamed sequence of strata, informally referred to as rocks of Morrow age, that is composed of a lower unit of limy glauconitic sandstone and sandy limestone and an upper unit dominated by gray to green mudstone. In southeastern Colorado, the basal sandy unit is well developed and is commonly termed informally the *Keys sand*. Rocks of assumed Morrow age in most places...
overlie either Precambrian granitic rocks or carbonate rocks of Ordovician or Mississippian age, and the base is readily distinguished by the contrast in lithology. In southeasternmost Colorado, however, rocks of Morrow age overlie a detrital sequence of mudstone, sandstone, and limestone of Late Mississippian (Chester) age. These Chester rocks contain some red beds and sandstone that are similar to but finer grained than those of the basal Pennsylvanian; they include little or no glauconite.

PALEOTECTONIC IMPLICATIONS

The paleogeologic map of eastern Colorado (pl. 2) shows the tectonic pattern that developed at or near the end of Mississippian time as well as some structural features that developed after Pennsylvanian sedimentation began.

Following deposition of Mississippian strata, eastern Colorado was upwarped and underwent a period of erosion prior to the onset of Pennsylvanian deposition; a regional erosional unconformity beneath Pennsylvanian rocks resulted.

A progressive northward truncation of Mississippian and older rocks from southeastern to northeastern Colorado indicates a greater arching and erosion in the northeastern part of the State than in the southern part before Pennsylvanian deposition. The preservation of youngest Mississippian (Chester) rocks in southeasternmost Colorado indicates pre-Pennsylvanian downwarping along the axis of the Hugoton embayment of the Anadarko basin and suggests that Pennsylvanian seas invaded this region first and transgressed from there northward and westward.

The physical status of the Apishapa–Sierra Grande uplift and the Red Creek arch, south of the Front Range uplift, at a time immediately preceding earliest Pennsylvanian deposition is not clear. The isopach patterns of interval A outline the locations of these elements in a general way, and the fine-grained character of the Lower Pennsylvanian deposits along the eastern margins of the positive element indicates that uplift could not have been great at that time.

INTERVAL A

FORMATIONS INCLUDED

Strata assigned to interval A in eastern Colorado (pl. 13) consist of: unnamed rocks, generally classified as rocks of Morrow age, in the subsurface of most areas; the Keyes sand, an informal term, and overlying undifferentiated Morrow rocks in southeasternmost Colorado; the Glen Eyre Shale Member of the Fountain Formation in outcrops near Colorado Springs; and the basal 250 feet of the undivided Fountain Formation in outcrops along the Front Range at least as far north of Colorado Springs as Perry Park (fig. 59).

The Glen Eyre Shale Member contains a fauna that includes Foraminifera, ostracodes, and conodonts, as well as scattered plant remains. Although the age of the fauna has been debated (Kellett, 1948; McLaughlin, 1952; Lehmann, 1953), most recently it has been considered to be Morrow (Chron, 1958, p. 13). Foraminifera and ostracodes of probable Morrow age have been reported from a few drill holes (locs. 174, 293, 370, 791, 799) in the subsurface of eastern Colorado. In addition, a varied marine fauna from about 250 feet above the base of the undivided arkosic Fountain Formation at Perry Park (Ellis, 1966), about 35 miles south of Denver, indicates that at least the basal part of the Fountain Formation in that area should be included in interval A.

Assignment of units to interval A in some places is based on lithologic criteria. The Glen Eyre Shale Member of the Fountain is similar in lithology to rocks of Morrow age in the subsurface throughout much of eastern Colorado. This unit and other units assigned to interval A are thought to be continuous with rocks known to be of Morrow age in western Kansas (Maher and Collins, 1952, sheet 1; Maher, 1953).

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A is a readily recognizable and generally consistent horizon throughout much of eastern Colorado, particularly in the eastcentral and southeastern parts of the State. At most localities in these areas, interval B is dark-gray to black siliceous mudstone and argillaceous siliceous limestone that contrast markedly with underlying greenish-gray to gray glauconitic mudstone and limestone that are characteristic of the upper part of interval A (pl. 10D, sec. W-W').

Near the northeast flank of the Apishapa uplift (fig. 60) and the east flank of the Front Range, interval B is more arkosic than elsewhere and lacks the argillaceous siliceous limestone and mudstone typical of other areas. In the Apishapa and Front Range localities, the upper contact of interval A is placed at the base of the first well-developed sandstone or arkosic sandstone above typical glauconitic limestone and mudstone of interval A. This horizon seems to be reasonably consistent in most places (pl. 10E, secs. Y-Y' and Z-Z').

The contact between the Glen Eyre Shale Member and the remainder of the Fountain Formation, where exposed in the Colorado Springs area, is a consistent, well-defined boundary. It is indicated by a lithologic change from variegated mudstone and quartzitic sandstone characteristic of the Glen Eyre to the reddish arkosic sandstone and conglomerate characteristic of the overly-
ing part of the Fountain Formation that is assigned to interval B.

At some localities adjacent to the Frontrange and Apishapa uplifts, such as at Perry Park, interval A is arkosic. The upper contact is arbitrarily projected into such places from nearby localities (for example, loc. 311) where the contact can be definitely located.

Locally in southeastern Colorado, interval B is absent; interval A is unconformably overlain by younger Pennsylvanian intervals (interval C at locs. 174, 795, 815, 822; interval E at loc. 275). In all these places the upper contact of interval A is located at a lithologic break between glauconitic limestone and variiegated mudstone of interval A and reddish arkosic sandstone, siltstone, and sandy limestone representing the basal parts of younger intervals.

On the east end of the Apishapa uplift (loc. 16), a thin remnant of interval A, as indicated by the presence of glauconite, is overlain unconformably by reddish siltstone and red and brown cherty dolomite of Wolfcamp age.

Interval B is absent in northeasternmost Colorado. Where interval A is present in this part of the State, it is overlain by interval C. The contact between the two intervals is somewhat arbitrary but generally is placed at the base of a sequence of brown to gray cherty limestone and dark mudstone or a quartzitic sandstone. These rocks overlie a sequence of variiegated siltstone, mudstone, and white sandstone, commonly glauconitic, that is assigned to interval A (pl. 10D, sec. $UU'$).

**THICKNESS TRENDS**

Interval A has a regional maximum thickness of more than 400 feet just north of the buried Apishapa uplift and in the Hugoton embayment of the Anadarko basin in Baca County, southeasternmost Colorado. From these areas the interval thins toward and laps onto the flanks of the Apishapa-Sierra Grande uplift, and it thins north and east to a pinch-out in northern Kansas and Nebraska. The interval seems to be absent in a small area in Yuma County near the Kansas and Nebraska boundaries.

An apparent thinning and pinching out of interval A toward the east margin of the northern Front Range may be the result of inability to distinguish in many places between intervals A and B; in the present study some rocks of Morrow age possibly are included in interval B, so any resulting thinness of interval A is not necessarily a depositional feature. On the other hand, north of Denver in outcrops along the Front Range, no evidence of the presence of Morrow rocks is known.

Thicknesses of interval A are noticeably variable adjacent to the northeast corner of the Apishapa-Sierra Grande uplift. Many of the thickness changes are probably due to block faulting and erosion during or following deposition of the interval (pl. 10E, sec. AA $AA'$). Elsewhere, particularly in southeasternmost Colorado, local variations in the thickness of interval A are probably the result of the cutting of channels into underlying Mississippian rocks before deposition of interval A and filling of these channels with sediments during interval A time. Thus, locally thick sections commonly indicate the presence of channel fills, and thinner sections represent interchannel areas.

Interval A is recognized in outcrop only at Perry Park, 35 miles south of Denver, and in the Colorado Springs area, where it is represented by the Glen Eyrie Shale Member of the Fountain Formation. The Glen Eyrie pinches out along the Front Range a short distance south of Colorado Springs and is believed to be absent both in outcrops and in most of the subsurface of the Canon City embayment.

**LITHOFACIES TRENDS**

Arkose and feldspathic sandstones are absent or compose only a small percentage of interval A in most of eastern Colorado, although they are common in parts of all other intervals. Arkose sandstone is the dominant lithology in interval A outcrops only at Perry Park (loc. 775) (Ellis, 1966, p. 157). It composes as much as 20 percent of the interval in some drill holes in northeastern Colorado and about 40 percent of the interval in a drill hole a few miles southeast of Pueblo (loc. 160), but elsewhere it is either absent or of only minor significance.

The stratigraphic sequence of lithologic types within interval A in eastern Colorado is shown by cross sections (pl. 10D, E). In almost all localities the interval consists of detrital rocks characterized by a basal unit of variable thickness composed of sandstone, conglomeratic sandstone, limy sandstone, sandy limestone, and calcarenite; it also commonly contains glauconite, which is a useful criterion for identification of the interval. This lower unit is overlain by a thicker, less variable unit of mudstone, argillaceous to pure limestone, and small amounts of sandstone. Locally, in southeastern Colorado, thin beds of coal and carbonaceous mudstone are present near the top of the interval.

In southeastern Colorado, particularly in Baca, Prowers, and Bent Counties, the basal sandy unit is commonly assigned to the Keyes sand, a prominent basal sandstone at the Keyes dome in Cimarron County, Okla. (Wilson, 1958, p. 69). This basal unit differs considerably in thickness from place to place but has a maximum thickness — in excess of 150 feet — in northeastern Baca and southeastern Prowers Counties; in general it thins from Baca and Prowers Counties southeastward; it both thins and becomes discontinuous to the east, north, and northwest.

In northeastern Colorado, division of interval A into
two units is not practical at many localities because the various lithologic components are rather evenly distributed throughout the interval.

Sandstone is dominant in interval A in three main areas (pl. 3B) in eastern Colorado: (1) northeastern Colorado, (2) an area along the Front Range between Denver and Colorado Springs, and (3) a small area southeast of the inferred juncture of the Front Range and Apishapa uplifts. The area in northeastern Colorado is the largest. In southeastern Colorado (pl. 10D, E) the Keyes sand is represented by an appreciable thickness of sandstone and sandy limestone at the base of the interval, but the distribution of this unit is partly masked, as shown on plate 3B, by the excessively thick overlying unit of mudstone and limestone.

Limestone is abundant, although not dominant, in two areas of interval A—a large, crudely ovate area in east-central Colorado north of the Apishapa uplift, and the Hugoton embayment in southeasternmost Colorado.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Interval A in eastern Colorado seems to have been derived from at least four source areas. One major source, indicated by an abundance of sandstone in northeastern Colorado (pl. 3B), was to the northeast in Nebraska and northwestern Kansas. A lesser source, also suggested by an abundance of sandstone in adjoining areas, was the northern part of the Front Range uplift. A local source was the Apishapa uplift, near its west end. The Keyes sand, which is the basal detrital unit of interval A in southeastern Colorado, was supposedly derived from Oklahoma to the southeast (Maher, 1953, p. 2486), although some parts of it, particularly the beds of calccrenite, may have come from erosion of middle and lower Paleozoic carbonate rocks exposed in the western part of the Apishapa uplift.

Interval A seems to be largely the product of a shallow-water marine and nearshore environment and was deposited during transgression of the sea from the Hugoton embayment in southeastern Colorado northward and westward across eastern Colorado (Mallory, 1960, p. 24). Fossils from drill holes and in outcrops attest to the marine origin of at least part of the interval. A flora in the Glen Eyrie Shale Member and coal beds near the top of the interval in southeastern Colorado imply that continental or coastal paludal environments were also present. Arkosic sandstone, locally present in the subsurface and in outcrops along the Front Range, probably was deposited largely by streams on alluvial plains.

Variegated colors characteristic of the interval and the presence of glauconite suggest unrestricted circulation of oxidizing waters, in contrast to the euxinic environments implied for many parts of interval B.

PALEOTECTONIC IMPLICATIONS

Of those tectonic elements that developed in eastern Colorado by Late Mississippian time, possibly only parts of the central and northern Frontrange uplift and the western Apishapa uplift were actively positive during Early Pennsylvanian time. Coarse sediment nearby provides evidence of a northern Frontrange uplift, mainly in southeastern Wyoming, but tectonism may have extended into northeastern Colorado, where interval A is no longer preserved or is little known (pl. 3B).

Smaller areas of abundant sandstone to the south (pl. 3B) are close to margins of inferred positive elements and probably indicate proximity to areas of slight uplift. Near Colorado Springs the dominance of mudstone in the Glen Eyrie Shale Member indicates the presence of a slightly elevated nearby source. On the north and east flanks of the Apishapa uplift, fine-grained rocks of interval A may be evidence that at least the eastern half of the uplift remained low enough to be a depositional area in interval A time.

The Freezeout Creek fault at the east end of the Apishapa uplift, a probable fault along the north flank of the uplift, and block faults to the east probably developed after deposition of interval A and possibly interval B.

INTERVAL B

FORMATIONS INCLUDED

Most strata assigned to interval B in eastern Colorado are Atoka in age and consist of (1) unnamed rocks, generally referred to as rocks of Atoka age, in much of the subsurface, (2) the lower part, or member, of the Fountain Formation along the east flank of the Front Range north of Lyons (fig. 59, loc. 757), and (3) an undetermined part of the undifferentiated Fountain in outcrops extending from Lyons to south of the Canon City embayment and in the subsurface immediately adjacent to the Front Range. Assignment of these strata to interval B is based upon scattered paleontological evidence in the subsurface and in the lower part of the Fountain Formation near the Wyoming border.

East and north of the Frontrange and Apishapa uplifts (fig. 60), the rocks of Atoka age in the subsurface form a distinctive lithologic unit but have never been given a formal name. Fusulinids of Atoka age, consisting of various species of Fusulinella and Profusulinella, have been noted in the sample logs of 16 drill holes used as control points in this part of Colorado and provide the basis for assigning the rocks of Atoka age to interval B.

The Fountain Formation is a sequence consisting of arkosic sandstone, conglomerate, sandy siltstone, and minor interstratified fine-grained quartzose sandstone and carbonate rock that crops out along the east flank of the Front Range from the Canon City embayment north
to the Wyoming border. Strata assignable to the Fountain Formation are also present in the subsurface adjacent to the Front Range and to the buried Apishapa–Sierra Grande uplift.

North of Lyons, Colo. (loc. 757), as far north as the southern Laramie Range and Laramie basin in Wyoming, the Fountain Formation has been divided into upper and lower parts (Maughan and Wilson, 1960).

The lower part of the Fountain Formation is a distinctive unit largely composed of sets of coarse-grained arkosic sandstone and conglomerate interstratified with sets of poorly sorted sandy siltstone and minor amounts of very fine grained quartzose sandstone and carbonate rock. It has a readily mappable upper contact; thus, informal member status seems justified. Fusulinids dated as late Atoka in age have been collected from a horizon near the top of the lower member at Box Elder Creek (loc. 334) near the Wyoming border (Maughan and Wilson, 1963, p. 103). On the basis of this evidence, the entire lower member of the Fountain is assigned to interval B, although part of the member possibly is equivalent in age to strata assigned to interval A in the subsurface.

Fusulinids of late Atoka age have been collected from strata in the lower part of the Casper Formation along the Laramie Range in Wyoming (Maughan and Wilson, 1963, p. 103); these beds, therefore, are considered equivalent to the lower member of the Fountain Formation.

Rocks of Atoka age probably are also present in the undifferentiated Fountain Formation south of Lyons, but in that area they cannot readily be separated from younger strata of overlying interval C. The extension of interval B from localities where paleontological data are available to localities where such data are absent is based upon lithologic similarities and upon the correlation of marker beds.

UPPER BOUNDARY OF INTERVAL B

In eastern Colorado, interval B is everywhere overlain by interval C. Lithologic criteria used to distinguish the contact between the two intervals differ from area to area. At only a few localities (173, 224, 318, 751) is fusulinid control adequate to pick the upper contact of interval B on biostratigraphic evidence alone.

In much of the eastern half of eastern Colorado, interval B is characterized by dark-gray to black argillaceous, siliceous limestone that contains brown to black chert nodules and by thin to thick sets of black carbonaceous mudstone interstratified with limestone. Interval C in this area is also characterized by interstratified limestone and mudstone; but the limestone, which is gray to brown, tends to be lighter and contains less argillaceous material and less silica than limestone in interval B. Mudstone in interval C is gray to green and is not as dark as mudstone in interval B. Hence, in this part of eastern Colorado, the contact between intervals B and C is placed at the top of the highest dark argillaceous siliceous limestone that occurs beneath lighter colored limestone typical of interval C. This contact probably constitutes a consistent horizon across much of northeastern and easternmost Colorado.

Along the east flank of the Front Range between Lyons (loc. 757) and the Wyoming border, the upper boundary of interval B is placed at the top of the lower member of the Fountain Formation. This member, as described earlier, consists largely of interstratified coarse-grained arkosic sandstones and conglomerates and poorly sorted sandy siltstones. The upper member of the Fountain Formation also consists of arkosic sandstone and conglomerate, but it contains, at most, only a few sets of poorly sorted sandy siltstones. The arkosic sandstones and conglomerates in the upper member are interstratified with bimodal, moderately well sorted, fine to very fine grained feldspathic sandstones. These sandstones commonly contain rounded medium to very coarse grains of quartz and feldspar, either disseminated through the rock or concentrated along lamination planes. Although the lower member of the Fountain contains a few thin sandstone sets similar in composition to the fine-grained units in the upper member, the sets in the lower member do not contain disseminated coarse sand grains. The upper boundary of the lower member is placed at the lowest unit of very fine to fine-grained sandstone that contains scattered medium to coarse sand grains. This unit is believed to be consistent and traceable from Wyoming at least as far south as Lyons, Colo. (Maughan and Wilson, 1963, p. 98).

Both the upper and the lower members of the Fountain Formation probably are continuous as far south as Left Hand Canyon (loc. 765), just north of Boulder, Colo., but the contact between the members is not everywhere clear. South of this point, the two members cannot be differentiated.

The boundary between interval B and interval C in the area directly east of the Fountain outcrop belt, along the flank of the Front Range, is difficult to pick because the two intervals contain similar types of rock. In places this boundary must be determined by projection from nearby localities where the contact is clear. In general, however, the lower part of interval C contains some dark siliceous limestone and more mudstone and other detrital rock than does the upper part of interval B; hence, the upper boundary of interval B at localities where these rock types are present is placed at the highest conspicuous carbonate unit below a sequence that is mostly mudstone or arkosic sandstone. In parts of Prowers County, the uppermost part of interval B locally
is marked by thin beds of anhydrite. In some other places, fusulinid data are useful in determining the approximate location of the contact.

THICKNESS TRENDS

Interval B is more than 400 feet thick in various small basins scattered along the northeast and east flank of the buried Apishapa–Sierra Grande uplift (pl. 6A). The interval thickens irregularly southward from the Wyoming border along the northern part of the Front Range to a maximum thickness of 650 feet near Lyons (loc. 757); south of the Lyons locality the lower member of the Fountain Formation can not be differentiated. At Colorado Springs and in the Canon City embayment, strata of Atoka age cannot be precisely separated from younger rocks, but they may be somewhat thicker than in the north.

Interval B pinches out along the border of the Apishapa–Sierra Grande uplift. It is absent from the crests of some small uplifts, possibly fault blocks, east of the east end of the Apishapa–Sierra Grande uplift.

Throughout much of its area of distribution, interval B maintains a fairly consistent thickness between 100 and 400 feet, but it exhibits a gradual northeastward thinning from points of maximum thickness along the east flank of the Front Range and the Apishapa uplift. The interval apparently pinches out as the result of erosion near the east border of Colorado north of the Kansas-Nebraska line.

In most parts of eastern Colorado, interval B is the thinnest and least extensive of the Pennsylvanian intervals.

LITHOFACIES TRENDS

In eastern Colorado, interval B grades from a dominantly sandstone unit along the flanks of the Front range and the Apishapa uplifts to a dominantly carbonate unit in northeastern Colorado (pl. 6B). Arkose constitutes as much as 30 percent of interval B throughout a considerable area in the subsurface northeast of the Apishapa uplift. The percentage of arkose decreases both eastward and northward from the Front range and Apishapa uplifts, but tongues of arkose extend eastward and northeastward into the mudstone and carbonate sequence which is typical of the interval in most of eastern Colorado (pl. 10D, E, secs. X-X', Y-Y', Z-Z').

Interstratification of arkosic sandstone, mudstone, and limestone in interval B is especially common in Crowley, Otero, El Paso, Pueblo, and Weld counties. In these areas, the mudstone is pale gray to greenish gray, and is lighter in color than the mudstone typical of interval B in easternmost Colorado.

In easternmost Colorado, interval B consists of interstratified dark mudstone and siliceous or argillaceous limestone that, for some areas, have been interpreted as cyclic deposits. In parts of southeastern Colorado, notably in Baca, Prowers, and Bent Counties, interval B is composed of interstratified limestone and mudstone that have high resistivity on electric logs. This sequence is referred to as the "thirteen-fingered limestone" (Wilson, 1968, p. 71). The repetition of beds also suggests a cyclic origin. This particular sequence, however, can be recognized only in a restricted area.

A marked facies change is evident on plate 6B, from sandstone in the western and southwestern parts of eastern Colorado to mudstone and finally to carbonate rock in eastern, east-central, and northeastern Colorado. This systematic change contrasts with the more irregular distribution of lithologic types in interval A (pl. 8B).

SOURCES AND ENVIRONMENTS OF DEPOSITION

The principal source of detritus in interval B of eastern Colorado, as indicated by the sandstone pattern on the lithofacies map (pl. 6B), was in the central and southern parts of the Front range uplift. Of lesser importance, but still a source of considerable sediment, was the western part of the Apishapa uplift. A positive area in Nebraska may have been a minor source of fine detritus deposited in northeastern Colorado (pl. 6B).

Along the east flank of the Front range uplift and probably also along the north flank of the Apishapa uplift, interval B apparently was formed in a continental environment on an alluvial plain bordering the sea that extended to the east and north. Evidence of fluvial deposition on coalescing fans is furnished by mean dip orientations of cross-stratification in the lower part of the Fountain Formation along the flank of the northern Front range and in the undifferentiated Fountain Formation along the central and southern parts of the Front range (Howard, 1966). The presence of thin units of fusulinid-bearing carbonate rock in lower parts of the formation in northernmost Colorado indicates minor periods of marine transgression into that area.

Interval B throughout most of eastern Colorado accumulated in a marine environment. Scattered fusulinid faunas support this conclusion. Black to dark-gray siliceous mudstones and limestones that characterize this interval in the easternmost parts of the State indicate a fairly restricted euxinic environment, as opposed to more normal marine environments that characterize interval A below and interval C above. The belt of dark mudstone and limestone is separated from the arkosic sandstone of the Fountain Formation to the west by a belt of lighter colored carbonate rock and mudstone representing deposits formed in a relatively normal marine environment between euxinic conditions to the east and fluvial conditions to the west.
PALEOTECTONIC IMPLICATIONS

Both the Frontrange uplift and the Apishapa uplift were active during deposition of interval B. The entire Frontrange positive element was elevated, in contrast with localized uplift of this element during interval A time. The large amount of arkosic sandstone and conglomerate in interval B along the east flank of the present Front Range supports the concept of vigorous erosion of a landmass that stood at moderate to great height. Lesser amounts of sandstone along the Apishapa uplift suggest that this positive element was less active than the Frontrange uplift and that its eastern part remained at a lower elevation than the western end.

The Hugoton embayment of the Anadarko basin in southeastern Colorado underwent only slight subsidence during the deposition of interval B (pl. 64). The initial rise of the Apishapa uplift may have restricted circulation of the sea within the embayment, causing the euxinic conditions indicated by dark siliceous mudstone and limestone.

The Freezeout Creek fault along the east flank of the Apishapa–Sierra Grande uplift and several small faults bordering fault blocks to the east probably are largely post-interval B features. Lithofacies patterns of interval B adjacent to these fault blocks give no evidence of nearby source areas, thus suggesting that most of the fault movement postdates deposition of the strata.

INTERVAL C FORMATIONS INCLUDED

Interval C composes that part of the Fountain Formation that is believed to be approximately of Des Moines age in outcrops at the base of the Front Range, in the Canon City embayment, and in the subsurface directly to the east. In the subsurface farther east, the interval consists of unnamed and undivided rocks commonly referred to as rocks of Des Moines age. Locally, in southeastern Colorado, it is divided into the Cherokee Group below and the Marmaton Group above (pl. 13).

Along the northern part of the Front Range in Colorado, the lower part of the upper member of the Fountain Formation, perhaps as much as 300 feet in places, represents interval C. Fusulinids collected from near the base of the upper member of the Fountain Formation at Box Elder Creek (fig. 59) and from equivalent strata of the Casper Formation in Wyoming have been identified as of early Des Moines or late Atoka age (Maughan and Wilson, 1963, p. 103) and provide the basis for assigning the lower part of the upper member to interval C. Rocks of Des Moines age probably are also present in the undivided Fountain Formation south of Lyons, Colo., but cannot be distinguished lithologically from other parts of the formation in that area.

Fusulinids of Des Moines age, reported from 24 drill holes used as control points in eastern Colorado, helped determine the presence of interval C in that area. The fusulinids consist of advanced species of Fusulinella, Fusulina or Beedleina, and Wedekindinella. Physical correlation of the rocks in eastern Colorado with the Cherokee and Marmaton Groups of known Des Moines age in western Kansas provides another basis for the assignment of these rocks to interval C.

UPPER BOUNDARY OF INTERVAL C

Interval C is everywhere conformably overlain by interval D except locally near the east flank of the Apishapa–Sierra Grande uplift in southeastern Colorado (fig. 59, loc. 850), where interval C is overlain unconformably by interval E.

Criteria used to distinguish the upper boundary of interval C differ from place to place. The contact is defined on biostratigraphic evidence from five drill holes (locs. 140, 181, 361, 782, 814), where fusulinids diagnostic of age are closely spaced in vertical sequence. Elsewhere, the contact is defined by tracing key beds or distinctive lithologic sequences from localities where the contact has been established.

In parts of eastern Colorado fairly remote from Pennsylvanian uplifts, the base of interval D is a detrital unit, 10-150 feet thick, between the dominantly carbonate rocks (Marmaton Group equivalents) of the upper part of interval C and the dominantly carbonate rocks characterizing most of interval D (pl. 10D, E, secs. W-W', X-X', Y-Y', Z-Z'). This detrital unit is considered to be the basal part of the Kansas City Group of interval D. Most of it is coarser grained than the mudstone units in interval C below, and its base forms a distinctive marker. Westward and southward toward the areas of Pennsylvanian uplift, however, this basal detrital unit of interval D is progressively thicker and tends to merge with dominantly detrital sections of intervals C and D in those directions (pl. 10D, E, secs. U-U', X-X', Y-Y', AA-A'A').

Differences in the carbonate rocks in upper parts of interval C as compared with interval D are helpful in locating the contact between these intervals. Carbonate rocks in interval C commonly contain more chert than those in interval D; likewise, fragments of the coral Chaetetes, not present in interval D, occur in C. Oolites are common in the basal carbonate rocks of interval D, but are rare in interval C. Thin to thick units of brown detrital rock, interstratified with carbonate rock, locally occur in interval D in areas remote from the Pennsylvanian uplifts, but they are not present in interval C.

Near areas of Pennsylvanian uplift, subsurface rocks of both interval C and interval D are predominantly...
detrital and apparently grade into the Fountain Formation of the outcrop. Selecting a consistent contact between these intervals is difficult in such areas. In most places, the contact is projected into the detrital sequence by extrapolating thickness trends of intervals C and D from areas where the contact is well established. In other places, however, the contact is established by the use of lithologic or biostratigraphic criteria.

In southeastern Colorado, at some localities near the Apishapa uplift, the upper contact of interval C is located at the horizon of change from dominantly medium to coarse grained arkosic sandstone of interval C to dominantly fine grained siltstone and mudstone interstratified with lesser amounts of arkosic sandstone and carbonate rock of interval D. This horizon has been shown to correspond fairly well to the position of the contact as located by projection from established localities.

In outcrops north of Lyons, Colo., and within the undivided Fountain Formation south of Lyons the contact between intervals C and D lies within the upper member of the Fountain Formation. At Owl Canyon (loc. 79), the contact is located by projection from the subsurface, but elsewhere in outcrops of that area, the contact has not been determined. Interval C is shown as present but is not assigned a specific thickness.

THICKNESS TRENDS

Interval C is the thickest of all the Pennsylvanian intervals in eastern Colorado. In an asymmetric trough just north of the Apishapa uplift, it attains a maximum thickness in excess of 1,500 feet (pl. 7A). In the Colorado Springs area, a comparable or greater thickness is preserved; but the age relations are uncertain there, and some of the rocks assigned to the interval may be younger than Des Moines. The interval thins within a short distance northeast of its areas of maximum thickness and is about 300-400 feet thick in the north half of eastern Colorado.

Interval C has a wide range in thickness in southeastern Colorado, notably to the east and northeast of the Apishapa-Sierra Grande uplift (pl. 10E, sec. AA-A'A') in Baca and Prowers Counties. The thickness differences are at least in part due to thinning of the interval across the crest of buried fault blocks. The isopach map (pl. 6A) indicates that a depositional trough — possibly fault bounded — is present along the northeast edge of the Apishapa uplift. Similar but smaller fault troughs are suggested by the isopach patterns of interval C in southeasternmost Colorado.

Outcrops of the Fountain Formation along the Front Range in eastern Colorado undoubtedly contain rocks of Des Moines age equivalent to those of interval C, but their thickness has not been determined.

LITHOFACIES TRENDS

Distribution of rock types in interval C of eastern Colorado corresponds closely to that of interval B, but facies differences are more pronounced (pl. 7B). Arkosic sandstone and conglomerate are the dominant lithologic types in the area adjacent to the northeast edge of the Apishapa uplift and along the east flank of the Frontrange uplift. Wedges of red-brown arkosic sandstone and conglomerate extend northward and eastward from the Apishapa and eastward from the Frontrange into the finer grained detrital rocks and the carbonate rocks that characterize the easternmost Colorado area (pl. 10D, E, secs. X-X', Y-Y', Z-Z').

The basinward spread of coarse detrital rocks derived from Pennsylvanian uplifts is somewhat greater in interval C than in the other Pennsylvanian intervals, except for possibly part of interval B. The only areas in easternmost Colorado in which detrital rocks are the dominant lithologic types are scattered patches in southeastern and east-central Colorado and a larger area in northeastern Colorado. The latter area is on the west edge of a belt of dominantly fine grained detrital rock that trends northwestward in Nebraska and Kansas.

Rocks of Des Moines age in the subsurface of eastern Colorado, in areas remote from the Frontrange and Apishapa-Sierra Grande uplifts, consist of interstratified gray to green mudstone and gray to brown limestone that commonly contain chert nodules. Fragments of the coral Chaetetes are common in cuttings of limestone from many localities of eastern Colorado. In drill holes where interval C consists dominantly of limestone, the interval generally can be divided into a lower part that has more mudstone than limestone and an upper part that is largely limestone (pl. 10D, E, secs. W-W', X-X', Y-Y').

The boundary between the upper and lower parts of interval C in most areas does not seem to follow a consistent horizon; it exhibits considerable vertical variation. In some areas, however, particularly near the Kansas border and in southeastern Colorado, this boundary is fairly consistent; there, rocks of the lower part of interval C are referred to the Cherokee Group, and those of the upper part, to the Marmaton Group.

Along the Front Range north of Lyons, where interval C consists of the lower part of the upper member of the Fountain Formation, the interval is composed primarily of arkosic sandstone and conglomerate alternating with beds of relatively fine grained, moderately well sorted feldspathic sandstone containing scattered medium to coarse sand grains. Farther north, near the Wyoming border, tongues of carbonate rock are present and thicken northward along the outcrop.
SOURCES AND ENVIRONMENTS OF DEPOSITION

Source areas and depositional environments for interval C are interpreted to be mostly the same as for interval B.

The coarse-grained arkosic rocks of interval C, as determined from lithofacies patterns and from cross-strata directions in outcrops along the Front Range (Howard, 1966), were derived primarily from the northern part of the Apishapa–Sierra Grande uplift and from the Frontrange uplift. The eastern part of the Apishapa–Sierra Grande uplift was a source of fine detrital sediment, including some sand, in southeastern Colorado. The belt of fine-grained detrital rocks in northeastern Colorado (pl. 7B) probably indicates a source in a positive area (Cambridge arch) in Nebraska to the northeast, but some of the detrital rocks in the belt possibly had their source in Pennsylvanian uplifts of Colorado.

Interval C was deposited on alluvial fans or alluvial plains adjacent to the Frontrange uplift and the northern part of the Apishapa uplift and in a marine environment to the north and east. Evidence for the alluvial origin of interval C rocks in outcrops adjacent to the uplifts is the same as for interval B and consists of low-angle trough cross-stratification dipping in divergent directions that indicate radial stream-channel patterns in the arkosic nonmarine environments in areas marginal to the uplifts are indicated by the interstratification of thin sets of fossiliferous carbonate rock in a predominantly arkosic sequence.

A normal marine environment with free circulation is indicated for areas where the dominant deposit was either carbonate rock or mudstone, as in northeastern Colorado. Scattered fusulinid and coral faunas support this interpretation. A euxinic environment, as indicated by dark siliceous limestone and mudstone in interval B of eastern Colorado, is not represented in interval C.

PALEOTECTONIC IMPLICATIONS

Interval C was the time of greatest uplift during the Pennsylvanian of the Frontrange and the Apishapa–Sierra Grande positive areas and of maximum sinking in troughs adjacent to these uplifts, particularly in the area along the northeast edge of the Apishapa uplift. The magnitude of the sinking at this time is shown by the thickness of interval C deposits, which is greater than for any of the other Pennsylvanian intervals in eastern Colorado (pl. 7B). Evidence for maximum uplift of positive areas during the time of interval C is the volume of coarse-grained predominantly arkosic sandstone and conglomerate derived from the highlands. These detrital sediments extend farther into the basin than similar deposits of any other interval in the Pennsylvanian (pl. 7B).

Both the Frontrange uplift and the western part of the Apishapa uplift are inferred to have been land areas of moderate to high elevation during the time of interval C. The eastern part of the Apishapa–Sierra Grande uplift probably stood at somewhat lower elevations, as indicated by the fineness of detrital sediment in rocks of the basin adjacent to these uplifts.

The Cambridge arch in Nebraska probably was a low land area and supplied only minor amounts of fine detrital sediments to the basin to the southwest. The Hugoton embayment of the Anadarko basin in southeastern Colorado was an active negative element during interval C time and underwent subsidence locally in excess of 800 feet in parts in southeastern Colorado (pl. 7A).

The Freezeout Creek fault bordering the east flank of the Apishapa–Sierra Grande uplift and other faults of less displacement bordering fault blocks on the east side of the uplift probably were active during or immediately preceding interval C time. The time of faulting is indicated by overlap of interval C on some of the fault blocks (pl. 10E, sec. AA′-AA′). Facies patterns of interval C in the area of the fault blocks do not suggest a nearby source of sediment, such as might be expected during uplift of the blocks. The small areal extent of the blocks, however, would have precluded the supplying of any large amount of sediment to the surrounding basin, and material eroded from the blocks would have been mostly fine grained; hence, indications of uplift furnished by coarsening of sediment adjacent to the blocks would be slight.

INTERVAL D

FORMATIONS INCLUDED

Interval D along the Front Range in eastern Colorado consists of the upper part of the upper member of the Fountain Formation in areas north of Owl Canyon, a part of the upper member of the Fountain Formation from Owl Canyon south to Lyons, and an unnamed upper part of the Fountain Formation south of Lyons and in the Canon City embayment. These rocks are of probable Missouri age. Strata assigned to interval D in the subsurface consist of the undifferentiated Lansing and Kansas City Groups in the easternmost part of the State and unnamed, undivided rocks referred to as rocks of Missouri age in other parts of eastern Colorado.

Strata assigned to interval D in the Fountain Formation north of Denver consist of interstratified reddish medium- to coarse-grained arkosic sandstone and conglomerate and lighter colored finer grained feldspathic sandstone that contains scattered medium to coarse sand grains. From Stonewall Creek (loc. 766) northward to the Wyoming border, interval D contains a few units of carbonate rock that thicken northward (Maughan and Wilson, 1960, fig. 3). South of Denver, strata of the
Fountain Formation included in interval D contain many thin to thick sets of poorly sorted sandy and clayey siltstone beds.

Although interval D is present in outcrops throughout the extent of the Fountain Formation, it is not easily differentiated from the underlying interval C or the overlying interval E, and no specific thickness is assigned except at Owl Canyon (loc. 79). There, the upper part of the upper member of the Fountain Formation, about 180 feet thick, is assigned to interval D on the basis of lithologic similarity to rocks in the subsurface farther east (pl. 10D, sec. X-X').

In the subsurface in the Canon City embayment and just east of outcrops of the Fountain Formation in northern Colorado, boundaries for interval D have been assigned within the Fountain. Individual beds or sequences of beds can be traced from these areas eastward into eastern Colorado and Kansas, where a Des Moines age has been established.

In easternmost Colorado, interval D consists of the undifferentiated Lansing and Kansas City Groups and is composed of a thick massive carbonate unit that contains a few interbeds of fine-grained detrital rock. This carbonate unit is readily distinguished from overlying and underlying units in that area, but as it is traced westward towards areas of Pennsylvanian uplift, it gradually changes facies, becoming a sandy detrital rock (pl. 10D, E, secs. X-X', Y-Y', Z-Z', AA-AA'). In this transition area interval D consists of unnamed strata generally referred to as rocks of Missouri age that are composed of interstratified limestone, dolomite, sandstone, and mudstone.

Fusulinids of Missouri age, characteristic of interval D, have been identified from 22 control points in the subsurface of eastern Colorado. The fusulinids include species of *Triticites* and *Eowaueringella*, notably *T. irregularis*, *T. burgessae*, and *E. ultimata*.

**UPPER BOUNDARY OF INTERVAL D**

Interval D is conformably overlain by interval E throughout most of eastern Colorado. An exception to this is in outcrops of the Fountain Formation along the east flank of the Front Range from Owl Canyon northward to the Wyoming border and eastward for a few miles in the subsurface (pls. 8A, 9A). In this area, interval D is overlain with erosional unconformity by Permian strata of Wolfcamp age.

Criteria used to distinguish interval D from interval E in eastern Colorado differ from place to place. At a few control points, fossils are adequate to permit differentiation on the basis of age (locs. 162, 181, 361, 814). At all other localities, the contact must be determined on the basis of lithologic criteria or by tracing the boundary from adjacent well-established control points.

In parts of eastern Colorado where interval D consists primarily of carbonate rock assignable to the undifferentiated Lansing and Kansas City Groups, the upper boundary of interval D is readily recognizable at the base of a sequence of dominantly fine grained sandstone and mudstone, from 5 to 50 feet thick, which belongs to the undifferentiated Pedee and Douglas Groups in overlying interval E (pl. 10D, E, secs. W-W', Z-Z', AA-AA'). (The Kansas Geol. Survey now regards Pedee strata, which are of Missouri age, as part of the Douglas Group.) This detrital unit is a consistent horizon marker separating carbonate rocks of the Shawnee Group in interval E from carbonate rocks in interval D below.

In northeastern Colorado, the Pedee and Douglas Groups cannot be recognized. Intervals D and E both consist of limestone and dolomite (pl. 10D, sec. X-X'). In this area the upper boundary of interval D is projected from adjacent localities where it is better defined.

The contact between intervals D and E is projected somewhat arbitrarily into areas of eastern Colorado, especially those marginal to Pennsylvanian uplifts, where the two intervals consist of dominantly detrital rock. At many localities, the contact is placed for convenience at the top of a unit of limestone which is overlain by a thick unit of gray-brown mudstone, sandstone, and some limestone, considered to represent the basal part of interval E. Where this limestone unit can be traced eastward to areas in which the age relations are better known, its top is near the top of rocks of Missouri age.

In outcrops along the Front Range between Redstone Creek (loc. 755) and Owl Canyon, the contact is placed at the top of the highest arkosic sandstone and conglomerate unit below unconformably overlying units of cross-stratified fine-grained orange sandstone and gray carbonate rock of the Permian Ingleside Formation. From Owl Canyon northward to the Wyoming border, the contact is at a slightly weathered zone which marks a period of erosion within a sequence of fine-grained quartzose sandstone and carbonate rock (Maughan and Wilson, 1960, fig. 3, p. 37). In this area the Ingleside strata onlap southward onto slightly beveled strata of the underlying Fountain Formation.

In the subsurface for a few miles east of Owl Canyon, the contact between interval D and Wolfcamp strata is placed, in some areas, at the top of the highest arkosic unit, above which are interstratified sandstone and carbonate rocks of the Ingleside Formation (loc. 216). In other areas, to the north and east, the contact is the top of a thick limestone and dolomite unit underlying a series of interstratified siltstone, sandstone, dolomite, and anhydrite in the Ingleside Formation.

**THICKNESS TRENDS**

Interval D exhibits a more consistent thickness than do older intervals in eastern Colorado. In general, it thins
gradually from about 500-600 feet in the south to 200-300 feet in the north. Along the border between southeastern Colorado and Oklahoma, slight southward thinning occurs. North of the Apishapa uplift a northwest-trending trough, conspicuous on the interval C isopach map (pl. 7A), is not apparent on the map for interval D (pl. 8A).

Minor irregular thicknesses of interval D northeast of the Apishapa uplift probably are the result of thinning of the interval above buried fault blocks. The interval is beveled by erosion and is overlapped southward by interval E in a narrow band on the north edge of the east end of the Apishapa uplift, and similar overlap probably is present along the east flank of the Apishapa-Sierra Grande uplift. Representatives of interval D are present in the Fountain Formation along the eastern boundary of the Front Range, but no thickness assignment is made because of the lithologic similarity with intervals E and C.

In general, thickness trends of interval D indicate less local basin filling and more blanketlike deposition than is represented in older Pennsylvania intervals, particularly interval C.

LITHOFACIES TRENDS

Lithofacies patterns in interval D are similar to those of interval C, except for some shifts in the position of facies boundaries (pl. 8B). Sandstones and conglomerates, mainly arkosic, are predominant in the area east of the east margin of the Frontrange uplift and in the northern part of the Canon City embayment. Fine-grained detrital rocks form a prominent northeast-trending band extending from southeasternmost Colorado and the southwest corner of Kansas northwestward past the east end of the Apishapa-Sierra Grande uplift and toward the south end of the Front Range. In northeastern Colorado, along the Nebraska border, detrital rocks are locally dominant. In the remainder of eastern Colorado, interval D consists mainly of brown to gray locally oolitic limestone that in places contains zones of brown or gray chert.

Comparison of the lithofacies map of interval D with that of interval C shows that during the time of interval D deposition the landward margin of dominantly carbonate deposits advanced toward the Pennsylvanian uplifts. This shift suggests that the supply of detrital sediments from these uplifts must have decreased during that time.

North of Owl Canyon (loc. 79), interval D consists mostly of light-brown to moderate-reddish-brown fine-grained sandstone containing scattered medium to coarse sand grains. Sandstone of this type increases at the expense of arkosic sandstone and conglomerate toward the Wyoming border (Maughan and Wilson, 1960, fig. 3).

As shown in cross section (pl. 10D, E), the thick units of limestone interstratified with thin sets of fine-grained sandstone and mudstone that characterize interval D in easternmost Colorado interfinger westward with strata consisting entirely or almost entirely of fine- to coarse-grained detrital rocks. This interfinger occurring in areas adjacent to the Frontrange and Apishapa-Sierra Grande uplifts. The only major detrital unit in interval D in areas where the interval consists mainly of limestone is a unit of interstratified sandstone, mudstone, and limestone, as much as 50 feet thick, that marks the base of the interval in southeastern and east-central Colorado.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Source areas and depositional environments remained about the same during deposition of interval D as they had been during deposition of interval C, but the depositional patterns shifted.

The Frontrange and Apishapa-Sierra Grande uplifts were the principal sources of detritus. Lithofacies patterns (pl. 8B) indicate that the southern part of the Frontrange uplift was the principal source of coarse detritus. The quantity of detrital sediments produced from the uplift was smaller than in earlier Pennsylvanian time, and carbonate rocks were deposited closer to the east margin of the Frontrange uplift (pl. 8B). The Apishapa-Sierra Grande uplift apparently was the source of fine-grained detrital rocks during the time of deposition of interval D, but the quantity was less than in interval C.

Transport directions were eastward from the Frontrange uplift, as indicated by lithofacies patterns and dip direction of cross strata in the Fountain Formation east of the present Front Range. Transport was both northward and eastward from the Apishapa-Sierra Grande uplift, as determined from lithofacies patterns. The extensive positive element active in eastern Nebraska in previous intervals probably was not a major source of sediment for interval D in eastern Colorado but may have served as a source for some fine detritus.

Interval D, like interval C, was deposited partly as alluvial fans adjacent to uplift areas; this fluvial environment graded eastward and northward from the positive areas into transitional continental-marine environments, and farther east and north into a normal marine environment, as indicated by thick carbonate rocks and by the presence of marine fossils in eastern Colorado. Oolitic and bioclastic limestone and thin sets of brown fine-grained detrital rock interstratified with carbonate sequences are indicative of fairly shallow marine waters that had moderately high energy and free circulation.

The alluvial-plain and alluvial-fan deposits consist principally of the coarse-grained arkosic sandstone and
conglomerate that characterize interval D in outcrops along the east flank of the present Front Range, particularly along the southern margins of the range. Inter­
tonguing of carbonate units with fine- and coarse­
grained detrital units in the northern part of the Front
Range (pl. 10D, E, secs. X-X', Y-Y', Z-Z') indicates
periodic encroachment of the sea from the east.

In southeastern Colorado, fine-grained detrital rocks
probably were formed from shallow-water marine
deposits, except possibly those strata adjacent to the
Apishapa-Sierra Grande uplift which may have been
shoreline or lagoonal deposits.

PALEOTECTONIC IMPLICATIONS

Interval D was a time of considerable local uplift in
eastern Colorado although tectonism was less than dur­
ing interval C time. The linear trough north of the
Apishapa uplift, clearly present in interval C, had filled
by the beginning of interval D. The Hugoton embayment
of the Anadarko basin in southeasternmost Colorado
also was not as apparent a feature in interval D as in in­
terval C. As a result, interval D formed a blanket of sedi­
ment of rather uniform thickness, as shown by isopach
patterns (pl. 8A).

Both the Frontrange and the Apishapa-Sierra Grande
uplifts are inferred to have formed smaller sources of sedi­
ment for interval D than for interval C. Evidence for this
decrease is the westward advance of the carbonate
deposits (pl. 8B) toward the Frontrange uplift and the
southward and westward advances toward the Apishapa
uplift. Furthermore, the Apishapa-Sierra Grande uplift
appears to have been predominantly a source of fine
rather than coarse detrital sediment, so it is inferred to
have been a low positive element at this time. Faults
that border this uplift and small fault blocks east of it
may have continued to be active during interval D,
although evidence for their movement is not clear. The
absence of interval D, inferred for a small area im­
mediately east of the Freezeout Creek fault on the east
flank of the Apishapa uplift, is considered evidence of
uplift in this region during interval D time and prior to
deposition of interval E.

INTERVAL E
FORMATIONS INCLUDED

Strata assigned to interval E in eastern Colorado con­
sist of the uppermost part of the upper member of the
Fountain Formation from the Lyons area northward to a
point north of Redstone Creek and the Pennsylvanian
part of the Fountain Formation (which also contains
rocks of Permian age) from the Lyons area southward
into the Canon City embayment. Rocks in the subsur­
face of eastern Colorado assigned to interval E consist of
unnamed and undivided rocks of Virgil age (pl. 13).

In parts of southeastern Colorado, interval E is com­
posed of the Pedee, Douglas, Shawnee, and Wabaunsee
Groups in ascending order. These group names have
been extended from Kansas on the basis of physical cor­
relation (Maher, 1958). (The Kansas Geol. Survey now
regards strata formerly assigned to the Pedee Group as
part of the Douglas Group.) The Wabaunsee may once
have extended far beyond its present limits but seems to
have been removed by pre-Permian erosion from
northeastern Colorado. The Pedee and Douglas Groups
which form a detrital unit 30-50 feet thick are present
only locally in southeastern Colorado, where they occur
between carbonate rocks of the Lansing and Kansas City
Groups of interval D and the overlying Shawnee Group
of interval E.

North of Lyons, rocks in the upper member of the
Fountain Formation that are assigned to interval E con­
sist of interstratified medium- to coarse-grained arkosic
sandstone and conglomerate and fine to very fine grained
relatively well sorted sandstone. South of Lyons, where
the Fountain is both Pennsylvanian and Permian in age,
strata believed to be equivalent to those north of Lyons
are composed of similar rocks but also include thin sets
of sandy, clayey siltstone. Assignment of these rocks to
interval E is based on physical correlation and, in part,
on projection of thickness trends from eastern Colorado,
where age dating of interval E is better established. Out­
crops of the Fountain Formation have no assigned
thicknesses or boundaries for interval E.

The strata assigned to interval E throughout much of
the subsurface in eastern Colorado are placed there on
the basis of both fossils and lithology. Fusulinids of
Virgil age, consisting of several species of Triticites, have
been reported from 26 drill holes used as control points in
eastern Colorado.

UPPER BOUNDARY OF INTERVAL E

Everywhere that interval E is recognized in eastern
Colorado, it is overlain by rocks of Wolfcamp age.
However, the nomenclature applied to the Lower Per­
mian rocks and the criteria used in determining the
boundary between the Pennsylvanian (interval E) and
the Permian differ from area to area. In northern
Colorado, the upper boundary of interval E is an uncon­
formity that bevels progressively lower beds of the inter­
val northward and westward. This beveling accounts for
the absence of interval E beneath Permian rocks along
the east flank of the Front Range in northern Colorado.
In southeastern Colorado, evidence of an erosional un­
conformity is less clear; interval E is presumed to be con­
formable with the overlying Permian.

The Ingleside Formation overlies interval E in out­
crops along the Front Range from the northern limit of
the interval southward to the Dry Creek area (loc. 754)
(Maughan and Wilson, 1960, fig. 3). In this area, in­
terstratified arkosic sandstone and conglomerate and
fine-grained feldspathic sandstone in interval E (Fountain Formation) are overlain directly by orange fine to very fine grained cross-stratified quartzose sandstone that forms a prominent cliff in the Ingleiside.

The Fountain and Ingleiside Formations appear to be conformable at Dry Creek, and they may be conformable as far north as Redstone Creek. However, thinning of the upper member of the Fountain Formation between Dry Creek and Redstone Creek (Maughan and Wilson, 1960, fig. 3) probably is accounted for by erosion of the uppermost part of the Fountain Formation prior to deposition of the Ingleiside. Sandstone dikes, composed of sand from the Ingleiside Formation, penetrate the uppermost beds of the Fountain Formation to a depth of several feet and are found at several localities, including Big Thompson Canyon (loc. 756). These sandstone dikes constitute additional evidence of an erosional unconformity.

South of Dry Creek the Ingleiside Formation interfingers with the Fountain Formation (Thompson and Kirby, 1940; Maughan and Wilson, 1960). Beyond the southernmost tongue of the Ingleiside south to the vicinity of Colorado Springs (loc. 767), the upper boundary of interval E lies within a sequence of interstratified arkosic sandstones, arkosic conglomerates, and feldspathic sandstones that constitute the Fountain Formation and that include some Permian rocks. In this area, evidence of an erosion surface within the Fountain in absent, and rocks of Virgil age forming interval E are interpreted as conformable with overlying strata of Wolfcamp age.

At Colorado Springs and south along the Front Range as far as Deadman Canyon (loc. 761), the upper boundary of interval E is placed at the base of a unit of cross-stratified and predominantly fine grained light-colored sandstone, from a few to 50 feet thick, that overlies arkosic strata of the Fountain Formation with apparent conformity. The Fountain Formation in this area is dominantly reddish-brown arkosic sandstone and conglomerate and contains small amounts of feldspathic sandstone and sandy clayey siltstone.

At Colorado Springs, the light-colored sandstone unit above the interval E boundary is mapped as the lower part of the Lyons Sandstone (McLaughlin, 1947) although it is separated from the main body of the Lyons by a unit of arkosic sandstone and mudstone. At Deadman Canyon this sandstone unit is thinner, is considered part of the Fountain Formation, and is about 300 feet below the base of the Lyons Sandstone. The sandstone unit has not been identified north of Colorado Springs and has not been traced more than a few miles south of Deadman Canyon; it is believed to be lenticular. In the Colorado Springs area, nevertheless, the base of this sandstone is a persistent horizon that is arbitrarily chosen as the top of interval E.

In the subsurface of eastern Colorado, various criteria are used to identify the upper boundary of interval E. Some biostratigraphic control is available and has been used locally. Lithologic criteria, however, must be used in most places. In northeastern Colorado, both interval E and directly overlying strata of Wolfcamp age are dominantly carbonate rocks. The top of interval E in this area generally is placed at an upward change from rocks that are dominantly limestone to rocks that are dominantly dolomitic limestone or dolomite. Locally, also, the Permian strata can be recognized by the presence of anhydrite and thin detrital beds.

In areas of southeastern Colorado where the Wabaunsee Group is recognized, the top of the Wabaunsee is considered the top of interval E. This horizon generally is located either on the basis of projection of the boundary from established sections in Kansas or on the basis of biostratigraphic control, where this is available. Generally, the top unit of the Wabaunsee is a limestone, 10-30 feet thick, which is overlain by a thicker mudstone sequence. Locally, the mudstone contains anhydrite and sandstone.

In parts of southeastern and east-central Colorado where the Wabaunsee is not recognized, the upper boundary of interval E is placed at the top of the highest limestone bed in a dominantly carbonate rock sequence that underlies a thick sequence of detrital rocks assigned a Wolfcamp age.

In predominantly detrital sections near the Apishapa–Sierra Grande uplift and the southern part of the Frontrange uplift, projection of an established surface is used to extend the upper boundary of interval E. The boundary at some places is at the top of the highest sequence of arkosic rocks (loc. 795); in others, at the base of a fine-grained sandstone bed overlying a mudstone sequence (loc. 311); and in still others, arbitrarily within a mudstone sequence several hundreds of feet thick (loc. 796). Selections of the upper boundary of the Pennsylvanian made on well logs of the American Stratigraphic Co. were used for the top of interval E at many localities, but for some places the boundary was adjusted by the author. The upper boundary of interval E in general becomes more tenuous and difficult to trace near the Pennsylvanian uplifts.

**THICKNESS TRENDS**

Thickness trends of interval E are similar to those of interval D. Interval E thickens gradually northward from more than 600 feet in southeastern Colorado to 100 feet in northeastern Colorado. The interval also thickens southward from its point of maximum thickness in southeastern Colorado to about 400 feet along the Oklahoma border.

Interval E is believed absent as a result of pre-Permian erosion along the Front Range from Dry Creek to the...
Wyoming border and in a small area in the subsurface to the east of the Front Range in the same area. It is present, but is not assigned a specific thickness, in outcrops of the Fountain Formation along the Front Range south of Dry Creek. Interval E pinches out along the flanks of the Apishapa uplift but may extend across part of the Sierra Grande uplift in northeasternmost New Mexico and in part of southeastern Colorado (pl. 9A). The only depositional basin outlined by the isopachs is a shallow, poorly defined basin northeast of the Apishapa uplift.

**LITHOFACIES TRENDS**

Lithofacies of interval E closely resemble the lithofacies of interval D; depositional patterns that prevailed during interval D time continued virtually unchanged.

Carbonate rocks, largely limestone, are predominant in northern and eastern Colorado but are interstratified with thin to thick sets of siltstone and fine-grained sandstone, and they grade laterally into fine-grained detrital rock in southeastern Colorado (pl. 9B). The interval is mainly sandstone and coarser detrital rock in a relatively narrow band bordering the east flank of the Front Range and in the Canon City embayment.

In northern Colorado, carbonate rock grades and intertongues westward into arkosic sandstone and conglomerate. Near the Front Range sandstone is feldspathic, and only a small amount of fine-grained detrital rock is present. In southeastern Colorado, near the Colorado-Kansas border, the dominant limestone grades and intertongues westward into mudstone.

Mudstone surrounds the eastern and northern borders of the Apishapa uplift. The amount of fine detritus introduced into southeastern Colorado apparently was somewhat greater during interval E time than during the time of earlier intervals, inasmuch as a small eastward shift in the boundary between predominantly carbonate rocks and predominantly detrital rocks is apparent in strata of interval E.

In southeastern Colorado, in places where the Pedee, Douglas, Shaawnee, and Wabaunsee Groups are differentiated, the Pedee and Douglas are a thin dominantly detrital unit. The Shaawnee Group consists primarily of brownish to gray limestone interstratified with thinner units of red to gray mudstone; the Wabaunsee Group, locally represented, consists of interstratified brown to gray limestone and red mudstone and fine to very fine grained sandstone. The amount of detrital components in most of the Wabaunsee Group is greater than that in the Shaawnee Group, and in places, the detrital units in the Wabaunsee are predominant.

**TOTAL THICKNESS OF PENNSYLVANIAN ROCKS**

**THICKNESS TRENDS**

During most of Pennsylvanian time the eastern Colorado region was bordered on the west and southwest by two prominent uplifts—the Front Range to the west and the Apishapa–Sierra Grande to the southwest. These uplifts were the principal sources of detrital sediments deposited in adjacent basins and in shelf areas beyond.

Areas of maximum deposition of Pennsylvanian strata are (1) the Colorado Springs area, where Pennsylvanian rocks are in excess of 4,400 feet thick, (2) a conspicuous northwest-trending linear trough north of the northern border of the Apishapa uplift containing over 3,000 feet of Pennsylvanian strata, and (3) Baca County in southeasternmost Colorado, where thicknesses differ
greatly from place to place but are in excess of 2,600 feet in two local areas (locs. 255, 252). Thickness trends are irregular.

Pennsylvanian rocks thin gradually from areas of maximum thickness northward and northeastward to areas in Weld and Logan Counties of northern Colorado where the Pennsylvanian is between 850 and 1,000 feet thick in the subsurface (pl. 11). Locally, in outcrops on the east flank of the Front Range in Larimer County, northern Colorado, Pennsylvanian strata are less than 600 feet thick.

Pennsylvanian strata thin markedly toward the south from an axis of maximum deposition that lies north of the Apishapa uplift and toward the west in the Baca County area. Pennsylvanian rocks wedge out on the flanks of the Apishapa-Sierra Grande uplift, and they thin southward in Baca County. They are less than 2,000 feet thick at four control points in southern Baca County on the Colorado-Oklahoma border.

Three linear structures are shown clearly by the isopachs on the total thickness map for the Pennsylvanian System (pl. 11). One of these structures is a northwest-trending linear trough north of the Apishapa uplift. A second is the northwest-trending Apishapa uplift itself. The third is a north-trending arch in eastern Yuma County near the Kansas-Nebraska border; the crest of this arch is covered by less than 1,000 feet of Pennsylvanian strata.

A fourth structure, less prominent than the others, is the south-plunging Red Creek arch in southwestern El Paso County, south of Colorado Springs. Less than 1,500 feet of Pennsylvanian strata were deposited on its crest.

PALEOTECTONIC IMPLICATIONS

The northward thinning of Pennsylvanian strata across eastern Colorado is, in part, the result of different rates of subsidence. Downwarp generally was less in northern Colorado than it was in the south during all intervals. Also, at the end of Pennsylvanian time, uplift and consequent removal by erosion of some, and locally all, of interval E in northern Colorado further accounts for northward thinning of Pennsylvanian strata.

The zero isopach of total Pennsylvanian strata along the north and east sides of the Apishapa-Sierra Grande uplift is the same as that of interval A (pl. 3A); none of the other Pennsylvanian intervals extended as far up the flanks of the uplift. Interval A is interpreted as having been deposited prior to uplift of the Apishapa-Sierra Grande area and as having originally extended across it. The zero isopach for the Pennsylvanian, therefore, is the result of post-interval A erosion during the Middle and Late Pennsylvanian when the Apishapa-Sierra Grande element was uplifted.

The northwest-trending linear trough north of the Apishapa uplift represents a yoked basin or fault trough, the movement of which apparently was coupled with movement of the Apishapa uplift. The time of maximum uplift of the Apishapa (interval C) was also the time of maximum sinking and sedimentation in this trough (pl. 7A).

Irregular thickness patterns in Baca County east of the Apishapa-Sierra Grande uplift and in southern Bent County north of the uplift reflect movement along small faults and fault blocks at various times during the Pennsylvanian Period. The areas of thin Pennsylvanian strata probably correspond to crests of upthrown fault blocks; the areas of thick Pennsylvanian, to grabens.

Original limits of the Front Range uplift are not readily apparent on plate 11. Margins of the uplift lay west of the present limits of Pennsylvanian strata along the Front Range. The only indications of its former position are (1) the irregular thickness patterns along the Front Range in northern Colorado, which probably result from burial of uneven topography along the flanks of the uplift, and (2) the Red Creek arch in southwestern El Paso County that probably was the southeast terminus of the uplift.

The north-trending arch in eastern Yuma County was active in Early and Middle Pennsylvanian time and resulted either in nondeposition or in deposition and later erosion of interval B, and locally of interval A, in this area (pl. 6A, 3A). Isopachs of later Pennsylvanian intervals do not suggest this feature.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

UNITS OVERLYING PENNSYLVANIAN

Throughout eastern Colorado, with the possible exception of local outcrop areas in the Canon City embayment and on the northeast flank of the Wet Mountains, Pennsylvanian strata are overlain by Permian rocks of Wolfcamp age (pl. 12). Basal Permian units consist of the Admire Group and undifferentiated rocks of Wolfcamp age in the subsurface of southeastern, east-central, and northeastern Colorado and the uppermost Permian part of the Fountain Formation in both the outcrop and the subsurface of the Canon City embayment. They also include the uppermost part of the Fountain Formation along the east flank of the Front Range north of the Canon City embayment as far as the Lyons area and the Ingleside Formation from the Lyons area north to the Wyoming border.

At two localities (locs. 298, 764) in outcrops of the Wet Mountains, Upper Jurassic rocks rest unconformably on the Fountain Formation, which at those localities is believed to be entirely Pennsylvanian, a Permian portion
probably having been removed by pre-Jurassic erosion. At Canon City (loc. 763), the Fountain Formation, interpreted as entirely Pennsylvanian at that locality, is overlain unconformably by the Lykins Formation of Permian and Triassic (?) age. The Fountain Formation is relatively thin (1,020 ft) at Canon City; the Permian portion of the formation, if ever present, was removed during a period of pre-Leonard erosion.

In the southern part of eastern Colorado, Permian rocks rest upon the Pennsylvanian with apparent conformity. In outcrops of northeastern Colorado, in contrast, an erosional unconformity is present between the Ingleside Formation and the underlying beveled strata of the Fountain Formation. This unconformity is believed to be present also in the subsurface of northeastern Colorado.

Erosion of interval E during development of the overlying unconformity probably contributed to a relatively abrupt northward thinning of the interval throughout eastern Colorado — a feature that contrasts with trends in the older Pennsylvanian intervals (pl. 10D, sec. W-W’) and that accounts for the absence of interval E along the Front Range in northernmost Colorado. The erosional unconformity has not been recognized in outcrops south of Redstone Creek (loc. 755). In the subsurface the unconformity is inferred, on the basis of an increase in the northward rate of thinning of interval E, to extend southward approximately to the northern border of Cheyenne County.

PALEOTECTONIC IMPLICATIONS

In southern Colorado, subsidence and deposition probably was continuous from Pennsylvanian into Permian time. Permian rocks overlap the Pennsylvanian around the borders of the Apishapa–Sierra Grande uplift and extend across the uplift to the Wet Mountains farther west. By the end of Early Permian time, the uplift was buried by Permian sediments (Mudge, 1967, p. 102, 107). Thus, movement of the Apishapa–Sierra Grande uplift had virtually ceased or had been reversed by the Early Permian.

In northern Colorado, mild but broad upwarping of Pennsylvanian strata and very slight tilting of these beds to the south caused erosion and beveling prior to Permian deposition. The hiatus represented by this unconformity was not great, as Lower Permian (Wolfcamp) beds everywhere overlie the uppermost Pennsylvanian.

REFERENCES


——, 1947, Subsurface geologic cross section from Scott County, Kansas, to Otero County, Colorado: Kansas Geol. Survey Oil and Gas Inv. Prelim. Cross Section 4, 11 p.


Mudge, R. M., 1967, Central Midcontinent region, chap. F, in
Middle and Southern Rocky Mountains, Northern Colorado Plateau, and Eastern Great Basin Region

By WILLIAM W. MALLORY

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-N
CONTENTS

Abstract ........................................... 265
Region defined ..................................... 265
Paleogeology ........................................ 265
Units underlying Pennsylvania ..................... 265
Lower boundary of Pennsylvania .................... 265
Paleotectonic implications ......................... 265
Interval A ........................................... 266
Formations included ............................... 266
Upper boundary of interval A ....................... 266
Thickness trends ................................... 266
Lithofacies trends ................................ 266
Sources and environments of deposition .......... 269
Paleotectonic implications ....................... 269
Interval B ........................................... 269
Formations included ............................... 269
Upper boundary of interval B ....................... 269
Thickness trends ................................... 270
Lithofacies trends ................................ 270
Sources and environments of deposition .......... 270
Paleotectonic implications ....................... 270
Interval C ........................................... 270
Formations included ............................... 270
Upper boundary of interval C ....................... 271
Interval C—Continued ................................ 
Thicknes trends .................................... 271
Lithofacies trends .................................. 271
Sources and environments of deposition .......... 272
Paleotectonic implications ....................... 273
Interval D ........................................... 273
Formations included ............................... 273
Upper boundary of interval D ....................... 273
Thickness trends ................................... 273
Lithofacies trends ................................ 274
Sources and environments of deposition .......... 274
Paleotectonic implications ....................... 274
Interval E ........................................... 275
Formations included ............................... 275
Upper boundary of interval E ....................... 275
Thickness trends ................................... 275
Lithofacies trends ................................ 276
Sources and environments of deposition .......... 276
Paleotectonic implications ....................... 276
Total thickness of Pennsylvanian rocks .......... 276
Thicknes trends and paleotectonic implications 276
Geologic units directly above Pennsylvanian System 277
References .......................................... 277

ILLUSTRATIONS

[For listing of plates (in separate case) see volume “Contents”]

Figure 61. Geographic features in Middle and Southern Rocky Mountains, northern Colorado Plateau, and eastern Great Basin region mentioned in text .......................... 267
62. Pennsylvanian and post-Pennsylvanian structural features in Middle and Southern Rocky Mountains, northern Colorado Plateau, and eastern Great Basin region mentioned in text ........................................ 268

TABLES

Table 28. Formations assigned to interval A ......................... 266
29. Formations assigned to interval B ......................... 269
30. Formations assigned to interval C ......................... 271
31. Formations assigned to interval D ......................... 273
32. Formations assigned to interval E ......................... 275
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES.
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

MIDDLE AND SOUTHERN ROCKY MOUNTAINS, NORTHERN COLORADO PLATEAU, AND EASTERN GREAT BASIN REGION

By William W. Mallory

ABSTRACT

Lower and lower Middle Pennsylvanian strata are irregular in thickness and distribution in the region; new tectonic patterns were more complex than any previous ones of the Paleozoic Era. By Des Moines time Pennsylvanian tectonism had reached a climax. In southern Wyoming, eastern Utah, and western Colorado, the Ancestral Rocky Mountains, a chain of actively rising uplifts, separated rapidly sinking basins. In northern Wyoming a broad stable shelf was the dominant tectonic feature. In Late Pennsylvanian time, uplift and subsidence diminished. In southeastern Idaho and western Utah, marked subsidence took place in the Cordilleran geosyncline.

Rocks in the basin areas are highly variable in composition but on the Wyoming shelf are relatively constant in composition. Lower and lower Middle Pennsylvanian rocks consist of relatively thin tongues of mudstone and carbonate rocks with fringing sandstones. Des Moines rocks in the Eagle and Paradox basins are thick lenses of halite, anhydrite, gypsum, and other salts that interfinger with coarse red beds in belts marginal to the Ancestral Rocky Mountain uplifts. The Paradox basin also contains dolomite and black shale. On the Wyoming shelf unusually well sorted sandstone is widespread.

In Late Pennsylvanian time much of the Wyoming shelf was emergent, and sandstone deposits on the shelf margins graded eastward into carbonate deposits in deeper waters. The Ancestral Rocky Mountains shed large volumes of red detritus into the Paradox and Eagle basins.

The Oquirrh basin in northwestern Utah and southeastern Idaho accumulated very thick deposits of sandstone, siltstone, limestone, and mudstone throughout the period. Intrageosynclinal uplift and subsidence related to the Antler orogeny probably occurred in southeastern Idaho.

REGION DEFINED

The region comprises the following areas: All of Wyoming, except the extreme northwest corner and Campbell, Crook, Weston, Niobrara, Goshen, and Laramie Counties in eastern Wyoming; all of Colorado west of the Rocky Mountain front; all of Utah; and that part of Idaho east of the 114th meridian.

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

The rock units mapped on plate 2 consist of the following:

Rocks of Mississippian age:
- Darwin Sandstone Member of the Amsden Formation, of Chester age, in central and western Wyoming, and Madison Limestone, of Osage and Meramec age, elsewhere in Wyoming
- Doughnut Formation, of Chester age, in northeastern Utah
- Horseshoe Mesa Member of the Redwall Limestone, of Meramec age, in south-central Utah
- Leadville Limestone, of Osage age, in western Colorado

Rocks of pre-Mississippian age:
- Isolated small areas of Devonian, Cambrian, and Precambrian rocks.

LOWER BOUNDARY OF PENNSYLVANIAN

The lower boundary of the system, except where the Darwin Sandstone Member and Doughnut Formation are present, is a disconformity generally marked by sandstone overlying limestone or dolomite. In many areas the disconformity has topographic relief that is measurable in scores of feet. Where the Darwin Sandstone Member and Doughnut Formation of Mississippian age are present, these strata commonly are gradational into the overlying Pennsylvanian rocks.

PALEOTECTONIC IMPLICATIONS

Lowering of sea level in the region at the beginning of Pennsylvanian time may have resulted from widespread epeirogenic upwarping. A marine withdrawal and depositional hiatus which preceded Pennsylvanian sedimentation in many areas began in late Meramec time and continued throughout Chester time. In much of western Wyoming and in northeastern Utah, a marine advance began during Chester time; rocks of this age, including the Darwin, lie on a surface of disconformity and grade upward into rocks of Pennsylvanian age. Seas continued to advance in Pennsylvanian time, and successively younger rocks of Pennsylvanian age lie disconformably on rocks of Meramec and older age.
INTERVAL A
FORMATIONS INCLUDED

The nomenclature for rocks in all parts of the region except the Sangre de Cristo Mountains (fig. 61) in south-central Colorado is shown in table 28 and on plate 13. Terminology and stratigraphic correlations are not completely standardized. For instance, the names Kerber, Sharpsdale, Minturn, and Sangre de Cristo are in use for parts of the Pennsylvania System, but assignment of age by epoch and correlation with other strata are tentative.

In much of the region, interval A is equivalent to rocks of Morrow age, as determined by fossils. In many areas lithologic criteria, in addition to paleontologic data, are helpful in establishing a mappable unit. In southwestern Utah, for example, fusulinids were used by Brill (1963, p. 313) to delimit a unit of Morrow age, but the Hall Canyon Member of the Oquirrh Formation was referred to the Morrow by Bissell (1962, p. 26-28) on the basis of a distinctive micritic to biolastic texture which he said is recognizable throughout Oquirrh basin (fig. 62). In Idaho, interval A is not easily defined, although some paleontological assistance is available. In northeastern Utah and northwestern Colorado, identification of interval A within the red mudstone and gray limestone of the lower part of the Morgan Formation is dependent on fusulinids (Thompson, 1945). Eastward, the lower part of the Morgan Formation grades into the dark Belden Shale of both Morrow and Atoka age. The boundary between rocks of these ages in the Belden is obscure; hence, interval A has arbitrarily been defined in the vicinity of the Sawatch Range (fig. 62) as the lower half of the Belden Shale.

In the vicinity of the Laramie and the Medicine Bow Mountains, southern Wyoming, the lower part of the Casper Formation has been assigned to interval A on the basis of dating by fusulinids from surface sections on the flanks of the Laramie Mountains. In central-western Wyoming, the Horseshoe Shale Member of the Amsden Formation composes interval A.

UPPER BOUNDARY OF INTERVAL A

Nowhere in the region is the upper boundary of interval A clearly and easily distinguishable except in central-western Wyoming, where cherty limestone of the Ranchester Member of the Amsden Formation in interval B lies on red mudstone of the Horseshoe Member in interval A. In Utah, according to Bissell (1962, p. 28-29), bioclastic and micritic limestone in the Hall Canyon Member of the Oquirrh Formation is overlain gradationally by bioclastic limestone, sandy limestone, and sandstone in the Meadow Canyon Member of Atoka age. In other formations, the separation is made on a faunal basis.

THICKNESS TRENDS

Interval A attains its greatest thickness of this region in northwestern Utah, where a maximum of about 1,500 feet is present within the Oquirrh basin southeast of Great Salt Lake (pl. 3A).

Two troughs extended from the Oquirrh basin eastward into the Rocky Mountain region (fig. 62). One of these, the Central Colorado trough, was located in northeastern Utah and northwestern Colorado (the Eagle, or Maroon, basin) and turned southeastward in central Colorado, where its trend is interrupted by Laramide uplifts of the Southern Rocky Mountains. A second trough, for convenience here called the Sweetwater trough, was in southwestern Wyoming and extended in a northeastward direction (fig. 62). Average thickness of interval A in the Central Colorado trough is about 300 feet; in the Sweetwater trough of southwestern Wyoming average thickness is about 200 feet. On the Wyoming shelf, isopachs form irregular patterns showing no well-defined trends.

LITHOFOACIES TRENDS

In western Utah and southeastern Idaho, interval A, as represented by the Callville, Ely, Hall Canyon Member (of Bissell, 1962) of the Oquirrh, and Wells Formations, is composed primarily of carbonate rock (pl. 3B). It probably was deposited in an open seaway that extended northward across Utah into eastern Idaho. In that part of Idaho, the carbonate facies grades into sandstone and mudstone of the Wood River and Copper Basin Formations. Sandstone carbonate rock composes the Topache Formation of southwestern Utah.

The Morgan Formation in northeastern Utah and northwestern Colorado contains limestone interbedded with sandstone and red mudstone. It grades
southeastward into the dark Belden Shale of the Eagle basin, northward into the Horseshoe Shale Member of the Amsden Formation, and northeastward into sandstone and limestone of the Casper Formation.
Locally in the northern part of the Sangre de Cristo Mountains, the Kerber Formation is dominantly sandstone but contains some dark mudstone and, locally, beds of shaly coal.
SOURCES AND ENVIRONMENTS OF DEPOSITION

Sandstone included in the Topache Formation seems to have been derived from the large land area in southeastern Utah and southwestern Colorado and possibly was formed as beaches and bars which transgressed landward. The Morgan Formation and Round Valley Limestone were deposited in a wide bay formed by the junction of the Central Colorado and Sweetwater troughs. Mudstone interbeds in the Morgan probably extended from areas of dominant mudstone deposition farther east and north. The dominance of dark-gray mudstone and dark limestone and the paucity of sandstone in the Belden Shale indicate that surrounding land was low and that restricted conditions in the Eagle basin created an euxinic environment. Coal in the Kerber Formation may represent a local swampy, landward expression of the shallow Central Colorado trough, where the depositional environments alternated locally among beach, offshore bar, and fluviatile types.

Red mudstone of the Horseshoe Member of the Amsden forms a tabular layer that was deposited on a broad, highly stable shelf. The color may have been derived from lateritic weathering in the source area or from erosion of red rocks of a Precambrian terrane possibly in the Lake Superior region. A hematite nodule zone in the Horseshoe Member suggests an iron-rich provenance in the area of deposition. Environment and source of the Casper Formation are not clear. The sandstone patterns on the facies map (pl. 3B) do not suggest any nearby source; hence, possibly the same distant source that supplied sand to the Darwin Sandstone Member contributed sand to the Casper. Abundant limestone, the absence of arkose, and the lithofacies trends which are at an angle to the trend of uplifts in the vicinity of the Laramie Mountains indicate that the Casper was deposited on a broad shelf before local uplift in late Morrow time (Mallory, 1967). The Molas Formation is a red regolithic deposit derived by lateritic weathering in the Paradox basin area before the marine invasion during interval C time.

PALEOTECTONIC IMPLICATIONS

Absence of interval A strata throughout all of southeastern Utah and most of western Colorado suggests that those regions were emergent during Morrow time and that marine rocks of Early Pennsylvanian age were never deposited there. Widespread presence of the Molas Formation, a red regolithic deposit derived largely from the weathering of Mississippian limestone, corroborates this inference. Elevation of the emergent region must have been moderate to low because coarse detrital material is scarce in strata on its margins. Presence of sandstone in south-central Colorado in the Kerber Formation implies that somewhat more pronounced uplift took place in Morrow time on the Apishapa uplift in the vicinity of the present Wet Mountains than elsewhere in Colorado.

Near the Laramie Mountains in southeastern Wyoming, the alignment, thickness, and lithofacies trends at the south limit of the interval suggest post-Morrow uplift. An irregular area in western Wyoming and eastern Idaho may have been emergent during or immediately after interval A time. The presence of coarse detrital material in the Wood River Formation in south-central Idaho indicates uplift in that vicinity (Roberts and Thomasson, 1964, p. D6).

INTERVAL B
FORMATIONS INCLUDED

The formations included in interval B in various parts of the Middle and Southern Rocky Mountains, northern Colorado Plateau, and eastern Great Basin are listed in table 29.

UPPER BOUNDARY OF INTERVAL B

The boundary between intervals B and C as determined on the basis of fusulinid zones is in most places the contact between rocks of Morrow and Atoka ages. In some areas where a lithologic change occurs at or near the time boundary, the lithologic change is used for the contact, and it provides a reasonably accurate means for tracing the series boundary on the surface and in the subsurface. For example, in Paradox basin and in part of Eagle basin, carbonate and calcareous shaly strata of interval B are overlain by evaporitic rocks of Des Moines age (interval C). In central and western Wyoming, the Tensleep Sandstone (interval C) contrasts markedly

<table>
<thead>
<tr>
<th>Area</th>
<th>Formations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eastern Idaho</td>
<td>Part of Wells, Copper Basin, and Wood River Formations.</td>
</tr>
<tr>
<td>Southwestern Utah</td>
<td>Part of Topache Formation, and Ely Limestone.</td>
</tr>
<tr>
<td>Northwestern Utah</td>
<td>Meadow Canyon Member of Oquirrh Formation.</td>
</tr>
<tr>
<td>Northeastern Utah</td>
<td>Part of Morgan Formation.</td>
</tr>
<tr>
<td>Southeastern Utah</td>
<td>Pinkerton Trail Formation of Hermosa Group; part of Molas Formation.</td>
</tr>
<tr>
<td>Northwestern Colorado</td>
<td>Part of Morgan Formation and part of Belden Shale.</td>
</tr>
<tr>
<td>South-central Colorado</td>
<td>Part of Kerber Formation.</td>
</tr>
<tr>
<td>Southeastern Wyoming</td>
<td>Part of Fountain and Casper Formations.</td>
</tr>
<tr>
<td>Northeastern Wyoming</td>
<td>Part of Minnelusa Formation.</td>
</tr>
<tr>
<td>Central Wyoming</td>
<td>Ranchester Limestone Member of Amsden Formation.</td>
</tr>
</tbody>
</table>
with the underlying red-stained cherty and shaly carbonate strata of the Ranche Member of the Amsden Formation. Elsewhere, the boundary between intervals has been determined by paleontologic means alone or by laterally projecting the contact from areas where it has been established on the basis of fossils or lithology.

THICKNESS TRENDS

Interval B is more than 1,000 feet thick in a northwest-trending band along the axis of the Oquirrh basin (pl. 6). In southeastern Utah and southwestern Colorado the interval is as much as 500 feet thick along an arc on the southwestern periphery of Paradox basin. In broad areas both north and east of the basin, the average thickness is about 200 feet. In northeastern Utah and northwestern and central Colorado, the interval is fairly thick along the axis of the northwest-trending Central Colorado trough.

Areas in which interval B is missing include a large region between the Paradox and Eagle basins, and in an irregular linear band north of the Eagle basin that extends along the Wyoming-Colorado line from the central part of the Uinta Mountains to the vicinity of Laramie, Wyoming. Rocks of interval B age are absent on the Pathfinder uplift in southern Wyoming but occur on its flanks and thicken away from it in all directions.

In southwest Wyoming, an area that seemingly lacks interval B strata appears to be the remnant of an emergent area that had been more extensive in interval A time. Interval B is absent in the area between the Paradox and Oquirrh basins in central and southern Utah.

LITHOFACIES TRENDS

In western Utah and in southeastern Idaho, carbonate rock is the dominant lithology, grading in south-central Idaho to coarse sandstone. Wide irregular belts of sandy carbonate rock are present near the margins of the interval in southwestern Wyoming and south-central Utah. Strata on the east and west margins of Paradox basin are mostly those of mudstone that grade into argillaceous carbonate and carbonate strata in the middle of the basin. Similarly, rocks at the margins of the Eagle basin are sandy mudstone.

The linear area in which interval B is absent along the Colorado-Wyoming line is flanked on the north by sandstone and mudstone, and the interval in the Central Colorado trough in south-central Colorado is mostly sandstone. In the Laramie basin of southeastern Wyoming, coarse arkosic conglomerate makes up the interval, and, in many places, shaly and sandy strata along the margins of the Pathfinder uplift interfinger basinward with carbonate rocks. Mudstone and sandy limestone are present west of the Wind River Range, north of an area where interval B is absent. Throughout most of Wyoming, however, carbonate rock is typical of interval B.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Carbonate rock is the dominant lithology of interval B and suggests open-marine conditions, but aprons of detrital strata that flank areas in which interval B is absent indicate that these areas were emergent during interval B time and supplied modest amounts of detrital material to the otherwise clear seas of the region. Along the Colorado-Wyoming-Utah line, a long west-trending projection from the Front Range uplift (spelled as shown to distinguish it from the present Front Range uplift of Laramide age) seems to have shed sand and mud both northward and southward. A broad area that may be considered the proto-Uncompahgre uplift apparently shed sand northward into the Eagle basin and mud southward into the Paradox basin.

In southern Wyoming, land in the area of the Pathfinder uplift shed weathered granitic particles into the Laramie basin and the northern part of the Denver basin. A large emergent area in south-central Utah and northern Arizona, here called the Piute positive element, contains the Emery uplift at the north end. It shed mostly sand westward and mud eastward.

The almost 300-foot thickness of arkosic conglomerate of the Fountain Formation indicates that the Laramie basin area probably was the site of very rapid detrital accumulation.

PALEOTECTONIC IMPLICATIONS

The tectonic patterns of interval B foreshadow the mature development of the Ancestral Rocky Mountains during Des Moines time. By interval B time (Atoka) both Paradox and Eagle basins were in existence and the nearby Uncompahgre uplift was emerging as a separate entity of moderate elevation. The Pathfinder uplift in Wyoming reached its greatest extent and degree of uplifting at this time, and the Piute positive element in south-central Utah was discernible. Sandy and muddy strata on the flanks of the Piute positive element suggest that it supplied some detrital material to adjacent seaways. An elevated area was also possibly present southwest of the Wind River Range, but evidence for this feature is weak. Sandstone in the Wood River Formation in eastern Idaho presumably was derived from an uplifted area farther west which may have been a part of the Antler orogenic belt (Roberts and Thomasson, 1964; fig. 68).

INTERVAL C FORMATIONS INCLUDED

Interval C in the Middle and Southern Rocky Mountains, northern Colorado Plateau, and eastern
Great Basin consists of the formations listed in Table 30. Interval C is made up of rocks mostly or entirely of Des Moines age.

**UPPER BOUNDARY OF INTERVAL C**

In much of western Wyoming, the Tensleep Sandstone is entirely of Des Moines age (Mallory, 1967) and is overlain directly by the Permian Phosphoria Formation. The contact is not difficult to determine, owing to the marked differences in color and lithology of the formations. At any particular location the contact may appear conformable but actually is paraconformable because all of Late Pennsylvanian and part of Permian time are represented by the hiatus.

In eastern and southern Wyoming the contact between intervals C and D is within the Minnelusa and Casper Formations and is located by paleontological dating at the top of rocks of Des Moines age. In northwestern Colorado, the upper boundary can be determined by a combination of paleontological and lithologic factors. The Jacque Mountain Limestone Member is a useful local marker bed at the top of the Minturn Formation. Age of the Minturn is equivocal, but at least the lower five-sixths of the formation is of Des Moines age (Tweto, 1949, p. 205); hence, the Jacque Mountain, defined as the top of interval C, approximates the top of the Des Moines. The age of Eagle Valley Evaporite in the Eagle basin is Des Moines, except for the lowermost part, which is late Atoka; the entire unit is arbitrarily placed in interval C. Farther northwest, that part of the Morgan Formation which contains abundant carbonate rock and which is equivalent to the Eagle Valley is also placed in interval C.

In the Paradox basin of southwestern Colorado and southeastern Utah, the upper boundary of the interval is placed at the top of the main body of the evaporite rocks in most wells. The lithologic change at this horizon does not everywhere mark the top of the Des Moines Series, but it is reasonably close.

**THICKNESS TRENDS**

Marked local variation in thickness is a common feature in interval C (pl. 7A). Three areas exhibit notably thick sections; these are (1) the Eagle basin, where thickness may be as much as 5,000 feet, (2) the Paradox basin, where, in tectonically undistorted areas, thickness may exceed 5,000 feet, and (3) Oquirrh basin, where it exceeds 6,000 feet near Provo but thins in all directions to 1,000–2,000 feet (some of the thickening may be due to repetition by faulting). Rocks assigned to interval C elsewhere in the region are less than 1,000 feet thick. In much of central and western Wyoming, they are less than 500 feet.

Both the Eagle and Paradox basins are elongated northwestward. Subsidiary northeast lineaments cause a series of deep crenulations in isopachs on the southwest flank of Paradox basin. An unusual feature of Paradox basin is a group of topographically and structurally prominent halite diapirs that have a distinct northwest trend. Present thickness of evaporitic rocks of interval C in these diapirs has been so greatly increased by faulting and flowage that the thickness cannot be shown by isopachs on a small-scale map. The maximum measured thickness is more than 10,000 feet. Black areas on plate 7A show where salt-cored anticlines of interval C are exposed at the surface or lie at shallow depth beneath alluvium.

In the Oquirrh basin at Provo, 6,000 feet of interval C is present in an allochthonous slice of Paleozoic rocks. The actual original thickness represented here is not clear and the problem is made difficult by lack of information on the distance and direction of movement of the Charleston thrust (Crittenden, 1961). In the Bighorn basin, a northeast orientation of lobate “thin” areas of deposition is evident.

**LITHOFACIES TRENDS**

Rocks of interval C in the Wyoming-Colorado-Utah region can be grouped into three major lithologic provinces (pl. 7B): (1) the Wyoming shelf, (2) the Cordilleran miogeosynclinal belt, and (3) basins associated with the Ancestral Rocky Mountain uplifts (fig. 62).

The Wyoming shelf covers most of western Wyoming and a part of Utah east of Great Salt Lake. In northwestern and central Wyoming, the rocks consist of unusually well sorted sandstone that is crossbedded and locally contains lenses of sandy carbonate rock and calcareous sandstone. This sheet of sandstone grades west, south, and east through calcareous sandstone and
sandy carbonate rock into carbonate rock. A tongue of sandy carbonate to carbonate rock is in central-southern Wyoming in the area occupied by the Sweetwater trough during both younger and older intervals.

In the Cordilleran miogeosynclinal belt of southeastern Idaho, Wyoming, and Utah, interval C is dominantly carbonate rock in most places, although locally it is composed largely of sandstone. The Wood River Formation in Idaho contains coarse clastic strata.

Interval C in the Paradox basin is a large lens of halite, potash, anhydrite, dolomite, and black shaly mudstone, enclosed within a sheath of anhydrite, dolomitic anhydrite, evaporitic dolomite, reeferid dolomite, and limestone (Wengerd and Strickland, 1954). The thickest part of the lens of evaporite rocks is in the Uncompahgre trough (Hite, 1968, p. 320), near and parallel to the southwest margin of the Uncompahgre uplift and on the east side of the basin. The evaporite rocks thin southeastward and terminate near the New Mexico State line; south of this part of the area, carbonate rocks are dominant.

Lithologic zones can be discerned at the periphery of the Paradox basin evaporite lens along the southwest margin of the basin as shown on plate 7B. Southwest of the Uncompahgre uplift, red arkosic conglomerate and sandstone intertongue basinward with the evaporite rocks. A wide band of sandy limestone borders the east margin of the Piute positive element.

The Eagle basin contains gypsum, anhydrite, and halite. The thickest evaporitic rocks occur in local subsidiary basins east of Glenwood Springs (Mallory, 1971). Northeastward and southwestward, the evaporitic rocks grade into red arkosic conglomerate and sandstone intertongue basinward with the evaporitic rocks. A wide band of sandy limestone borders the east margin of the Piute positive element.

The Eagle basin contains gyspum, anhydrite, and halite. The thickest evaporitic rocks occur in local subsidiary basins east of Glenwood Springs (Mallory, 1971). Northeastward and southwestward, the evaporitic rocks grade into red arkosic conglomerate and sandstone intertongue basinward with the evaporitic rocks. A wide band of sandy limestone borders the east margin of the Piute positive element.

The north end of the Frontrange uplift in southern Wyoming is bordered by a narrow band of arkosic conglomerate and sandstone which grades basinward into sandstone and carbonate rock.

**Sources and Environments of Deposition**

The source of the sand in the Tensleep Sandstone in central Wyoming has long posed a problem. No entirely satisfactory answer is at hand. The alternation of noncalcareous, crossbedded sandstone with calcareous sandstone suggests that the Wyoming shelf was unusually stable and that ebb and flow of marine waters caused by eustatic changes (possibly responsible also for cyclothemes of the Mississippi Valley) allowed the surface of deposition to become intermittently emergent. At these times, wind and waves may have reworked the sand into crossbedded noncalcareous dunes and beaches.

A predominance of carbonate rock in western Utah indicates that this area was distant from strongly active uplifts. Coarse detritus in the Wood River Formation in eastern Idaho probably was derived from highlands to the west in the Antler orogenic belt of central Idaho (Roberts and Thomasson, 1964).

The presence of black mudstone interbedded with the hypersaline rocks of Paradox basin indicates deposition at considerable depth, as deep water would favor undisturbed accumulation of organic-rich muds in an euxinic environment. Rapid downfaulting (or sharp bending) at the basin margin adjacent to the Uncompahgre uplift must have taken place during interval C time (especially early in the interval) to permit the basin floor to subside 5,000 feet while keeping pace with or exceeding rate of deposition.

The alternation of evaporite beds was explained by Hite (1970) to be the result of major changes in sea level. During low sea level, the passageway to the open sea, located to the south, became shallow and permitted only inflowing currents. Deposits of halite and potassium salts in the basin mark these periods. During high sea level, both influx and reflux occurred; anhydrite and dolomite deposits in the basin mark these periods.

Along the southwest margins of the evaporite lens, some of the carbonate strata contain abundant calcareous organic remains. These faunules almost certainly lived on the seaward side of a topographic barrier that is inferred to have surrounded the area of more saline waters.

On the west margin of the Paradox basin, largely unfossiliferous carbonate strata are considered to have been formed by chemical rather than biogenic processes. These unfossiliferous units wedge out westward toward the Piute positive element, which suggests that a low topographic barrier had developed between the evaporite basin and the open sea in that direction (R. J. Hite and O. B. Raup, oral commun., 1966).

On the northeast side of the basin, adjacent to the Uncompahgre uplift, red arkosic conglomerate and sandstone and red mudstone were derived from weathering and erosion of the basement rocks of the then actively rising Uncompahgre uplift. They were deposited as fans marginal to the uplift.

In the Eagle basin, halite was deposited in rapidly subsiding subsidiary troughs where circulation was restricted. Gypsum and(or) anhydrite formed across a wider area in a belt extending from western Rio Blanco County to the vicinity of the Sawatch Range. Local algal bioherms on the margins of the evaporite basin may have helped restrict the circulation in the basin. Saline brines have been obtained from springs and wells at Salt Creek near Antero Reservoir in southwestern Park County, Colo. Red and gray arkosic conglomerate and sandstone and red and black mudstone of the Minturn Formation of the Eagle basin were derived by weathering and erosion of basement rocks and pre-Pennsylvanian sedimentary rocks on the Frontrange uplift and were deposited as fans marginal to the uplift.
Red shaly sandstone and calcareous sandstone north of the Uncompahgre uplift seem to have been derived from the Uncompahgre probably as fans or mudflats marginal to the uplift.

The Pathfinder uplift in interval C time was smaller than in the time of interval B and supplied little or no sediment to surrounding areas.

**PALEOTECTONIC IMPLICATIONS**

The Wyoming shelf apparently was one of great stability. The Oquirrh basin was still in existence, but it was smaller and subsidence was less than it was earlier in the Pennsylvanian Period. In southwestern Utah, the Piute positive element apparently was tectonically neutral, whereas the shelves on its east and west margins subsided only enough to permit deposition of a few hundred feet of carbonate and sandy carbonate rocks.

The narrow, linear, northwest-trending Central Colorado trough continued to be active during the time of interval C. Tectonic complexity of the Eagle basin, at the northwest end of the trough, increased with the appearance in Des Moines time of two subparallel subsidiary northwest-trending troughs.

The parallel trends of the Uncompahgre and Frontange uplifts, the bordering Paradox basin, and the intervening Central Colorado trough strongly suggest that these major features were tectonically related. Subsidence in excess of 5,000 feet in the Paradox basin and concomitant elevation of the Uncompahgre uplift only a few miles to the northeast, which provided large volumes of coarse material, suggest the action of complementary vertical forces. Similarly, vertical forces probably were coupled in the southeastern part of the Central Colorado trough and adjacent uplifts; subsidiary troughs may have been present in the Eagle basin where the pattern of tectonic forces apparently was more complex.

Absence of fringing detritals around the Pathfinder uplift suggests that in interval C time this element was either tectonically neutral or slowly subsiding and served only to influence the force and direction of longshore currents.

**INTERVAL D**

**FORMATIONS INCLUDED**

Formations composing interval D in the Middle and Southern Rocky Mountains, northern Colorado Plateau, and northeastern Great Basin are listed in table 31. Most of the rocks forming this interval are of Missouri age.

**UPPER BOUNDARY OF INTERVAL D**

The Lewiston Peak Member of the Oquirrh Formation was described by Bissell (1962, p. 31) as a distinctive assemblage of cherty limestone, silty-argillaceous limestone, and sandstone. Because this member contains a fusulinid fauna that has been assigned a Missouri age by Bissell, it is identifiable and separable from the overlying member on the basis both of fossils and of lithology. The top of interval D in northeastern Utah is, therefore, placed at the top of the Lewiston Peak Member.

In southeastern Wyoming the middle part of the Casper Formation, of Missouri age, has been assigned to interval D. Age identification is based primarily on work by Thomas, Thompson, and Harrison (1953) in the Laramie Mountains. Correlation of beds into the subsurface east, west, and north of the Laramie Mountains allows separation, with some degree of confidence, of rocks of Missouri age from the overlying rocks of Virgil age.

Elsewhere in the Wyoming-Colorado-Utah region, rocks were arbitrarily assigned to intervals D and E. In most places deposition seems to have been continuous from Missouri to Virgil time. Local hiatuses in the Missouri-Virgil sequence differ in age and, hence, may not be reliable guides to the series boundary. Facies changes are common in Upper Pennsylvanian rocks; hence, correlation by lateral tracing of lithologic sequences is less dependable than in the older strata. Interval D has been separated from interval E in much of the Paradox basin and the Central Colorado trough by the arbitrary division of strata beneath rocks of Permian age and above interval C into two parts of equal thickness.

**THICKNESS TRENDS**

The thickness of interval D generally is no more than 1,000 feet. The interval is absent across wide areas in the Wyoming-Colorado-Utah region (pl. 8A). All of southwestern Utah, nearly all of northwestern Wyoming, and parts of southeastern Idaho lack strata assigned to it. The Oquirrh basin was restricted; rocks deposited in...
that basin reached a maximum thickness of 2,000 feet at Provo. As with other intervals, isopachs in the Oquirrh basin generally trend north to northwest, except on the west margin, which was farther east than it was during interval C time. Maximum thickness of interval D in the basin was about 500 feet and average thickness about 350 feet.

The Central Colorado trough during interval D seemed to maintain the same approximate extent and the same northwest axial trend that it had during the time of interval C. The maximum thickness of sediment deposited was about 500 feet and the average thickness along the axis about 350 feet. Interval D is present in Wyoming only southeast of a roughly straight line that transects the central part of the State in a northeast direction. Average thickness of interval D in Wyoming is about 100 feet. Thickened deposits define the Sweetwater trough in Sweetwater County, but thickness trends only weakly reflect it farther northeast.

**LITHOFAECIES TRENDS**

Interval D in southeastern Idaho consists of interbedded sandstone and carbonate (pl. 8B). Along the axis of a tongue of thick strata extending northward from the Oquirrh basin into Idaho, carbonate rocks seem to dominate. Eastward, northward, and westward they grade into sandstone of the Wood River and Tensleep Formations. Control for this facies pattern is meager, however, and the simplicity of this interpretation may be more apparent than real. Carbonate rock dominates in the Oquirrh basin but seems to grade northeastward into sandy and argillaceous facies. In the Paradox basin carbonate rock is the dominant lithology within a wide, central, northwest-trending band. This belt is composed primarily of the Honaker Trail Formation (Wenger and Matheny, 1958, p. 2075), which is largely carbonate rock but which contains appreciable amounts of chert, mudstone, sandy siltstone, mica-siltstone, and arkose. Northeastward and southwestward these strata contain progressively more sand and mud. Arkosic conglomerate and sandstone are present along the southwest margin of the Uncompahgre uplift.

In the Central Colorado trough interval D consists of sandstone and some conglomerate, siltstone, and mudstone. The red Maroon Formation occupies most of the Eagle basin. This formation is coarest near the margins of the Uncompahgre and Frontrange uplifts and, in general, contains finer material toward the center of the basin, which is in the vicinity of the White River Plateau. In extreme northwestern Colorado and adjacent parts of Utah, the Maroon Formation intertongues with the white to cream-colored Weber Sandstone that is conspicuously exposed on the north and south flanks of the Uinta Mountains. The Weber is equivalent to part or all of the Tensleep Sandstone in Sweetwater County, Wyo., and the Tensleep, in turn, grades eastward into the Casper Formation of central Wyoming. A narrow tongue of carbonate rock and sandy carbonate rock in Converse and Carbon Counties occupies the site of the northeastern part of the Sweetwater trough of Morrow time.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

Interval D can be classified, according to source, into three major rock types: (1) carbonate rock, which was largely deposited in place, (2) red terrigenous detrital rock derived from the Ancestral Rocky Mountains, and (3) light-colored well-sorted detrital rock from a distant source.

Siltstone, mudstone, and sandstone beds in the Lewiston Peak Member of the Oquirrh Formation and the Honaker Trail Formation were derived from nearby source areas along the margins of the Oquirrh and Paradox basins.

Coarse arkosic fanglomerate adjacent to the southwest margin of the Uncompahgre uplift and red fanglomerates and sandstones (Maroon Formation) on both sides of the Eagle basin clearly were derived from the Uncompahgre and Frontrange uplifts. Light-colored well-sorted quartz sandstone in northwestern Colorado and adjacent parts of Wyoming and Utah (Weber and Tensleep) may have been derived by the erosion of older sandstone outside the region, although possibly a large part of the sandstone in interval D may have been derived from reworking of that part of the Tensleep Sandstone of Des Moines age of northwestern Wyoming.

 Beds of carbonate rock in southern Wyoming may have been deposited on an offshore shelf between nearshore shoals that received sand (Casper Formation) in central Wyoming and open-water carbonate deposits to the east.

**PALEOTECTONIC IMPLICATIONS**

The broad area in northwestern Wyoming and southwestern Utah where interval D is missing may represent emergence during interval D time or inundation during this time followed by sufficient uplift at a later time to allow removal of interval D strata, or some combination of these conditions.

Interval D strata thin gradually toward broad areas where the interval is missing in northwestern Wyoming and southwestern Utah. Moreover, along the margins of these areas, sandstones, or carbonate rocks containing sandstone, are present. These relations suggest uplift of adjacent areas during interval D time. In Emery County, Utah, however, the isopach patterns clearly indicate truncation following deposition of the interval.

The modest thickness of strata of interval D in the Paradox and Eagle basins suggests that subsidence decreased substantially during the time of that interval.
By analogy, the degree of uplift in adjacent areas was correspondingly less. In comparison, the large volume of red conglomerate, sandstone, and mudstone in northeastern Utah, representing the part of the Maroon Formation that is younger than Des Moines, suggests that uplift of the northern part of the Ancestral Rocky Mountains was still vigorous in the latter part of Pennsylvanian time.

**INTERVAL E**

**FORMATIONS INCLUDED**

Formations constituting interval E in the Middle and Southern Rocky Mountains, northern Colorado Plateau, and eastern Great Basin are shown in table 32; formations in major subdivisions of the region are shown also on plate 13. The interval consists of rocks largely of Virgil age.

<table>
<thead>
<tr>
<th>Area</th>
<th>Formations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eastern Idaho</td>
<td>Part of Copper Basin, Wood River, and Wells Formations.</td>
</tr>
<tr>
<td>Northwestern Utah</td>
<td>Pole Canyon Member of Oquirrh Formation.</td>
</tr>
<tr>
<td>Southwestern Utah</td>
<td>Part of Calvillo and Ely Limestones.</td>
</tr>
<tr>
<td>Southeastern Utah and</td>
<td>Part of Rico Formation and part of Honaker Trail Formation of Hermosa Group.</td>
</tr>
<tr>
<td>southwestern Colorado</td>
<td></td>
</tr>
<tr>
<td>South-central Colorado</td>
<td>Part of Sangre de Cristo Formation.</td>
</tr>
<tr>
<td>Northwestern Colorado and</td>
<td>Part of Maroon Formation and Part of Weber Sandstone.</td>
</tr>
<tr>
<td>northeastern Utah</td>
<td></td>
</tr>
<tr>
<td>Central and southern</td>
<td>Part of Tensleep Sandstone and part of Casper, Fountain, and Minnelusa Formations.</td>
</tr>
<tr>
<td>Wyoming</td>
<td></td>
</tr>
</tbody>
</table>

**UPPER BOUNDARY OF INTERVAL E**

The upper part of the Pole Canyon Member of the Oquirrh Formation in northwestern Utah as described by Bissell (1959, p. 126) contains massive orthoquartzites which contrast with conformably overlying argillaceous and silty-sandy limestone, calcareous sandstone, and some orthoquartzite beds considered to be of Permian age.

In the Paradox basin the top of interval E is difficult to establish with certainty, inasmuch as few faunal data are available for dating and the Pennsylvanian and basal Permian rocks commonly are lithologically similar. In most parts of the basin the upper boundary is placed at the top of the main body of limestone, which is overlain by rocks that are mostly sandstone. At the margins of the basin the contact is within the sandstone sequence. This boundary is probably not everywhere picked at the same horizon.

In the Central Colorado trough, the systemic boundary, which is at or near the upper boundary of interval E, is especially obscure. Cotylosaur and pelycosaur bones, found in beds 1,570 feet above the base of the Sangre de Cristo Formation, are from a location south of Salida (Brill, 1952, p. 822). Because similar bones occur in strata of Wolfcamp age in Kansas, Oklahoma, Texas, and New Mexico, the Sangre de Cristo Formation at and above this horizon can be considered to be of Permian age. Elsewhere in the trough, fossils are limited almost entirely to algae that occur in very thin isolated limestone lenses. In the absence of other age criteria, a horizon below that of the reptile bones has been arbitrarily extended north and south throughout the trough as the top of the interval. Hence, the thickness given for interval E in northwestern and south-central Colorado is only an estimate.

Interval E in the Laramie Mountains is coextensive with rocks of Virgil age as based on fusulinid dating by Thomas, Thompson, and Harrison (1953). Lithologic units can be correlated from the mountains into the subsurface with a fair degree of confidence. Identification of the upper boundary of interval E in southeastern Wyoming is markedly aided by the occurrence of a distinctive red marker bed at the contact with the overlying Permian System (Maughan, 1967, p. 132). Locally in that area the contact is a disconformity. In some places the entire interval has been removed by erosion beneath the disconformity, but elsewhere no beveling is demonstrable.

In the Uinta Mountains the top of the Pennsylvanian occurs within the Weber Sandstone, according to Bissell and Childs (1958). For the present study, the boundary is placed just below the zone where Bissell and Childs (1958, pl. 1) indicate the occurrence of Schwagerina.

**THICKNESS TRENDS**

Interval E has limited extent in the Wyoming-Colorado-Utah region (pl. 9A). It is missing in nearly all of western Wyoming, in central and southwestern Utah, and in much of central and western Colorado. It occupies the Oquirrh basin but is restricted to the Provo area in the vicinity of Great Salt Lake, where the greatest thickness of strata is about 2,000 feet, and the thickness averages about 1,000 feet. The axial part of the basin, as shown by isopach patterns in this area, trends generally northwest, as for other intervals.

Interval E is present in the Paradox basin, where it attains a maximum thickness of less than 1,000 feet and averages about 600 feet. Isopach patterns are irregular, but an overall north to northwest trend for the basin is discernible.

In the Eagle basin, where interval E is also represented, the greatest thickness is about 500 feet; however, thickness patterns in this basin should be accepted with
reservations in view of uncertainties in consistently establishing the upper boundary of the interval. In southwestern Wyoming the average thickness is about 100 feet. Presence of the Sweetwater trough is vaguely suggested by a slight thickening of the interval.

**LITHOFACIES TRENDS**

In the Oquirrh basin a well-defined facies change takes place from sandstone on the east through calcareous sandstone and calcareous shaly sandstone to carbonate rock on the west (pl. 98).

In the Paradox basin, mixtures of sandstone, siltstone, mudstone, and carbonate rock are irregularly distributed (upper part of the *Honaker Trail Formation* and the lower part of the Rico Formation). A band of arkosic conglomerate and sandstone parallels the southwest margin of the Uncompahgre uplift. In the western part of the basin, anhydrite is interbedded with sandstone and mudstone.

In the Central Colorado trough, red sandstone, fine conglomerate, arkosic conglomerate, siltstone, and mudstone (*Maroon and Sangre de Cristo Formations*) compose the entire interval. Local carbonate beds a few inches thick, commonly gray or black, are present but account for only a very small fraction of the total volume of rock. Coarsest strata are present along the margins of the Frontrange and Uncompahgre uplifts. Red sandy siltstone, siltstone, and mudstone compose most of interval E in the center of the Eagle basin, especially in the vicinity of the White River uplift.

The red beds of the Maroon Formation in northwestern Colorado intertongue with light-colored beds of the Weber Sandstone in northwestern Utah; the Weber, in turn, grades into calcareous sandstone, sandy carbonate rocks, and carbonate beds of the Casper Formation in eastern Wyoming.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

Sandstone and related coarse sedimentary rocks of interval E in central-southern Idaho (Wood River Formation) were probably derived from a highlands to the west which may have resulted from uplifts along parts of the Antler orogenic belt in Nevada. Sandstone in interval E in central Wyoming, northwestern Colorado, and northeastern Utah, and east of Great Salt Lake suggests a local source. The adjoining area in western Wyoming and adjacent States, which lacks strata assignable to interval E, was emergent and may have supplied the detritus. Mudstone, in the vicinity of Great Salt Lake, may have come from a different source to the east or southeast. Carbonate rocks west of a sandstone and mudstone band in northwestern Utah probably were deposited in a trough or embayment far from land.

The high proportion of sandstone, siltstone, and mudstone in the upper part of the *Honaker Trail Formation* and the lower part of the Rico Formation in the Paradox basin, in contrast to the high content of carbonate rock in the interval D part of the *Honaker Trail Formation*, suggests that marine withdrawal took place with concomitant invasion of the basin area by terrigenous sediments from adjacent land. Presence of anhydritic strata in the western part of the basin supports a concept of shoaling waters and restricted circulation. A belt of arkosic fanglomerate along the southwest margin of the Uncompahgre uplift indicates that the area was rising and contributing detrital material to the adjacent basin.

In the Central Colorado trough and to a limited extent in the Laramie basin, red terrigenous rocks of the Sangre de Cristo, Maroon, and Fountain Formations apparently were derived from the Uncompahgre and Frontrange uplifts. Farther north, in central Wyoming, the Tensleep Sandstone grades eastward into calcareous sandstone, sandy carbonate, and carbonate rocks of the Casper and Minnelusa Formations.

**PALEOTECTONIC IMPLICATIONS**

As with interval D, interval E progressively thins to broad areas that have no deposits in northwestern Wyoming and southwestern Utah. Sandstone, or carbonate rock containing sandstone, is present along the flanks of interval E basins, suggesting that emergence of source areas took place during interval E time. In parts of southern Utah where isopachs abut the Piute positive element and elsewhere in the region, some erosion of interval E rocks probably took place.

The relatively high proportion of sandstone and mudstone in the sediments of this interval in the Paradox basin area seems to foreshadow a blanketing of the basin by sand and mud of the Cutler Formation during Permian time. The belt of arkosic fanglomerate on the southwest margin of the Uncompahgre uplift and the large volume of red sandstone, fine conglomerate, siltstone, and mudstone in the Maroon and Sangre de Cristo Formations in the Central Colorado trough indicate that elevation of the adjacent Uncompahgre and Frontrange uplifts was extensive during interval E time and provided a large supply of coarse terrigenous sediments to adjacent areas of deposition.

**TOTAL THICKNESS OF PENNSylvANIAN ROCKS**

**THICKNESS TRENDS AND PALEOTECTONIC IMPLICATIONS**

Rocks of Pennsylvanian age in the region were deposited in four tectonic provinces (fig. 62): (1) The Cordilleran miogeosyncline, which includes the Oquirrh basin in northwestern Utah and the similar Wood River basin in southeastern Idaho, (2) the Paradox basin in
southeastern Utah and southwestern Colorado, (3) the Central Colorado trough in northeastern Utah, northwestern Colorado, and south-central Colorado, and (4) the Wyoming shelf. The Sweetwater trough is a northeast-trending branch of the Eagle basin part of the Central Colorado trough which extended into southern Wyoming and at certain times had mild expression on the Wyoming shelf. The Plute positive element and the Uncompahgre and Frontrange uplifts separated the basins and were not sites of deposition.

Thicknes of Pennsylvanian strata along the axis of the Cordilleran miogeocline reaches 10,000–15,000 feet in the Oquirrh basin and decreases southward and northeastward. Original depositional thickness (prior to diapiric thickening) in the central part of the Paradox basin seems to have been about 6,000 feet. In the narrower, tectonically more complex Eagle basin, rocks of Pennsylvanian age may reach a maximum thickness of 12,000 feet, but they thin in all directions from local foci of subsidence.

Isopach patterns on the Wyoming shelf have no well-defined trends, indicating that the shelf may not have subsided as a simple tilted plane but instead was mildly unstable locally. Average thickness of Pennsylvanian rocks is about 500 feet on the shelf.

Location and trends of the Ancestral Rocky Mountain uplifts coincide with those of the existing Rocky Mountain ranges only in a few local areas where Laramide trends follow lines of weakness attributable to Pennsylvanian mountain-building.

GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM

Rocks of Early Permian (Wolfcamp) age overlie directly the Pennsylvanian rocks in the region except in western and central Wyoming (pl. 12). Deposition was continuous in Utah and western Colorado from Virgil to Wolfcamp time, making it difficult to separate rocks assignable to these series. In southeastern Wyoming, a brief interval of emergence caused local beveling at the close of Virgil time and allowed the deposition of a thin bed of fluvial origin, called informally the “red marker.”


REFERENCES


1962, Pennsylvanian-Pennsylvanian Oquirrh basin of Utah, in Geology of the southern Wasatch Mountains and vicinity, Utah: Brigham Young Univ., Geology Studies, v. 9, pt. 1, p. 20–49.


Montana, North Dakota, Northeastern Wyoming, and Northern South Dakota

By EDWIN K. MAUGHAN

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-O
CONTENTS

Abstract .................................................. 279
Region defined ........................................... 279
Paleogeography ........................................... 279
Units underlying Pennsylvanian ...................... 279
Lower boundary of Pennsylvanian ..................... 280
Paleotectonic implications ............................. 280
Interval A ................................................ 281
Formations included .................................... 281
Upper boundary of interval A ......................... 282
Thickness trends ........................................ 282
Lithofacies trends ...................................... 283
Sources and environments of deposition ............ 283
Paleotectonic implications ............................ 284
Interval B ................................................ 284
Formations included .................................... 284
Stratigraphic relations ............................... 285
Upper boundary of interval B ......................... 285
Thickness trends ........................................ 285
Lithofacies trends ...................................... 285
Sources and environments of deposition ............ 285
Paleotectonic implications ............................ 286
Interval C ................................................ 286
Formations included .................................... 286
Interval C—Continued
Stratigraphic relations ............................... 286
Upper boundary of interval C ......................... 287
Thickness trends ........................................ 287
Lithofacies trends ...................................... 287
Sources and environments of deposition ............ 287
Paleotectonic implications ............................ 288
Interval D ................................................ 288
Formations included .................................... 288
Upper boundary of interval D ......................... 288
Thickness trends ........................................ 289
Lithofacies trends ...................................... 289
Sources and environments of deposition ............ 289
Paleotectonic implications ............................ 289
Interval E ................................................ 289
Total thickness of Pennsylvanian rocks ............. 290
Thickness trends ........................................ 290
Paleotectonic implications ............................ 290
Geologic units directly above Pennsylvanian System 290
Units overlying Pennsylvanian ....................... 291
Paleotectonic implications ............................ 291
References ................................................. 292

ILLUSTRATIONS

[For listing of plates (in separate case) see volume "Contents"]

Figure 63. Map showing major paleogeographic and structural features affecting deposition or distribution of Pennsylvanian strata in Montana and parts of Idaho, Wyoming, and the Dakotas ........................................ 280
64. Geographic features in Montana and parts of Idaho, Wyoming, and the Dakotas mentioned in text ........................................ 281
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVIANIAN SYSTEM IN THE UNITED STATES.
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVIANIAN SYSTEM

MONTANA, NORTH DAKOTA, NORTHEASTERN WYOMING, AND NORTHERN SOUTH DAKOTA

BY Edwin K. MAUGHAN

ABSTRACT

Pennsylvanian rocks in the northern Rocky Mountains and Great Plains were deposited mostly on a broad shelf between the Cordilleran geosyncline, which extended into the western part of the region of this chapter, and the land area of the Canadian Shield to the east. Represented in the region are the five Pennsylvanian intervals, A through E, which approximate the midcontinent Pennsylvanian time-stratigraphic subdivisions, Morrow through Virgil. Formations of Pennsylvanian age are the Amsden and Quadrant Formations in western Montana and northeastern Wyoming; the Tyler Formation, Alaska Bench Limestone, and Devils Pocket Formation of the Amsden Group and, locally, an overlying remnant of the Quadrant Formation in central Montana; and parts of the Minnelusa Formation in eastern Montana and adjacent parts of North and South Dakota and northeastern Wyoming. In south-central Montana, Pennsylvanian rocks consist of the Amsden Formation and Tensleep Sandstone.

Pennsylvanian detrital sediments of limited areal extent, and some limestone, were deposited in earliest Pennsylvanian time in a transgressing sea after the regional uplift, folding, and faulting that ended the Mississippian Period. The initial Pennsylvanian sediments are represented in central Montana by the Tyler Formation (Early Pennsylvanian, Morrow age). They were largely derived by erosion of Mississippian strata from broadly uplifted areas in south-central Montana and adjacent parts of Wyoming and from the craton to the east. The Pennsylvanian sea expanded eastward from the Cordilleran geosyncline; downfolded areas were filled with sediment; and intervening land areas were progressively flooded. Source areas became more distant. In early Middle Pennsylvanian (Atoka) time, deposition of detritus was replaced in central Montana by widespread deposition of calcium carbonate, represented by the Alaska Bench Limestone. This limestone is believed to have thinned toward a shore in the central or eastern part of the Dakotas. The Alaska Bench is now absent at many places in the region because of erosion from areas where it was uplifted by warping and, possibly, by minor faulting of the shelf in middle or late Atoka time.

During the Middle Pennsylvanian, chiefly in Des Moines time but possibly beginning as early as late Atoka, deposition of sand, represented by part of the Quadrant Formation, was renewed in the western part of the region. The sand seems to have come from land to the west, perhaps within the Cordilleran geosyncline. Mostly carbonate sediment, represented by part of the Minnelusa Formation, was deposited in the eastern part of the region. Local uplift in the western land area probably occurred contemporaneously with the regional upwarping elsewhere and terminated the deposition of calcium carbonate.

During interval C (Des Moines) time, an actively sinking trough in western Montana, either the main part of the Cordillerangeosyncline or a branch of it, received an accumulation of at least 1,700 feet of mostly quartz sand; adjacent parts of the shelf lying east of the geosyncline received a thickness of about 300 feet, mostly of mud and calcium carbonate.

Epigenetic movements of the craton similar to those that began in Middle Pennsylvanian time continued in Late Pennsylvanian (Missouri and Virgil) time; however, sediments accumulated in increasingly smaller areas. In Late Pennsylvanian to Early Permian time, much of the shelf area was uplifted. Upper Pennsylvanian strata, especially those of Virgil age, were removed at most places before deposition of Permian rocks. Before Middle Jurassic time, renewed uplift and erosion removed all Pennsylvanian strata that once covered northern Montana and large parts of North and South Dakota.

REGION DEFINED

Pennsylvanian rocks in Montana, North Dakota, northern South Dakota, and a small part of northern Wyoming, are described in this chapter (fig. 1, chap. A). The included part of South Dakota is north of the Black Hills base line in the western part of the State; also included are two northeastern counties of Wyoming, Campbell and Crook, and the northern part of Yellowstone National Park in the northwestern part of the State. To clarify the geology of western Montana, reference is made to the geology of central Idaho; for further discussion of Idaho, see chapter N of this publication. For brevity, the region described in this chapter is referred to as the “Montana and Dakota region” or the “northern Rocky Mountains and Plains region.”

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVIANIAN

Pennsylvanian rocks rest unconformably upon Mississippian rocks throughout this region except in a small area near central South Dakota (pl. 2). The Mississippian rocks consist of the Madison and Big Snowy Groups, and the Pahasapa Limestone.
The Madison Group is composed of Lodgepole Limestone of Kinderhook and Osage age, overlain by the Mission Canyon Limestone of Osage and Meramec age, and the Charles Formation of Meramec age. The upper part of the Mission Canyon Limestone intertongues with the Charles Formation in the eastern part of the Montana and Dakota region. The equivalent of the Madison Group in the Black Hills area of northeastern Wyoming and northwestern South Dakota is the Pahasapa Limestone.

The Big Snowy Group, which overlies the Madison Group in central and eastern Montana and in North Dakota, is composed in ascending order of the Kibbey, Otter, and Heath Formations of Late Mississippian (Chester) age. Equivalent strata in southwestern Montana are identified as the Big Snowy Formation.

The Big Snowy Group and equivalent rocks are preserved beneath Pennsylvanian rocks in many parts of this region. A relatively narrow band of these rocks extends northeasterly from Idaho into southwestern Montana and is probably connected with another narrow band of Chester rocks that extends eastward across central Montana. The latter band broadens farther east into a wide remnant of Chester rocks in eastern Montana and adjacent parts of the Dakotas and is approximately coextensive with the Williston basin (fig. 63). Elsewhere, especially in southern Montana, northeastern Wyoming, and most of South Dakota, Chester rocks are absent and the Pennsylvanian rests upon rocks of Meramec age. Farther southeast in South Dakota, rocks of Meramec age are truncated and the Pennsylvanian strata rest upon Lower Mississippian beds (Osage and possibly, locally, Kinderhook).

Pennsylvanian rocks are known to rest upon rocks older than Mississippian only in one small area, chiefly in Stanley County, S. Dak. (pl. 2; fig. 64). Here, Precambrian crystalline rocks underlie Pennsylvanian rocks in one drill hole (loc. 22, pl. 1), and thin Mississippian strata, probably no younger than Kinderhook age, underlie Pennsylvanian in a nearby hole (loc. 62). Paleozoic rocks older than Mississippian may underlie Pennsylvanian in a narrow belt between these two drill holes.

**LOWER BOUNDARY OF PENNSylvIAN**

Mississippian rocks throughout the northern Rocky Mountains and Plains region were beveled before deposition of the overlying Pennsylvanian strata. This beveled surface truncates increasingly older strata southward from central Montana into Wyoming and southeastward in the Dakotas. In general, the surface forms a low angle with underlying strata, but in a few places strata are sharply truncated, suggesting the probability of local Late Mississippian to Early Pennsylvanian tectonism.

**PALEOTECTONIC IMPLICATIONS**

The Montana and Dakota region was relatively stable during most of Late Mississippian time. Very late in Mississippian or very early in Pennsylvanian time the region was differentially uplifted above sea level and Chester rocks were removed from south-central Montana and adjacent parts of Wyoming. Uplift was at a minimum in the region of the Williston basin, which seems at that time to have had its center in eastern Montana and western North Dakota adjacent to the faulted Cedar Creek structure (fig. 63). Streams were incised in the upland surface, creating considerable local and regional relief. This locally rough surface was buried and preserved when the sea returned in Early Pennsylvanian time.

Faults or fault systems probably were active in at least three areas within the region at the end of Mississippian time, as indicated by the juxtaposition beneath Pennsylvanian strata of Mississippian rock sequences having markedly different thicknesses. Lithofacies and thickness of individual stratigraphic units of the Mississippian near these probable faults suggest that the uplifted blocks were not sources of detritus during the Mississippian, and that originally Mississippian rocks were evenly and widely distributed across the region. Faulting and erosional beveling apparently took place at the end of or after the Mississippian.

In central Montana a probable fault of considerable magnitude (Maughan and Roberts, 1967, pl. 3) trends west-northwest not far south of the Big Snowy Mountains. In these mountains the Big Snowy Group attains a thickness of 1,200 feet. It is abruptly truncated and is absent south of the fault, although originally rocks of the group probably extended southward well into northern Wyoming. Probably these rocks were only weakly indurated at the time of faulting, and erosion on the upthrown side of the fault or faults proceeded rapidly.

![Figure 63](image-url) — Major paleogeographic and structural features affecting deposition or distribution of Pennsylvanian strata in Montana and parts of Idaho, Wyoming, and the Dakotas.
In western Montana the Big Snowy Formation is truncated and generally absent east of a probable north-northeast-trending fault that is about coincident with the Greenhorn fault of Cretaceous age (fig. 64) in Beaverhead and Madison Counties and with the Cretaceous Lombard thrust fault in Gallatin and Broadwater Counties. The amount of displacement of faults active in Early Pennsylvanian time in Beaverhead and Madison Counties can be measured by comparison of the thickness of the Big Snowy Formation preserved on the downthrown side with that on the upthrown side, the displacement being about 800 feet.

In eastern Montana a third probable Late Mississippian to Early Pennsylvanian fault or fault system occurs. It coincides with the Cedar Creek anticline in eastern Montana. Strata of the Big Snowy Group are present beneath Pennsylvanian rocks on both sides of this fold, but are absent from the crest; the thickness preserved on the east side is greater than on the west.

Big Snowy rocks, together with Pennsylvanian rocks, have been abruptly truncated and removed to the north of another system of probable faults directly north of the Big Snowy Mountains. This faulting, however, took place after Pennsylvanian time. Uniform lithofacies and bedding in the Big Snowy Group suggest that the Upper Mississippian rocks near the Big Snowy Mountains were deposited well offshore, near a basin center, and not near the basin margins or in a narrow trough. The facies relations in Mississippian rocks, therefore, support the interpretation of post-Mississippian faulting and erosion.

**INTERVAL A**

**FORMATIONS INCLUDED**

The Tyler Formation of the Amsden Group constitutes interval A in Montana, and strata of Early Pennsylvanian (Morrow) age in the lower part of the Minnelusa Formation constitute interval A in adjacent parts of North Dakota, South Dakota, and Wyoming (pl. 13). Detrital beds in the upper part of the Tyler Formation intertongue in some areas with the Alaska Bench Limestone and equivalent limestone strata of the Minnelusa as shown by key beds and therefore, they are partly of Morrow age. On this basis, interval A in parts of eastern Montana and western North Dakota is believed to also include some of the Alaska Bench Limestone although most of that formation is in interval B.

Rocks mapped as interval A in Montana are
equivalent to the lower sandstone and red mudstone sequence of the Amsden Formation in north-central Wyoming and have been traced by me, through subsurface correlation, into Division V (Reclamation Member) (Condra and Reed, 1935) of the Hartville Formation in central-eastern Wyoming.¹ Equivalent rocks make up the upper part of unit B and all of unit C of Shannon (1961) in central Idaho. Mamet, Skipp, Sando, and Mapel (1971) suggest that the upper part of unit B may be of Mississippian age. Sandstone, equivalent to the Tyler Formation, that has been mapped as part of the Quadrant Formation in the Beaverhead Mountains near Leadore, Idaho (Ruppel, 1968) has also been included in interval A.

The equivalent to the Tyler Formation of central Montana in the the Rocky Mountains of Alberta and British Columbia is considered by Halbertsma and Staplin (1960) to be the Tunnel Mountain Formation, the lowest formation of the Rocky Mountain Group. The Tunnel Mountain is largely or entirely of Early Pennsylvanian (Morrow) age; however, it may be partly Late Mississippian (late Chester) (McGugan and Rapson, 1962, p. 360-361; and McGugan and May, 1965). The Rocky Mountain Quartzite (sandstone of Sweeney, 1955) has been used for approximately equivalent rocks of Pennsylvanian age adjacent to the Canadian border in the Whitefish Range near Trail Creek in northwestern Montana. Although these exposures of Rocky Mountain Quartzite in northwestern Montana are geographically near and physically somewhat similar to the Tunnel Mountain Formation, it is included in interval C in this publication because of its greater similarity to the Quadrant Formation of interval C to the south in western Montana.

The upper part of the Yakinikak Limestone, which underlies the Rocky Mountain Quartzite in the Trail Creek area, possibly should be included in interval A. The lower part of the Yakinikak has yielded fossils of Late Mississippian age (Willis, 1902, p. 316; Sweeney, 1955; Blackstone, 1934), but the upper part is undated paleontologically and could include strata of Pennsylvanian age.

Strata assigned to interval A in the northern Rocky Mountains and Plains region are considered to be of Early Pennsylvanian (Morrow) age on meager paleontological data. The Stonehouse Canyon Member of the Tyler Formation in central Montana and the lower part of the Amsden Formation of southwestern Montana are dated chiefly by pollen (Maughan and Roberts, 1967, p. B21). The Tyler Formation also has produced a few Early Pennsylvanian brachiopods in Montana (Willis, 1959, p. 1959), and an extensive Early Pennsylvanian os-tracode fauna, a few brachiopods, and some pelecypods in North Dakota (Willis, 1959, p. 1959-1962; Foster, 1961, p. 93; Ziebarth, 1964, p. 124-125).

The lower part of the Minnelusa Formation in the Black Hills has furnished foraminifers identified as Early Pennsylvanian forms (B. L. Skipp, written commun., 1966). Also, both the Cameron Creek Member of the Tyler Formation, which probably is the equivalent of the basal Minnelusa red beds, and the lowermost strata in the overlying Alaska Bench Limestone contain fossils of Early Pennsylvanian age (J. T. Dutro, written commun. to W. W. Mallory, 1962; Gorman, 1963; Maughan and Roberts, 1967, p. B21).

**UPPER BOUNDARY OF INTERVAL A**

Where interval A is overlain by interval B, the contact is gradational, but in many places interval B strata have been removed and the upper boundary of interval A is an unconformity. In most places where intervals A and B are both present in Montana, the boundary is arbitrarily placed at a surface indicated by change from dominantly red beds and interbedded limestone beds below to dominantly limestone beds above. This change occurs in western and central Montana at the boundary between the Tyler Formation and the Alaska Bench Limestone. At most places in eastern Montana and adjacent parts of North Dakota, South Dakota, and northeastern Wyoming, the contact between intervals A and B has been placed at an approximate time-stratigraphic boundary within the lower part of the Alaska Bench Limestone and within the lower part of the Minnelusa Formation (pl. 10F, section II-1 "I' and IV-1 "J' ).

In most places where rocks of interval B are absent, interval A is overlain unconformably by rocks of late Atoka to early Des Moines age that are assigned to interval C. Along and near the northern and eastern limits of Pennsylvanian rocks, strata of interval A are unconformably overlain by Jurassic and, in a small area, by Cretaceous rocks (pl. 12).

**THICKNESS TRENDS**

Interval A in much of the Montana and Dakota region ranges in thickness from a few feet to 250 feet (pl. 3A). In most parts of southeastern Montana, northeastern Wyoming, and northwestern South Dakota it is less than 100 feet thick and locally it is absent. Interval A thickens northeastward and northward in eastern Montana toward the center of the Williston basin where interval A rocks locally reach a maximum thickness of about 250 feet. The zero edge of these rocks on the northeast is erosional and approximately parallels the course of the Missouri River through the Dakotas. This erosion occurred during pre-Middle Jurassic time.

Interval A is somewhat thicker in central and southwestern Montana than elsewhere in the region. The
Tyler Formation is as much as 800 feet thick on the northeast slope of the Little Belt Mountains (fig. 64), and the part of the Amsden Formation in interval A is about as thick in the Tendoy Mountains.

LITHOFAICIES TRENDS

Interval A in this region is composed chiefly of mudstone (pl. 3B) but sandstone is also present in much of the lower part and limestone in much of the upper part. The many variations in facies shown on plate 3B result in large measure from differences in the depth of erosion in the upper, relatively limy part of the interval, rather than from regional facies differences in the originally deposited sediments. Only in southwestern Montana does an increase in percentage of sandstone seem to be the result of a depositional trend.

In the center of the Williston basin of eastern Montana and western North Dakota, the upper part of interval A consists largely of limestone, but along the basin margins this limestone intertongues with mudstone and sandstone. Although the trends is somewhat obscure on the lithofacies map because of local erosion at the top of the interval, the nature of the transition is readily apparent in cross section II-I’. In the same area the lower part of interval A consists dominantly of sandstone and mudstone. Lithofacies patterns depicting these rocks, like those of the strata above, are very irregular. Inconsistencies in the lower strata result, however, from the surface relief upon which sediments of earliest Pennsylvanian age were deposited. In general, local depressions were filled with detrital sediments at an early stage, whereas topographically higher areas were covered later by carbonates that formed a blanket across the region.

In central Montana the lower or Stonehouse Canyon Member of the Tyler Formation consists of three distinct stratigraphic units. The amount of sandstone within each of these units differs from place to place within the Big Snowy Mountains and locally little or no sandstone is present. This three fold development of sandstone is conspicuous in the Big Snowy Mountains; a similar sequence has been noted in the Northwest Sumatra oil field, about 75 miles to the east (Foster, 1956), in the subsurface of eastern Montana (Willis, 1959) and in North Dakota (Willis, 1959; Ziebarth, 1964). Whether the three sandstone units in these different areas formed contemporaneously or can be correlated is not known, but clearly each sequence represents three distinct episodes of sand accumulation in or near the margins of the Tyler Sea.

In southwestern Montana, dominantly dark gray mudstone and interbedded lenses of fine conglomeratic and fine- to medium-grained sandstone form the lower part of interval A deposits. These deposits seem to be limited in lateral extent and likely accumulated rapidly. Mudstone and associated sandstone that compose the upper part of interval A are more widespread and everywhere are reddish brown. Eastward some of these red beds intertongue laterally into the Alaska Bench Limestone (pl. 13; p. 10F, section II-I’).

SOURCES AND ENVIRONMENTS OF DEPOSITION

As the sea advanced across western and central Montana in Early Pennsylvanian time, sediments derived from nearby, recently exposed land areas were at first deposited only in restricted areas of depression. Later as the sea transgressed farther eastward, covering a wide area, sediments from more distant sources apparently were spread rather uniformly across all of the Montana and Dakota region.

The main sources of the lowest Pennsylvanian sediments in south-central Montana and adjacent parts of Wyoming were the Siouxia land area southeast of the Williston basin and local areas within the Montana-Dakota region that had been elevated late in Mississippian or early in Pennsylvanian time. From these newly uplifted areas, rocks of the Mississippian Big Snowy Group were eroded and stripped (Todd, 1959, p. 69; Cooper, 1956, p. 84; Maughan and Roberts, 1967) with the result that lowest Pennsylvanian strata closely resemble in composition and color the rocks of the subjacent Big Snowy Group. An exception is sandstone in the Tyler, the composition of which differs from that in the Kibbey (Willis, 1959, p. 1954), suggesting that sand in these two formations may have been derived from distinctly different sources.

The distribution of sandstone and limestone in different parts of interval A in central Montana, although not clearly shown by the lithofacies pattern, is important in the interpretation of sources and environments of deposition and in determining the paleotectonics. Sandstone beds in the Stonehouse Canyon Member thin northward in the Big Snowy Mountains, and limestone beds, especially the Bear Gulch Limestone Tongue that forms the top of the lower member of the Tyler (Foster, 1956; Mundt, 1956a), thicken in the same direction. This relation suggests that a depositional edge was nearby to the south of the exposures in the Big Snowy Mountains.

Lower Pennsylvanian sediments seem to have been deposited at least partly as deltas at the edge of a sea advancing across Early Pennsylvanian land areas. Both marine (Gardner, 1959) and nonmarine environments (Mundt, 1956b, p. 1930) have been suggested for the Tyler Formation. The lithofacies and biofacies indicate a range from fresh water through brackish water to normal marine conditions (Willis, 1959; Foster, 1961; Ziebarth, 1964; Maughan and Roberts, 1967). Distinctly different depositional environments seem to have existed simultaneously in adjacent areas (pl. 14) and to have followed one another in succession in each area. A
marine environment in some areas is indicated by limestone beds containing brachiopods and some pelecypods. Elsewhere, brackish- to freshwater environments are indicated by much richer carbonaceous mudstone, in which inarticulate brachiopods including *Lingula*, intermingled marine and nonmarine ostracodes, and plant fragments are common (Willis, 1959; Foster, 1961; Ziebarth, 1964).

Lenticular sandstone bodies occur within the Tyler Formation at the three separate horizons that represent three distinct episodes when sand was the dominant sediment accumulating at the margins of the Tyler Sea. These sandstones are unfossiliferous except for plant remains, including probable detrital coal fragments (Easton, 1962, p. 26; Mundt, 1956b, p. 1922). In central Montana these sandstone lenses have been called river-channel deposits (Foster, 1956, p. 119, 122; Mundt, 1956b, p. 1922; Todd, 1959, p. 69, 71). Farther east, principally in the Williston basin, they are considered to be offshore bars (Harris, 1958, p. 44; Willis, 1959, p. 1958-1959). Probably both kinds of sand deposits occur in the Tyler. In the Big Snowy Mountains, sandstone structures seem indicative of deposition as offshore bars. In most of eastern Montana and in North Dakota the facies patterns suggest deposition of sand in deltaic river channels and levees (pl. 14).

The complex facies relationships and juxtaposition of marine and nonmarine deposits suggest sedimentation on a delta such as that of the Mississippi River where sand, mud, and calcium carbonate are included, and freshwater, brackish-water, and marine deposits are intermingled. The environment during deposition of the Tyler has been described as "***a broad flatland of sluggish streams, lagoons, and steaming Pennsylvanian jungles" (Foster, 1961, p. 93).

A source area for lowest Pennsylvanian sediments of western Montana and nearby areas probably was somewhere in central or northern Idaho. Rocks in the Beaverhead Mountains of western Montana and adjacent parts of Idaho consist of sandstone which tongues eastward into dark-gray shaly mudstone of the Tyler Formation. This relationship suggests a source of sand to the west or northwest. The sand probably was deposited in a marine environment, but no fossils having genetic implications have been reported. Possibly the sand was a nearshore deposit flanking a geanticlinal positive element west of or within the Cordilleran geosyncline.

Late in interval A time, local source areas were leveled and inundated by the spreading Early Pennsylvanian sea. Terrigenous muds were brought into the region from increasingly greater distances but in less volume than previously. Land lying somewhere east of the Williston basin, possibly as near as the eastern part of North and South Dakota, was a probable source of some of these muds, as indicated by an eastward facies change from carbonate to detrital rock within the basin.

A sparse fauna in the upper part of the Tyler Formation is marine. Terrigenous muds either remained oxidized or later became oxidized to form red beds. The nearly uniform thickness and wide lateral extent of these red beds suggest shallow-water deposition on a broad shelf. By the end of interval A time the entire region was covered by a shallow sea and deposition of lime muds was widespread. The sea expanded far beyond the limits of the northern Rocky Mountains and Plains region.

**PALEOTECTONIC IMPLICATIONS**

Lower Pennsylvanian sediments of interval A were deposited in the eastern and south-central part of this region on a broad, gently undulatory surface between the Cordilleran geosyncline to the west and a cratonic land area to the east. Thicker sections of interval A in southwestern Montana accumulated in a depositional trough that extended northeastward into central Montana. The trough was bounded on the south by an uplift, probably faulted, that had been active following deposition of Upper Mississippian (Chester) strata. The area of relatively thick Lower Pennsylvanian rocks in central Montana is the remnant of a probably larger depositional basin that extended into northern Montana; Pennsylvanian rocks have been removed from that area so the original extent of those rocks to the north is unknown.

Sinking began in local basins that either formed or were reactivated by Late Mississippian or Early Pennsylvanian time. These small, local movements were superimposed on widespread epeirogenic sinking that took place throughout the region. The sinking of basins and uplift of adjacent land areas may have been periodic; three episodes are suggested by the sandstone bodies at three stratigraphic levels in the lower part of the Tyler Formation.

By the end of Early Pennsylvanian time local basins apparently had ceased to form and the entire region seems to have become part of a slowly and rather uniformly sinking continental shelf.

**INTERVAL B FORMATIONS INCLUDED**

The Alaska Bench Limestone is the principal formation included in interval B in Montana (pl. 13), although some of the Amsden Formation constitutes the interval in western Montana. Part of the lower member of the Minnelusa Formation forms interval B in North and South Dakota and northeastern Wyoming (pl. 10F, section II-I' I'). The name Amsden Formation has been extended by some geologists into North Dakota for
Lower Pennsylvanian strata; and the names Tyler and Alaska Bench are used in North Dakota by some geologists.

**STRATIGRAPHIC RELATIONS**

The Alaska Bench Limestone and its equivalents form a widespread, lithologically and stratigraphically distinct unit that grades downward into the Tyler Formation of interval A. In most parts of the region, the boundary between these formations is arbitrarily placed with a dominant limestone unit above a dominant mudstone unit.

In some places where mudstone of the Tyler tongues laterally into limestone of the Alaska Bench, the interval boundary is placed within the lower part of the limestone sequence where certain beds approximate a time-stratigraphic boundary.

Limestone of Atoka age makes up part of the Amsden Formation of central and western Wyoming. Correlatives are probably the *Round Top Member* (Division IV) of the Hartville Formation of eastern Wyoming (see footnote 1) and the *Kananaskis Formation of the Rocky Mountain Group* in British Columbia and Alberta. Unnamed limestone of Atoka age (Mamet and others, 1971), comprising part of unit D of Shannon (1961), is assigned to interval B in the Arco area of central Idaho.

The Alaska Bench Limestone is probably of Early to Middle Pennsylvanian age although paleontological evidence is weak. A specimen of *Linoproductus (sensu stricto)*, a new species described by Gordon (in manuscript) from the base of the Alaska Bench Limestone in central Montana indicates an Early Pennsylvanian (probably late Morrow) age (J. T. Dutro, written commun. to W. W. Mallory, 1962). The foraminifers *Millerella marblensis* Thompson and *M. advena* Thompson, from the lower part of the Alaska Bench, indicate a similar age (Scott, 1945; Mundt, 1956b, p. 1931, 1933); and the upper part of the Alaska Bench Limestone may be of early Atoka age (George Verville, oral commun., 1962). These fusulinids extended into Early Atoka time (Lloyd C. Henbest, written commun., 1972). A microfauna from the Amsden Formation of the northern Bighorn Mountains suggests that the Amsden is of Early Pennsylvanian age (Gorman, 1963).

**UPPER BOUNDARY OF INTERVAL B**

An unconformity marks the top of interval B. In most places it separates strata of early Atoka age or older, from strata of late Atoka or early Des Moines age. Middle Pennsylvanian strata overlying the unconformity are assigned to interval C. Where the Middle Pennsylvanian rocks have been removed by later erosion, most notably near the northern and eastern limits of Pennsylvanian rocks, overlying strata are of Middle Jurassic and, locally, Cretaceous age (pl. 12).

**THICKNESS TRENDS**

Interval B thins from western Montana, where it is as much as 265 feet thick, eastward to the Dakotas, where it does not exceed 125 feet in thickness (pl. 6A). An anomalous maximum thickness of 290 feet near Judith Gap in central Montana (fig. 64) probably is the result of local preservation of an upper part of the Alaska Bench Limestone not represented elsewhere in the vicinity. Fairly abrupt changes in local thickness seem to result chiefly from the truncation of beds at the top of the unit rather than from any appreciable differences in the depositional thickness of individual beds. Strata of this interval probably were originally uniformly deposited throughout most or all of the region, as indicated by similar stratigraphic sequences in widely separated areas.

**LITHOFAICIES TRENDS**

Interval B is composed chiefly of limestone interstratified with mudstone and is relatively uniform in composition throughout the Montana-Dakota region. Argillaceous beds are more plentiful in the upper part of the interval than in the lower part. This difference is especially evident in the Judith Gap area, which probably has the most complete section of rocks of late Morrow and early Atoka age in the region. At Judith Gap the upper part of the Alaska Bench Formation, which in most other places is not preserved or is only partly preserved, is composed of nearly 50 percent red mudstone.

Lithofacies variations in the Montana-Dakota region, shown on plate 6B, would be slight except for the differential removal of strata from the argillaceous upper part of the interval. Lateral depositional changes in facies are discernible only in a few areas. Limestone, abundant in the central part of the region, is relatively scarce and mudstone relatively more abundant in western Montana, in North Dakota, and in South Dakota.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

The large amount of limestone and the relatively minor amounts of clayey mudstone indicate that deposition was in clear water throughout most of the region. An increase of terrigenous materials both in parts of western Montana and in the Dakotas suggests land areas to the west and to the east of the region. Because the detrital components in these areas are principally fine sand, silt, and clay-size particles, both the western and the eastern sedimentary source areas probably were moderately distant or low lying or both.

The few fossils that have been found in this interval imply a shallow-water marine environment in most places. The ubiquitous red color of the mudstone imparted by abundant hematite suggests deposition within an oxidizing environment, and much of the region may have been, at times, within the intertidal zone. Locally,
in western Montana, detrital dolomite grains, possible root fillings, and vermiculite suggest deltaic deposition above the tidal zone in parts of the sequence (Strickler and Zeisloft, 1965).

**PALEOTECTONIC IMPLICATIONS**

Uniformity of bedding that everywhere characterizes rocks of interval B suggests that the entire region and the adjacent areas were part of a broad shelf during most of interval B time. Continued gradual epeirogenic sinking or eustatic sea-level rise is indicated. Minor uplift of bordering land areas is implied by an upward increase of mudstone in the Alaska Bench Limestone.

Late in interval B time the shelf became less stable. Lower Atoka and underlying rocks were gently folded and possibly faulted. Erosion entirely removed rocks of the interval at some places, particularly on the crests of anticlines and along some faults (pl. 6). Instability of the shelf area during the last part of interval B time may have coincided with the beginning of a postulated geanticlinal uplift to the west. This uplift initiated deposition of sand in western Montana, and this sand now makes up the Quadrant Formation, assigned to interval C.

**INTERVAL C FORMATIONS INCLUDED**

The lower part of the Quadrant Formation forms interval C in western Montana and northwestern Wyoming (northern Yellowstone Natl. Park) (pl. 13). An upper part of the Quadrant is arbitrarily excluded from interval C and included in intervals D and E in southwestern Montana; however, the formation is poorly fossiliferous and not well dated in that area, and a Late Pennsylvanian age for the upper part of the formation is unproven.

The Rocky Mountain Quartzite near Trail Creek in northeastern Montana has not been paleontologically dated. It is included in interval C because of its lithologic similarity to the Quadrant Formation.

The Devils Pocket Formation and the overlying Quadrant Formation compose interval C in central Montana. An unconformity below the Devils Pocket Formation and equivalent strata is a practical lower boundary for interval C because it separates rocks that contrast lithologically and paleontologically.

The lower part of the middle member of the Minnelusa Formation of Des Moines age forms the interval in the Dakotas and northeastern Wyoming. The lower part of the Tensleep Sandstone and upper part of the underlying Amsden Formation at its type section in the northern Bighorn Mountains (Maughan and Roberts, 1967, fig. 3) make up interval C in north-central Wyoming. This usage differs from that of W. W. Mallory in some parts of Wyoming (pl. 13) where the upper part of the Amsden Formation is included by him in interval B rather than in interval C. These strata are stratigraphically equivalent to the Hayden Member (Division III) of the Hartville Formation in central eastern Wyoming. (See footnote 1.)

Rocks of Des Moines age included in interval C in central Idaho compose the middle part of unit D of Shannon (1961). The lower part of the Wood River Formation, northwest of the Arco area, also includes Middle Pennsylvanian (Des Moines) rocks according to Bostwick (1955). The lower part of the Wells Formation in southeastern Idaho is similar in lithology and age to the Quadrant Formation (Cressman, 1964, p. 27).

**STRATIGRAPHIC RELATIONS**

The age of the Devils Pocket Formation in central Montana, of the lowermost part of the Quadrant Formation in western Montana and northwestern Wyoming, and of the upper part of the Amsden Formation in the northern Bighorn Mountains of south-central Montana is equivocally late Atoka or early Des Moines. An Atoka age was assigned (Henbest, 1954, 1956; Easton, 1962) to the rocks in all these places, principally on peculiar species that were tentatively classed as *Profusulinella*. These forms from Montana and Wyoming, according to L. G. Henbest (1956, p. 59; and written commun., 1972), are more advanced than the species of *Profusulinella* from rocks of known Atoka age in other parts of the country. Question remains as to their classification and age—whether Atoka or possibly early Des Moines.

The Devils Pocket Formation has been assigned late Atoka age by George Verville (oral commun., 1963), on the basis of fusulinid studies. The lower part of the Quadrant Formation in several places in Montana is assigned a Des Moines age on the basis of fusulinids (Thompson and Scott, 1941; Henbest, 1954, 1956) and because of its continuity with the dated lower part of the Tensleep Sandstone and the Hayden Member (Division III) of the Hartville Formation in Wyoming (pl. 13).

The lower part of the middle member of the Minnelusa Formation near Rapid City, S. Dak., is considered to be of late Atoka age by George Verville (oral commun., 1963); however, its stratigraphic relations are the same as those of other strata included in interval C elsewhere in the region. This member rests unconformably upon red mudstone of Early Pennsylvanian age and includes strata of Des Moines age in its upper part.

The Quadrant Formation and the Devils Pocket Formation intertongue between western and central Montana. A change from sandstone in the Quadrant to dolomite in the Devils Pocket occurs progressively higher stratigraphically from west to east across the region. In eastern Montana, the two formations are mostly interbedded dolomite and mudstone and are indistinguishable from the lower part of the middle member of the Minnelusa Formation in North and
South Dakota and northeastern Wyoming (pl. 10F, section II-I 1').

**UPPER BOUNDARY OF INTERVAL C**

The boundary between intervals C and D in northeastern Wyoming and in adjoining parts of South Dakota is within the middle member of the Minnelusa Formation where rocks of Late Pennsylvanian age conformably overlie the Middle Pennsylvanian strata of interval C. The boundary as drawn between these intervals is an isochronous surface.

In most of Montana and in some parts of North and South Dakota, strata of Late Pennsylvanian age are absent and interval C is bounded at the top by unconformity. Overlying rocks range in age from Early Permian to Middle Jurassic and, locally, Cretaceous (pl. 12).

In a small area of southwestern Montana, the Quadrant Formation includes strata thought to be of Late Pennsylvanian age, as shown on maps of intervals D and E and on plate 10F, section II-I 1'). In this small area the upper boundary of interval C is arbitrary.

**THICKNESS TRENDS**

Interval C is as much as 2,000 feet thick in southwestern Montana where branches of the Cordilleran geosyncline project northeastward from Idaho (pl. 7A). The interval thins abruptly northeastward, and in central Montana and farther east in North and South Dakota, it generally is less than 300 feet thick (pl. 10F, section II-I 1').

In most of the region other than western Montana, sediments of interval C were deposited as a wedge that gradually thinned eastward across the shelf. Thickness irregularities, especially in central and eastern Montana, are chiefly due to postdepositional erosion, much of which probably took place in Early Permian time (Maughan, 1967, p. 132-133). Furthermore, interval C rocks were beveled and truncated northward and eastward during subsequent erosional episodes and were then covered by Middle Jurassic sediments.

Minor depositional thinning is evident where interval C rocks locally overlap and bury pre-existing highs. These highs, in the form of north-northwest-trending linear uplifts, are distributed throughout western Montana from Yellowstone National Park northwestward, in western South Dakota along the northeast edge of the Black Hills, and in Stanley County, central South Dakota. Rocks as old as Precambrian underlie Pennsylvanian rocks assigned to interval C in the central South Dakota area. The positive areas probably are the result of middle to late Atoka folding and faulting.

**LITHOFACIES TRENDS**

Fine- to medium-grained sandstone is a common component of rocks assigned to interval C in this region. The Quadrant Formation in southwestern Montana is composed dominantly of sandstone and quartzite. Dolomite beds, generally sandy, are also common, especially in the lower part of the formation, and they increase in number and thickness from western to central Montana.

Sandstone extends from southwestern Montana eastward, forming tongues that project into carbonate rock, chiefly dolomite and, in the Devils Pocket Formation, into mudstone. This intertonguing is conspicuous in the lower part of interval C in western and central Montana and in progressively higher strata farther east (pl. 10F, section II-I 1'). Thin beds of mudstone, mainly gray or green, are common in the Devils Pocket Formation, especially in eastern Montana. In a few places, red mudstone is interbedded with dolomite and sandstone in the lower part of the formation. Sandstone beds and lenses thin eastward and are absent in the lower part of the formation at most places in eastern Montana. The calcareous sandstone facies extends far to the east in southern Montana; but sandy and muddy carbonate facies predominate in central Montana (pl. 7B) where the upperchiefly sandy part of interval C has locally been removed.

Near the middle of the Powder River Basin (fig. 64) in northeastern Wyoming and adjacent parts of Montana, the upper part of interval C changes from dominantly sandstone in the west to dominantly carbonate rocks, forming part of the middle member of the Minnelusa Formation, in the east. The dolomite is mostly thin to medium bedded and is interbedded with many thin beds of gray to black mudstone. The mudstone is richly carbonaceous and radioactive, and in the Black Hills contains abundant pollen spores (G.O.W. Kremp, written commun. to C.G. Bowles, 1962). Sandstone, although rare in the lower part, is a common component of the upper part; it decreases eastward and grades into mudstone in central North Dakota and South Dakota.

A carbonate facies of the Minnelusa includes both a lower argillaceous dolomite and an upper sandy dolomite; this facies extends northward from the northern Powder River Basin into the Williston basin and southeastward into the Black Hills area. Only the argillaceous carbonate facies is preserved in eastern Montana, north of the Powder River Basin.

In most of northeastern South Dakota and in parts of western North Dakota, interval C consists mostly of argillaceous carbonate rocks and calcareous mudstones. In this area, unlike farther west, both upper and lower parts of interval C are argillaceous.

**SOURCES AND ENVIRONMENTS OF DEPOSITION**

The great thickness of sandstone in western Montana, the rapid eastward thinning, and the overlap of sandstone over carbonate rocks eastward suggest a source in a rising land area to the west or northwest. The
sandstone is well sorted and quartzitic, and is probably a multicycle deposit. Middle Ordovician (Chazy) sandstone, preserved as scattered remnants in Idaho, northeastern Washington, British Columbia, and Alberta, once may have been widespread in that region, forming the principal source for the Pennsylvanian sand. Other possible sources are Cambrian and Precambrian sandstones of the same region.

Mudstone which is abundant in North and South Dakota was probably derived from land areas to the east or southeast. Mudstone is dominant in parts of northwestern South Dakota, and an eastward increase of sandy mudstone is apparent in central and southern South Dakota.

A northern or possibly northeastern source area within the southern part of the Canadian Shield in Manitoba and Saskatchewan has been suggested for Middle Pennsylvanian sandstone in Montana and Wyoming (Mallory, 1967). The distribution pattern of the sand, however, implies that it entered the region from the west, though this pattern does not preclude a northern or northeastern source.

The kinds of fossils in the Middle Pennsylvanian rocks and the presence of dolomite, believed to be of primary or penecontemporaneous origin, indicate a marine environment of above normal salinity. Dolomite is abundant in strata of interval C, but is largely lacking in underlying strata of the Alaska Bench Limestone and other Pennsylvanian carbonate rocks of the region. Presumably, the sea covering the shelf area of the northern Rocky Mountains and Plains region was partially isolated from the open ocean. Land areas at indeterminate distances to the west, north, and east of the region, probably barred the circulation of marine water from these directions. To the southwest, entry of marine water apparently was limited by accumulation of detritus filling the Cordilleran geosynclinal trough that extended into Montana. Mountain uplifts in Colorado and elsewhere to the southeast barred the sea from the south and the southeast.

**PALEOTECTONIC IMPLICATIONS**

Most of the region apparently was subjected to slow, uniform epeirogenic subsidence during the time of interval C, but in western Montana, sinking was more rapid than in bordering areas to the east.

Branches of the Cordilleran miogeosyncline, which extended into western Montana from Idaho in very late Mississippian to Early Pennsylvanian time and had been filled with sediment during Early Pennsylvanian time, renewed active sinking in the Middle Pennsylvanian. Subsidence probably began in middle or late Atoka time and seems to have continued through Des Moines time.

The mass of sand that filled the geosyncline and spread eastward onto the subsiding shelf in Middle Pennsylvanian time indicates that uplift was taking place to the west or northwest of western Montana. This uplift, possibly a geanticline within the Cordilleran geosyncline, probably began during middle to late Atoka time and may have continued simultaneously with subsidence in western Montana during interval C time.

Conglomerate in the lower part of the Wood River Formation, of probable late Atoka or early Des Moines age, thickens northward in the Wood River area of central Idaho and is believed to indicate a nearby source area of marked relief (Bostwick, 1955, p. 948). Probably the source area for the conglomerate and quartzose sand deposited in central Idaho was the same as the source area that was responsible for sand deposited in western Montana, that is, the region to the northwest in Idaho, Washington, and adjacent parts of Canada.

**INTERVAL D FORMATIONS INCLUDED**

Rocks assigned to interval D occur in southwestern Montana and in the Dakotas, Wyoming, Idaho, and southeastern Montana; they are absent from central Montana.

In the southwestern Montana area, an upper part of the Quadrant Formation is arbitrarily assigned to this interval although it is not paleontologically dated. Above it at a few places in that area, the uppermost part of the Quadrant has been determined as probably of Early Permian age and this suggests that the formation may also include Late Pennsylvanian strata. Poorly dated cherty limestone of the Copper Basin Formation lies between rocks dated as Des Moines and those dated as Early Permian are found near Arco, south-central Idaho (Shannon, 1961); therefore, it is probably Late Pennsylvanian in age and equivalent to strata included in interval D.

In the eastern part of the region, strata of known or probable Missouri age in the middle member of the Minnelusa Formation are assigned to interval D; parts of the Minnelusa Formation in the southern Black Hills (McCauley, 1956) are well dated on a paleontologic basis as of Missouri age (Condra, Reed, and Scherer, 1940; Love, Henbest, and Denson, 1953; Henbest, 1958). The middle part of the Minnelusa is continuous with the lower part of the Meek Member (Division II) of the Hartville Formation (Foster, 1958) in southeastern Wyoming, and therefore, is assigned to interval D. In Montana, especially in those places where interval C is more than 300 feet thick, a few remnants of Missouri age strata may occur, but they are unrecognized; therefore, they are not differentiated from Des Moines rocks of interval C.

**UPPER BOUNDARY OF INTERVAL D**

Interval D contains the youngest Pennsylvanian rocks in the eastern part of the region. At most places these
rocks are unconformably overlain by the upper member of the Minnelusa Formation of Early Permian (Wolfcamp) age. At other places interval D is overlain by the Opeche Shale of Early Permian (probable late Leonard) age or by the Saupe Formation of Ziegler (1955-1956) of probable Middle Jurassic age (pl. 12).

In the southeastern half of the Black Hills and adjacent areas in South Dakota and Wyoming, a red mudstone unit about 20-50 feet thick, known as the "red marker," forms the basal unit of the upper member and helps to separate Permian from Pennsylvanian parts of the Minnelusa. The Permian strata are more terrigenous where they thin and overlap northwestward onto the Milk River uplift than elsewhere.

THICKNESS TRENDS

Although originally the thickness of interval D probably was uniform across much of the region, in many places, much of this interval was removed later. Exceptions are in part of southwestern Montana along a northeast-trending band that projects into southwestern Montana from Idaho, and in a few areas of southeastern Montana, northeastern Wyoming, and the Dakotas (pl. 8A) where interval D is overlain by interval E. Because of irregular depths of erosion at the upper boundary, variations in thickness do not accurately reflect original thickness trends.

Part or all of interval D was eroded from eastern Montana and adjacent parts of Wyoming and the Dakotas before deposition of Permian sediments. Where present in the eastern part of the region the individual beds tend to be uniform in lithology and thickness for long distances. Individual beds are thicker and the total thickness of correlative parts of the interval is slightly greater in northeastern Wyoming and northwestern South Dakota than in North Dakota; probably, the interval originally was blanketlike, thickening slightly southwards. This original southward thickening can be reasonably projected also into southwestern South Dakota and into adjacent parts of the region farther south.

In the central part of North and South Dakota, interval D was removed from the area east of the Missouri River probably in Late Triassic and Early Jurassic time, before deposition of Middle Jurassic sediments.

LITHOFACIES TRENDS

Facies relationships are difficult to interpret because of the incompleteness in preservation of interval D. Facies patterns on plate 8B are more a reflection of this irregular preservation, which emphasizes vertical differences in rock type, than they are facies changes within the rocks. In general, however, lithofacies relationships are believed to have been similar to those of interval C. Sandstone is the dominant lithology of interval D rocks in western Montana, whereas dolomite and sandy dolomite are dominant in the eastern part. Sand seems to have spread farther eastward in Missouri time than during the Des Moines; and gypsum, which is inconspicuous in Des Moines rocks, is locally plentiful in northwestern South Dakota in rocks of Missouri age. Mudstone seems to increase southeastward in eastern Montana and South Dakota.

SOURCES AND ENVIRONMENTS OF DEPOSITION

As during the preceding interval, uplift west and northwest of Montana seems to have been continuous and to have provided a source for large amounts of quartz sand. A source of mud and silt to the east and southeast of the region also seems to have persisted from Middle into Late Pennsylvanian time.

Detritus accumulated in a shallow sea which inundated the entire region well beyond the present limits of Upper Pennsylvanian strata during interval D time. Circulation of marine waters seems to have been restricted, as in interval C time, and primary or penecontemporaneous dolomite was again widely deposited. An abundance of gypsum in northwestern South Dakota suggests that this region was more completely isolated from the open sea than before and that the climate may have been warmer; consequently, evaporation from the sea was greater than before. An alternative explanation is that an incipient basin began to form in South Dakota at this time and that water of increased salinity and density, from which gypsum was precipitated, accumulated in it.

PALEOTECTONIC IMPLICATIONS

Tectonic patterns established during interval C time probably continued throughout the time of interval D. Lands to the east and west contributed sediment to this region but no evidence of any pronounced uplift nearby has been found. Southwest Montana continued to subside at a slightly faster rate than the rest of the region which, from Des Moines to Virgil time, was only slightly negative.

Gentle southward tilting of the region may have begun in Missouri time. Later, in the Pennsylvanian and in the Early Permian the broad Milk River uplift in Montana and adjacent parts of Canada was exposed to erosion (Maughan, 1966; Maughan and Roberts, 1967, p. B23).

INTERVAL E

Part of the Quadrant Formation in Beaverhead County, Mont., probably consists of rocks of Virgil age and is assigned to interval E (pl. 13; pl. 9). These strata are continuous with Virgil rocks in southeastern Idaho (pl. 9). Isolated remnants of Virgil rocks, as yet unrecognized, may occur at other places in Montana and in
North and South Dakota. Rocks of this age possibly were once distributed throughout most of the northern Rocky Mountains and Plains region, and were subsequently largely removed by post-Pennsylvanian erosion. On the other hand, because this area was located on the south flank of the Late Pennsylvanian or Early Permian Milk River uplift (Maughan, 1966; Maughan and Roberts, 1967), by Virgil time uplift may have been sufficient to cause the Pennsylvanian sea to withdraw from most of the region; thus, no interval E deposits would have accumulated there.

**TOTAL THICKNESS OF PENNSylvanian ROCKS**

**THICKNESS TRENDS**

The thickest Pennsylvanian rocks known in the region total about 4,430 feet. They are in the southern Tendoy Mountains (fig. 64) in southwestern Montana near the Idaho border. Sediments in this area were deposited almost continuously from Early Pennsylvanian into Permian time in a northeast-trending trough that projected into central Montana from Idaho and in a similar, but smaller and less well defined, northeast-trending trough extending into the Pioneer Mountains of Montana. The trough in the Pioneer Mountains contains Pennsylvanian rocks as much as 950 feet thick, northwest of and parallel to those in the deeper trough.

In central Montana, rocks of Pennsylvanian age are as much as 970 feet thick on the northwest flank of the Little Belt Mountains, but they thin eastward in the Big Snowy Mountains and from there farther east in Montana and North and South Dakota. Deposits of thickness comparable to those in the Little Belt Mountains probably were once extensive in central and north-central Montana.

Pennsylvanian rocks tend to be thin along an east-northeast-trending band just south of the Big Snowy Mountains in central Montana, and along a north-northwest-trending band approximately coincident with the Cedar Creek anticline in eastern Montana. Shallow depositional basins are indicated by slightly thicker Pennsylvanian rocks in northeastern Wyoming and in the Williston basin, east of the Cedar Creek anticline.

Post-Pennsylvanian erosion has significantly modified the thickness of Pennsylvanian rocks in the region. Some beveling of Pennsylvanian rocks took place before deposition of Lower Permian (Wolfcamp) rocks; subsequent erosion locally removed additional Pennsylvanian rocks before deposition of the younger Permian and Triassic rocks. These erosional episodes account for the present-day position of the zero isopach in northern Montana and North Dakota. Middle Pennsylvanian beveling has also produced minor thickness variations, especially in south-central and southeastern Montana—including the Black Hills region—and in an area of thin to absent Pennsylvanian rocks in the vicinity of Stanley County, near central South Dakota.

**PALEOTECTONIC IMPLICATIONS**

Two regional tectonic disturbances during the Pennsylvanian affected the distribution and thickness of Pennsylvanian rocks in Montana and North and South Dakota. Several post-Pennsylvanian tectonic events additionally caused Pennsylvanian rocks to be beveled and partially or totally removed in some places.

The earliest widespread tectonic disturbance of this region followed deposition of the Upper Mississippian Big Snowy Group. North-central Idaho and western Montana are believed, on sparse evidence, to have been uplifted either in a broad antitcline or in fault blocks, whereas southwestern Montana was downfolded into shallow troughs or was cut by grabens. Some broad, very low uplifts and shallow basins also formed on the shelf farther east. Among these eastern structural features were an incipient Williston basin, a low platform in south-central and southeastern Montana, and a shallow basin in north-central Montana.

About middle Atoka time, many older faults and folds seem to have been reactivated. The activity was more intense than it had been previously in western Montana, but was less intense in the central and eastern parts of the region. Uplift seems to have continued into Middle Pennsylvanian time in north-central Idaho and in areas farther north, and a geanticline may have formed within the northern part of the Cordilleran miogeosyncline. Troughs in western Montana continued to sink, and epeirogenic downwarping characterized the shelf farther east. This geanticlinal uplift furnished much of the sand deposited in the subsiding troughs of southwest Montana, and the sand extended as a wedge onto the adjacent shelf in eastern Montana and Wyoming during Middle Pennsylvanian (Des Moines) time.

Late Pennsylvanian to Early Permian uplift centered chiefly about the Milk River positive area in north-central Montana. Pennsylvanian rocks were totally removed from this uplift and partly removed in adjacent areas before deposition of overlapping Lower Permian sediments.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSylvanian SYSTEM**

**UNITS OVERLYING PENNSylvanian**

Rocks overlying the Pennsylvanian range in age from Early Permian to Cretaceous. Near the Beaverhead and Tendoy Mountains (fig. 64) in southwestern Montana, Lower Permian strata lie conformably on the Pennsylvanian within the upper part of the Quadrant Formation, but elsewhere the overlying strata are separated from
Pennsylvaniaian rocks by an unconformity. In general, overlying units are increasingly younger to the northeast in southwestern Montana and to the north and northwest in North and South Dakota and eastern Montana.

The Meade Peak Phosphatic Shale Member of the Phosphoria Formation of Leonard age and equivalent strata of the Shedhorn Sandstone locally lie unconformably upon Pennsylvaniaian rocks of the Quadrant Formation of southwestern Montana. Near Three Forks, also in southwestern Montana, the Grandeur Member of the Park City Formation of probable late Wolfcamp age (Frenzel and Mundorff, 1942; J. S. Williams, in McKelvey and others, 1959, p. 36) forms the base of the overlying Pennsylvaniaian rocks (Sheldon and others, in McKee, Oriel and others, 1967, p. 154).

In South Dakota and adjacent parts of Wyoming, Montana, and North Dakota, the upper member of the Minnelusa Formation of Early Permian (Wolfcamp) age unconformably overlies Pennsylvaniaian rocks. A red mudstone unit about 20-50 feet thick, known as the “red marker,” forms the basal unit of the Minnelusa’s upper member and helps to separate Permian from Pennsylvaniaian parts of the formation in much of this area. The Permian strata are increasingly terrigenous northwesternward where they thin and overlap onto the Milk River uplift.

The upper (Permian) member of the Minnelusa Formation is progressively more beveled northward and so in the Williston basin it is absent and the Opeche Shale of probable late Leonard age overlies Pennsylvaniaian rocks. Around the southeast edge of the Milk River uplift of central Montana, Pennsylvaniaian rocks are overlain by Upper Permian (upper Guadalupian Ochoa (?) equivalents) of the Park City Formation and Lower Triassic of the Dinwoody Formation in southwestern Montana, by the equivalent part of the Goose Egg Formation in Wyoming, and by the equivalent lower, Permian, part of the Spearfish Formation in part of southeastern Montana. Around the periphery of the Milk River uplift these formations are beveled and truncated by erosion that took place before deposition of Middle Jurassic rocks, and Middle Jurassic rocks overlie Pennsylvaniaian rocks north of the truncated edge of Permian and Triassic units.

Jurassic rocks that overlie the Pennsylvaniaian are progressively younger from the Williston basin southward into central South Dakota. In this region Pennsylvaniaian rocks are overlain in turn by the Saude Formation of Ziegler (1955-1956), the Ness Formation of Nordquist (1955, p. 104-106), the upper part of the Sundance Formation, and possibly the Morrison Formation. Progressively younger Jurassic rocks, comprising the Saude Formation, the Gypsum Spring or Ness Formation, the Ellis Group, and the Morrison Formation, overlie Pennsylvaniaian rocks from the Williston basin westward in eastern and central Montana.

**PALEOTECTONIC IMPLICATIONS**

Uplift of all but possibly southwesternmost Montana ended deposition in the region by Late Pennsylvanian or earliest Permian time. The uplift was probably epeirogenic, but was greatest in north-central Montana and may have included faulting in an arcuate band in central Montana around the south side of the Milk River uplift where Pennsylvaniaian strata are truncated. The age of the faulting and folding is not accurately known; however, it may have been as late as Early Jurassic.

Early Permian (Wolfcamp) seas inundated an area peripheral to the Milk River uplift from the midcontinent to the southeast and from the Cordilleran geosyncline to the southwest. The uplift remained land and was the principal source of sediments deposited in the Williston basin, on the Wyoming shelf, and along the eastern margin of the Cordilleran geosyncline during Early Permian time.

Renewed epeirogenic uplift late in Early Permian (Leonard) time caused the sea to withdraw from the entire Montana-Dakota region. Permian and Pennsylvaniaian strata were eroded from a broad area in southern Montana and adjacent parts of Wyoming and North and South Dakota. Some faulting occurred at this time in eastern Montana along north-northwestern structural elements such as the Cedar Creek anticline and possibly in western Montana in a zone between Livingston and Helena. Subsequently, the sea readvanced into central Montana where apparently it remained until Early Triassic time. All of Montana, including possibly the Milk River positive area, and most of the Dakotas were covered; however, the sedimentary record has been largely destroyed by later uplift and erosion.

Late Triassic and Early Jurassic were times of uplift; and older rocks, including Pennsylvaniaian rocks in some places, were eroded and beveled to a northeastward-sloping plain of low relief. In the Late Jurassic, a sea in which the Jurassic Sundance Formation was deposited transgressed from the northeast. “Belt Island” near the present-day Big Belt Mountains of west-central Montana, the Sheridan arch in Johnson and Sheridan Counties, northern Wyoming (Peterson, 1954, p. 474), and a low uplift in central South Dakota remained land during the early part of this transgressive episode, but eventually were covered.

Final important tectonic events in Cretaceous and early Tertiary time that brought Pennsylvaniaian rocks to the surface and exposed them to erosion were uplift associated with emplacement of the Idaho and Boulder batholiths, and thrusting and normal faulting in western Montana, and folding in the eastern part of the region associated with the Laramide orogeny.
REFERENCES


1961, Tyler sand trend becoming better known in Dakotas: World Oil, v. 152, no. 1, p. 89-93.


Montana, North Dakota, Northeastern Wyoming, and Northern South Dakota

Geology, geochemistry, mining: Cleveland, Ohio, Northern Ohio Geol. Soc., p. 35-47.


Petsch, B. C., 1949, North part of the Whitewood anticline [S. Dak.]: South Dakota Geol. Survey Rept. Inv. 65, 30 p.


Thompson, M. L., and Scott, H. W., 1941, Fusulinids from the type sections of the lower Pennsylvanian Quadrant formation [Wyo.]: Jour. Paleontology, v. 15, no. 4, p. 549-553.


Arizona

By EDWIN D. MCKEE

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-P
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES,
PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

ARIZONA

By EDWIN D. MCKEE

ABSTRACT

The paleotectonic record of the Pennsylvanian System in Arizona involves repeated advances and retreats of the sea from three principal directions. Marine deposits accumulated on the floors of shelf seas; and continental sediments, mostly red beds, were deposited on adjoining coastal plains and deltas. Most land areas were low and nearly flat. The Defiance-Zuni positive element in the northeastern part of the State persisted as a landmass throughout the period, but contributed little detrital sediment; another positive element probably existed in the southwestern part of the State, but the record of any contribution of sediment from this source is obscure.

At the time that Pennsylvanian sediments began to accumulate in Arizona, the surface of the region was largely carbonate rock of Mississippian age. Only in a few places, notably on the Defiance positive element and around its margin are rocks older than Mississippian known to have been exposed. On the limestone-dolomite surface a karst topography with abundant sinkholes, caverns, and collapse features was developed, especially in the central Arizona area. Red regolith, insoluble residue, and solution breccia that covered the surface across wide areas indicate that a once much greater thickness of limestone was largely removed by solution before Pennsylvanian deposition. An extensive but small vertical movement apparently was responsible for this work of solution.

The earliest Pennsylvanian strata to be deposited in Arizona were in the northwestern and southeastern areas during interval A (Morrow) time. The maximum thickness of interval A is about 400 feet in the northwestern part of the State and 300 feet in the southeastern. These sediments apparently accumulated in restricted areas that did not connect across central Arizona. Only slight uplift or perhaps warping occurred at this time.

During interval B (Atoka) time, depositional areas still were much restricted; one extended southwestward into Arizona from a large basin in Utah and Colorado and another northwestward into Arizona from the Sonoran geosyncline, but they did not connect. Rocks of interval B, unlike those of interval A, were not deposited in the northwest corner. Thickness of sediment in places was in excess of 500 feet, indicating moderate downwarping, but evidence of more severe crustal disturbance is lacking.

Interval C (Des Moines time) represents a time of maximum transgression of the seas. Nearly 1,000 feet of strata accumulated in parts of the southeast region. During deposition of interval D (Missouri time), similarly thick and widespread accumulations formed in the southeast; but, elsewhere, deposition largely or entirely stopped.

In interval E (Virgil time), which was the final part of the Pennsylvanian, depositional areas that centered mainly in the northwest and southeast were connected, and some sediment accumulated in northeastern Arizona, so most of the State was below the base level of deposition. After Pennsylvanian time the same general pattern of deposition was resumed in the Permian. A time break and minor erosion are recorded over extensive areas; in some places, however, deposition was continuous and as a result Lower Permian rocks rest conformably upon Upper Pennsylvanian.

REGION DEFINED

The region discussed in this section of the Pennsylvanian Paleotectonic Map text is restricted to the State of Arizona; however, many of the paleogeographic and paleotectonic features that are referred to were largely or partly outside of Arizona and they straddled the present international and State boundaries. Strata of Pennsylvanian age, partly continental and partly marine, were deposited across most parts of the State at one time or another during the period. Transgressions and regressions of the sea were mostly across low flat land surfaces and adjacent shelves, which were marginal to the Cordilleran and Sonoran geosynclines to the northwest and southeast, respectively, and to the Paradox basin on the northeast (fig. 65). The Defiance-Zuni positive element on the New Mexico border was partly within the State, but probably the positive element was not high at any time during the Pennsylvanian.

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

Rocks of the Pennsylvanian System in most parts of Arizona rest upon strata of Mississippian age. A notable exception is on the flanks of the Defiance positive element in the northeastern part of the State. Here the Pennsylvanian rests mostly on Precambrian and, locally, on Devonian (pl. 2).

A hiatus of considerable magnitude is at the base of the Pennsylvanian almost everywhere in the State, including most places where Mississippian beds are below the Pennsylvanian. Only in extreme southeastern Arizona (Armstrong, 1962, p. 24) and in two widely separated localities farther north (Mckee and
For distances of many miles in the Grand Canyon area of northwestern Arizona, the Pennsylvanian-Mississippian boundary is a continuously exposed disconformable contact of the Supai Formation on the Redwall Limestone (McKee and Gutschick, 1969, p. 74). From a broad view, the contact is flat and even, but in detail the Redwall surface is actually very irregular and has a relief of as much as 30 feet along the walls of channels, sinks, and remnant platforms or hills. In the vicinity of the East Kaibab monocline of eastern Grand Canyon, an angular unconformity is reported, where gently dipping beds of Redwall were beveled prior to Supai deposition.

In northeastern Arizona, stratigraphic relations at the Mississippian-Pennsylvanian contact are much like those in the Grand Canyon area farther west, except that the Pennsylvanian sea in northeastern Arizona advanced across that area at a somewhat later time and from a different direction. Fossils in the basal Supai deposits of Grand Canyon are considered to be of Morrow age, and the sea advanced eastward (McKee, unpub. data), whereas faunas from the Mol a Formation at the base of the Pennsylvanian sequence of northeastern Arizona indicate an age ranging from Atoka to Des Moines, with the younger forms occurring in the more southwestern deposits (Peterson, 1959, p. 506). The Mol a Formation has been examined by many geologists, and a comprehensive summary of these investigations has been presented by Merrill and Winar (1958). The lower part of the formation is generally considered to be a residual deposit accumulated in place through solution of the underlying carbonate rocks; the upper part is clastic debris, at least partly derived from the regolith but reworked and redistributed by streams and an advancing sea.

In the Chiricahua Mountains of southeastern Arizona, a disconformity between the Paradise Formation, the upper part of which is of Chester age, and the Horquilla Limestone, the lower part of which is of Morrow age, marks the Mississippian-Pennsylvanian boundary (Ross and Sabins, 1965, p. 176). Farther west, in the Little Dragoon Mountains (Gilluly and others, 1954), Whetstone Mountains (Tyrrell, 1957), and elsewhere, the Black Prince Limestone, shown to be of Morrow or Atoka age by Nations (1961, 1963), overlies strata of Meramec age in the upper part of the Escabrosa Limestone. Thus, from southeast to northwest or in the direction that Pennsylvanian seas transgressed, the magnitude of the hiatus increases. The upper surface of the Escabrosa is slightly undulating; it contains holes and small pits and is covered, in places, with chert-pebble and chert-cobble conglomerate (Tyrrell, 1957, p. 48, 49). The detrital sediment, 15–27 feet thick, is considered to be an insoluble residue (Gilluly and others, 1954; Nations, 1963, p. 1255).

In most of east-central Arizona, the Naco Formation of

---

LOWER BOUNDARY OF PENNSYLVANIAN

Strata underlying the Pennsylvanian rocks of Arizona are largely bedded of Mississippian age on which is a karst surface characterized by many sinkholes, abundant collapse features, numerous caverns, and various other products of solution. The overlying basal Pennsylvanian deposits commonly consist of red regolith, solution breccia, and accumulations of insoluble residues though, locally, normal conglomerates or beds of finer transported materials cover the uppermost Mississippian strata.

---

**Figure 65.** General tectonic provinces in the Arizona region during Pennsylvanian time. Position of Supai delta (Pennsylvanian) and Butte fault (Precambrian) also shown.

Gutschick, 1969, p. 74) are rocks of Chester age, representing youngest Mississippian time, known to occur. Elsewhere, the Mississippian strata are much older, being of early Meramec age (McKee and Gutschick, 1969, p. 74, 80).

The oldest of Pennsylvanian rocks in Arizona, located in the southeast and northwest areas, are of Early Pennsylvanian age (pl. 3A) as shown by determination of fossils from rocks in these areas (Ross and Sabins, 1965, p. 177; McKee, unpub. data). Thus, a stratigraphic break involving only uppermost Mississippian and lowermost Pennsylvanian time probably is represented in areas where the hiatus is least.
Pennsylvania aged overlie the Redwall of Mississippian age. The basal Naco in most of this area is early Des Moines in age and is younger than the lowest Pennsylvaniaan strata in southern Arizona, which are of Morrow age (Brew, 1965). Thus, transgression of the Pennsylvanian sea from southeastern to central Arizona took considerable time, during which a maximum amount of residual material developed in central Arizona.

The contact between the Redwall Limestone and the Naco Formation in central Arizona commonly is placed above the highest continuous ledge of gray limestone, and beneath either a dark red-brown chert rubble and mudstone mixture or an undisturbed mudstone residuum (Brew, 1965, p. 28). In some places, extensive solution has caused the boundary to appear gradational. Broken, disarticulated rubble in the uppermost few feet of the Redwall is infilled with reddish-brown mudstone, and occasional large blocks of Redwall are embedded in residuum (McKee and Gutschick, 1969, p. 84). This residual material may have accumulated during latest Mississippian time or the early part of Pennsylvanian time, or both. In central Arizona the entire Horseshoe Mesa Member of the Redwall has been destroyed and its residue either partly or, in places, entirely removed.

For some areas, reasonably good estimates are possible for the amount of Mississippian rock removed during this interval of erosion. Remnants or residual boulders of Mississippian rocks containing datable fossils are locally preserved. Reconstruction of the extent of originally deposited Mississippian rocks throughout central and southern Arizona has been undertaken by Brew (1965, fig. 3).

In southern Arizona, Gilluly, Cooper, and Williams (1964) have estimated, on the basis of the amount of insoluble residue accumulated in basal beds of the Black Prince Limestone, that "tens and possibly hundreds of feet of the underlying Escabrosa" have been removed. In central Arizona, Sando (1964, p. C41) found the coral genus Aulina, of probable Meramec or Chester age, preserved in residuum of the Redwall, which demonstrates that Redwall strata much younger than those at the present surface once covered the area. Near Bright Angel trail in Grand Canyon, a small remnant of strata belonging to the Redwall and of Chester age (the latter determined by brachiopods and fusulinids) shows that rocks of Late Mississippian age once covered much of this area, even though now the uppermost Redwall through most of the Grand Canyon ranges from late Osage to early Meramec age. Thus, compelling evidence shows that deposition occurred throughout large parts of Arizona until late in Mississippian time.

**PALEOTECTONIC IMPLICATIONS**

The stratigraphic relations at the basal Pennsylvanian unconformity indicate clearly that a small vertical movement affected a wide area. Much, perhaps most, of the erosion that followed uplift was in the form of solution, as indicated by widespread karst topography and, in many places, the appreciably thick residual material accumulated in situ. In some areas, however, through-flowing streams introduced gravels of rock types that apparently came from distant points and thus indicate moderate gradient in some of the channels (McKee and Gutschick, 1969, p. 76). The amount of rock removed and the degree of karst development seem to have been greater in central Arizona than in northern Arizona and probably greater also than in southern Arizona, as indicated by both the thickness of strata known to have been eroded and the local relief on the post-Redwall surface.

At only one place, the eastern part of Grand Canyon, has evidence of angular discordance between strata of the two systems been reported (McKee and Gutschick, 1969, p. 80, fig. 31e). A beveled surface on gently dipping strata of the Redwall at Tanner trail (fig. 66) indicates warping followed by erosion along a belt that had been much earlier disturbed by movement along the Precambrian Butte fault and that was later deformed by development of the Tertiary East Kaibab monoclone.

**INTERVAL A**

**FORMATIONS INCLUDED**

Interval A is the most restricted of the Pennsylvanian intervals in Arizona (pl. 34). It is represented (pl. 13) by the lowest part of the Supai Formation in the Grand Canyon area of northwestern Arizona, by the basal part of the Horquilla Limestone of the Naco Group in southeastern Arizona (Ross and Sabins, 1965, p. 177), and by the Black Prince Limestone farther west in southeastern Arizona (Nations, 1963, p. 1259). Elsewhere in the State, no fossils of Early Pennsylvanian age have been reported.

**STRATIGRAPHIC RELATIONS**

(NOTE: The discovery, after preparation of this report, of invertebrate fossils, including fusulinids and brachiopods, extensively distributed in the Grand Canyon region throughout the rocks called Supai has greatly clarified the age relations of various rock units in that formation, but it necessitated a revision of certain age assignments that were used in preparing maps for this publication. Principal changes that affect the Grand Canyon region isopach and lithofacies maps are summarized by McKee (1975). They include a reduction in thickness of interval A, an increase in thickness of interval B, and near or complete elimination of interval C, but little or no change in interval E.)

In the northwestern part of Arizona, assignment of basal Supai strata to interval A is partly based on
brachiopods of Early Pennsylvanian age (R. E. Grant, written commun., 1964) from Quartermaster and Twin Springs Canyons, both in western Grand Canyon (fig. 66). In adjacent parts of Nevada to the west, foraminifers in the lower part of the lithogenetically equivalent Callville and Bird Spring Formations likewise are assigned a Morrow age (Welsh, 1959, p. 56a, 56b).

In the Four Corners area of northeastern Arizona, no rocks of interval A have been recognized. The basal Pennsylvanian formation in that area, known as the Molas, is generally regarded as a residual red soil developed on a karst surface of Mississippian age and reworked in its upper part by an advancing sea (Peterson, 1959, p. 506). Although in parts of Colorado this formation is considered to be of Morrow age, available evidence indicates a largely or entirely Atoka age in Arizona. Many fusulinid collections from the subsurface of the Four Corners area have been studied by John Chronic (written commun., 1964) of Boulder, Colo. A tabulation of species determined by him shows no Morrow forms from northeastern Arizona or from the adjoining counties of Utah and New Mexico. The oldest
determinable fossils from Pennsylvanian rocks in this area are all of Atoka age (table 33).

TABLE 33. — Faunal assemblages from Pennsylvanian well samples in Four Corners area showing distribution and age assignments by John Chronic

<table>
<thead>
<tr>
<th></th>
<th>Morrow</th>
<th>Atoka</th>
<th>Des Moines</th>
<th>Missouri</th>
<th>Virgil</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northeastern Arizona</td>
<td>1</td>
<td>2</td>
<td>5</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>New Mexico, San Juan County</td>
<td>0</td>
<td>5</td>
<td>9</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Utah, southern San Juan County</td>
<td>0</td>
<td>2</td>
<td>19</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Utah, middle San Juan County (T29S-T30S)</td>
<td>0</td>
<td>1</td>
<td>9</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>Utah, northern San Juan County (T27S-T31S)</td>
<td>0</td>
<td>4</td>
<td>10</td>
<td>3</td>
<td>6</td>
</tr>
</tbody>
</table>

*Age assignment uncertain.

The presence of interval A in southeastern Arizona was established by the discovery of Morrow fossils in several desert ranges near the New Mexico border (Sabins, 1957, p. 486). Farther westward in southeastern Arizona, the interval is composed of the Black Prince Limestone, considered to be of Late Mississippian or Early Pennsylvanian (?) age by Gilluly, Cooper, and Williams (1954), but it was later shown by Nations (1963, p. 1259) on the basis of its coral and fusulinid fauna to be of Morrow age.

UPPER BOUNDARY OF INTERVAL A

In northwestern Arizona, the upper boundary of interval A is arbitrarily defined. It must be somewhere between the lower middle cliff units of the Supai Formation in western Grand Canyon. Fossils considered to be of Morrow age occur at the top of the lowest limestone cliff at Quartermaster Canyon (fig. 66) and foraminifers of post-Morrow age (L. G. Henbest, written commun., 1964) occur in the middle cliff unit at Hidden Canyon. A boundary that is readily mappable is the top of the middle slope unit and this horizon is used to separate intervals A and C; interval B apparently is missing in this area.

Where the Black Prince Limestone is present in southeastern Arizona, the contact between it and the overlying Horquilla Limestone is considered to be the boundary between intervals A and B. Marked differences in lithology occur with carbonate rock in interval A and interbedded carbonate and mudstone in interval B; distinctive faunas occur in these formations. The Black Prince is characterized by the *Millerella* zone; the basal part of the Horquilla by the *Profusulinella* zone in the east and the *Fusulinella* zone in the west (Ross and Tyrrell, 1965, p. 619; Ross and Sabins, 1965, p. 177).

THICKNESS TRENDS

Interval A, as recognized in northwestern Arizona, attains a maximum thickness of more than 400 feet near the Nevada border. It thins from this area south and east toward a zero line roughly one-third of the distance across the State. The rate of thinning and the position of the zero line are subject to question because they are determined by projecting the upper boundary from Nevada eastward into areas in which there is no fossil control.

In southeastern Arizona, a maximum thickness of about 300 feet has been reported for Morrow rocks near the New Mexico border (Sabins, 1957, p. 486). Sixty to 80 miles farther west, thicknesses ranging from 100 to 200 feet are assigned to the Black Prince Limestone. In that area, as in northwestern Arizona, the limits of Morrow rocks are not yet well known. At localities along the Mexican border, the Black Prince is missing and beds of Atoka age in the Naco Group rest on Mississippian strata, as has been noted by P. T. Hayes (oral commun., 1964). These data indicate southern limits of interval A; but to the north and west, as shown on plate 3A, control points are too few and widely separated to establish precise limits for the interval. In general, the basin of deposition is similar to, but far more restricted than, the basins of subsequent intervals of the Pennsylvanian System.

LITHOFACIES TRENDS

In two areas where interval A was deposited in Arizona — the northwest and southeast corners — the lithofacies are typical of shelf deposits. Interval A in the northwestern area has more than 80 percent carbonate rock along the Nevada border, but this amount progressively decreases eastward, and the proportion of mudstone increases in the same direction. In the area of eastern Grand Canyon, more than 80 percent red shaly mudstone alternates cyclically with about 20 percent medium-bedded limestone (pl. 3B). Interval A in the southeastern Arizona area, in contrast, contains only carbonate rock.

SOURCES AND ENVIRONMENT OF DEPOSITION

The cyclic alternation of beds, particularly of red shaly mudstone and medium-bedded limestone, and a fauna of large mollusks and brachiopods in the limestones suggest that in northwestern Arizona the depositional environment was alternately terrestrial and near-shore shallow-water marine. Derivation of the mudstone from the east is suggested by the lithofacies trend and, especially, by a tongue of relatively pure mudstone in the vicinity of eastern Grand Canyon that thins and narrows eastward (pl. 3B). The Cordilleran seaway was mainly to the west in Nevada. The source must have been pre-Mississippian rocks exposed rather far to the east or northeast, inasmuch as mud could not have been derived from carbonate rocks of the underlying Mississippian Redwall Limestone.

Deposits that accumulated in southeastern Arizona
Paleyotectonic Implications

Interval A represents a time of minor crustal disturbance in Arizona as shown by the limited extent and thickness of the rocks that were deposited at this time. The northernmost end of the Sonoran geosyncline extended from Mexico into southeastern Arizona, and was slightly negative. Small uplift, possibly accompanied by gentle warping, is indicated in central–northern Arizona by a slight angular discordance between Pennsylvanian and Mississippian strata at Tanner trail (fig. 66) in eastern Grand Canyon.

Interval B

Formations Included

Interval B extends a short distance into northeastern Arizona (pl. 6A). In that area it consists of part of the deposits that were formed in the extensive Paradox basin, which was mostly in Utah, Colorado, and New Mexico to the north, northeast, and east, respectively. These deposits are basal Pennsylvanian and are referred to the Molas Formation and to the basal 100 feet of the Lower Hermosa.

Rocks assigned to interval B extend across most of southeastern Arizona and form the lower part of the Horquilla Limestone (pl. 13). Farther north, in east-central Arizona, where these rocks constitute the thin northern edge of strata of Atoka age, they form the lowest part of the Naco Limestone.

Stratigraphic Relations

Rocks of Atoka age are unknown in northwestern Arizona, though extrapolation of poorly controlled isopachs from adjoining States suggests that rocks of that age formerly may have extended into the extreme corner. Fossils at several localities indicate that Des Moines rocks lie directly on those of Morrow age; Atoka beds have not been recognized in between. In southeastern Nevada, a similar relationship exists within the Callville Limestone. Farther west in Nevada, in much thicker sections, fusulinids indicate a stratigraphic sequence that includes equivalents of all the midcontinent series of the Pennsylvanian (Welsh, 1959, p. 56a).

Fusulinids studied by John Chronic (written commun., 1964) show the Molas Formation and Lower Hermosa Formation in this area to be of Atoka age. Diagnostic fossils are recorded from outcrops at two localities in Arizona and from wells (table 33) in adjoining parts of San Juan County, N. Mex., and San Juan County, Utah.

In east-central Arizona, assignment of an Atoka age to the rocks referred to interval B is largely based on fusulinids reported by H. R. Wanless (written commun., 1964). The lower boundary of the interval is placed at the contact between the Naco Formation and the Escabrosa or the Redwall Limestone.

Assignment of rocks in southeastern Arizona to interval B is largely based on fusulinids in the lower part of the Horquilla Limestone which are considered to be of Atoka age (H. R. Wanless, written commun., 1949; Tyrrell, 1957; Sabins, 1957, p. 466).

Upper Boundary of Interval B

The upper boundary of interval B in northeastern Arizona is arbitrarily placed in the lower part of the Lower Hermosa, between the positions of fossils dated as late Atoka and of those determined to be early Des Moines. In stratigraphic sections near Four Corners, this boundary is approximately 100 feet above the base of the Lower Hermosa. Farther south and west, where fusulinids have not been found and where thinner sections of Pennsylvanian rocks cannot readily be subdivided, this boundary is not recognizable. Atoka strata are assumed to have pinched out.

In southeastern Arizona, the upper boundary is placed to include the highest Atoka fossils in interval B and to exclude the lowest Des Moines fossils. It can be located within relatively narrow limits at numerous widely scattered localities (for example, Sabins, 1957, p. 486; Ross and Tyrrell, 1965, p. 620; Ross and Sabins, 1965, p. 177).

Thickness Trends

Southward and southwestward from the northeast corner of Arizona, interval B decreases in thickness from nearly 300 feet to a thin edge in approximately 50 to 60 miles. The pattern of isopachs in this area (pl. 6A) and the trend in rock thickness indicate that these strata are marginal to thicker deposits in the Paradox basin to the northeast.

Interval B extends across most of southeastern Arizona and into east-central Arizona (pl. 6A). Within this area, its maximum known thickness, 640 feet (Epis, 1956), is in the Pedregosa Mountains near the border of New Mexico. To the northwest, at Superior, Coolidge dam, and Holy Joe Peak, thicknesses of 51, 81, and 200 feet, respectively, have been determined, and farther north at Salt River crossing and near the junction of the Black and White Rivers, measured thicknesses are 42

1According to D. C. Brew (written commun., 1969) the oldest fusulinid-bearing beds at Black River crossing are probably earliest Des Moines and not Atoka in age.
The zero isopach at the northern limit of interval B in east-central Arizona probably is near the Mogollon Rim (fig. 66); in several stratigraphic sections along the rim, rocks of Atoka age seem to be absent. Control is insufficient to determine precisely the location of this zero line, but its approximate position seems clear.

**LITHOFACIES TRENDS**

Lithofacies data for interval B (pl. 6B) show a progression from carbonate rock to mudstone in the direction of thinning of the interval in both northeastern and southeastern Arizona. Within a few tens of miles westward and southward from the center of the Paradox basin — northwest of Four Corners, for example — the lithofacies composition changes from more than 50 percent carbonate rock to less than 20 percent, and the percentage of mudstone increases correspondingly. Likewise in southeastern Arizona, a definite change in facies occurs from mostly carbonate rock in the east to more than half mudstone in the west and northwest.

**SOURCES AND ENVIRONMENT OF DEPOSITION**

Although rocks of interval B in Arizona may well have once extended beyond their present limits, which are indicated by the zero isopachs (pl. 6A), existence of a very limited seaway during Atoka time seems probable in view of the large area from which interval B is absent. Furthermore, there seems little likelihood that waters of the Paradox basin in the north ever had a direct connection with those of the Sonoran geosyncline to the south; this barrier is indicated by depositional thinning of the interval toward the intermediate area.

A fairly small percentage of sand on the lithofacies map for this interval suggests that regions near the seaways in Arizona were not greatly elevated. A high percentage of mudstone in the western part of the basin to the northeast and a moderately high percentage along the western margins of the seaway to the southeast suggest sources of fine detrital material to the west or northwest of those areas.

**PALEOTECTONIC IMPLICATIONS**

Interval B apparently represents a time of little crustal disturbance in Arizona and was marked only by slight sinking of areas marginal to the Paradox basin and Sonoran geosyncline; no evidence of sinking marginal to the Cordilleran geosyncline is evident. The region was devoid of any major uplift.

**INTERVAL C FORMATIONS INCLUDED**

Rocks in northwestern Arizona that have been assigned to interval C (pls. 7A, B) consist of the lower middle part of the Supai Formation. Fossils probably of Des Moines age have been found in these rocks and fossils definitely of Des Moines age occur in strata considered equivalent not far to the west in Nevada.

In most wells in the Four Corners area of northeastern Arizona, rock units believed to be of Des Moines age are, in ascending order, the Lower Hermosa (except the basal 100 feet), the Desert Creek, and the Isumay (pl. 13). The age of these units is reasonably well established by fossils. The overlying Upper Hermosa is generally dated as Missourian or Virgil, or both, but may include rocks of latest Des Moines age and of Wolfcamp age.

Rocks of Des Moines age in east-central Arizona, assigned to interval C, compose a part of the Naco Limestone. In southeastern Arizona, they are a part of the Horquilla Limestone.

**STRATIGRAPHIC RELATIONS**

In the western Grand Canyon area, red beds and carbonate rocks occur in the Supai Formation between rocks that contain Morrow fossils assigned to interval A and rocks near the top that contain a Permian Wolfcamp fauna. In Hidden Canyon (fig. 66), the Pennsylvanian strata above interval A contain foraminifers that range in age from Des Moines to Virgil. Strata that are in similar stratigraphic position in the Callville and Bird Spring Formations in adjacent parts of Nevada also contain fusulinids of Des Moines age in the lower beds and Virgil in the upper beds; none of Missouri age are present.

The thickness of Des Moines rocks compared to the thickness of Virgil rocks in Nevada is in the proportion of 3 to 2 (Welsh, 1959, p. 56a); post-Morrow, pre-Wolfcamp beds in the Supai Formation are arbitrarily assigned to intervals C and E in the same proportion. The lower limit of this interval C–interval E sequence is placed at the top of the second slope unit in western Grand Canyon, and the top of this sequence is marked by a conglomerate that underlies the Esplanade cliff throughout the area.

**UPPER BOUNDARY OF INTERVAL C**

In northeastern Arizona, the upper boundary of interval C is easily determined, for it is placed at the base of the Upper Hermosa. The contact is a lithologic break between limestone and mudstone that normally can be recognized without much difficulty, even in the subsurface.

Along the Mogollon Rim (fig. 66) and south of it in east-central Arizona, the approximate upper boundary for interval C, based on fusulinids, has been determined by H. R. Wanless (written commun., 1960) for eight localities. Interval C has a thickness about equal to the thickness of overlying parts of the Pennsylvanian System at these localities and this thickness relation has then been applied in calculating interval thicknesses and
boundaries in the subsurface farther north (Little Colorado area).

In southeastern Arizona, a precise boundary between strata of Des Moines age and those of the interval directly above is difficult to establish in many places, but the boundary is located as closely as possible by datable fossils. Despite probable small errors in calculating exact thicknesses of interval C which are caused by uncertainties of the boundary in this area, the total thickness of interval C clearly is greater than that of any other Pennsylvanian interval, and isopach trends on plate 7A are believed to be approximately correct.

THICKNESS TRENDS

The interval most widely deposited in Arizona during Pennsylvanian time was interval C. Relatively thick deposits, indicating areas of maximum depression and sediment accumulation during Des Moines time, extend into the State from the northwest, northeast, and southeast. Deposits are absent from positive areas along the north and east boundaries of the State, between these negative areas. No record is preserved of Des Moines rocks in the southwest and in most of the western part of the State.

Isopachs in the northwest corner of Arizona suggest that a structural depression in that area connected with the Cordilleran geosyncline of Nevada and allowed the sea to transgress southeastward toward and across central Arizona. Rocks of interval C in northwestern Arizona are largely nonfossiliferous and probably are nonmarine. Deposition apparently was continuous across central Arizona; thin deposits connected with thicker, largely marine strata that had formed in an embayment to the southeast. Likewise, isopach trends suggest that deposits possibly extended continuously across the extreme northern part of Arizona, connecting those in the northwest with those in the northeast corner.

Interval C deposits of northeastern Arizona are progressively thinner from the central part of the Paradox basin southward and westward, but they extend farther into that part of Arizona than do the deposits of any other part of the Pennsylvanian. Thin, unfossiliferous sections in the little-known, middle parts of the Navajo Reservation may be entirely of Des Moines age, but uncertainty persists as to whether they are continuous across this area, connecting with strata of Des Moines age in either the northwest area, as already mentioned, or the east-central Arizona area (pl. 7A).

A broad extension of the Sonoran geosyncline developed in southeastern Arizona and extended northward through most of east-central Arizona, as far as the much-restricted Defiance positive element, during Des Moines time. The thickness of rock of Des Moines age is nearly 900 feet in the Gunnison Hills in southeastern Arizona (fig. 66) (H. R. Wanless, written commun., 1949). How far westward the rocks originally extended is not known, but in the Vekol Mountains, more than halfway across the State, the thickness is almost 600 feet. Northwestward, the rocks apparently connected with rocks of Des Moines age of the Grand Canyon area, and northward, they possibly connected across the Black Mesa area (fig. 66) with the relatively thick section near Four Corners. In the Black Mesa area, however, a westward extension of the Defiance positive element may have formed a barrier between deposits, as indicated by trends in thinning and by increase in detrital material near that area (pl. 7B).

LITHOFACIES TRENDS

The lithofacies map of interval C (pl. 7B) shows rock types in Arizona distributed roughly in accordance with position relative to positive and negative features. Limestone predominates in the northwest, northeast, and southeast corners of the State where the interval is thickest, adjacent to the Cordilleran miogeosyncline, Paradox basin, and Sonoran geosyncline, respectively. Toward the margins of these seaways, the proportion of mudstone to limestone increases, and in areas bordering the Defiance positive element where the interval is thin, mudstone dominates. Appreciable amounts of sandstone are present only along the southern margin of the Piute positive element (fig. 65) near the Utah border.

SOURCES AND ENVIRONMENT OF DEPOSITION

The Piute positive element, which was centered in the middle part of southern Utah and extended as a land area a short distance into northern Arizona, seems to have been not only a partial barrier of moderate relief between waters of the Cordilleran geosyncline and the Paradox basin but also a source of sand for sediments deposited along its southern and eastern margins. These are the only places in Arizona where sand accumulated in considerable amounts during deposition of interval C.

The region of the Defiance positive element seems to have been a widely emergent land area that formed a major barrier to the three principal seaways in Arizona at the time of interval C, though the precise positions of the shorelines cannot be located with available information. The land apparently was not greatly elevated at this time, however, for mudstone represents the coarsest detrital sediment that accumulated in appreciable amounts offshore.

Carbonate rock that is very extensive in interval C in many parts of Arizona suggests an environment of relatively shallow waters. Bordering landmasses apparently were of low relief. Abundant medium-scale crossbeds in both limestones and sandstones indicate strong currents and high-energy components; these alternated with quiet, low-energy conditions as
represented by structureless limestones and shaly mudstones.

Throughout the region, interval C has characteristics of cyclic deposition as described for the Horquilla Limestone by Sabins (1957) and as are apparent in strata of other areas. Furthermore, superimposed on these cycles is the record of a general deepening of the sea during early and middle Des Moines time and a gradual shallowing near the end of Des Moines time (Ross and Sabins, 1965, p. 176). Evidence of the deep-water conditions is furnished “by the relatively few beds having fusulinids and by the dominance of dark gray, fine-grained limestone in the middle part of the formation.”

PALEOTECTONIC IMPLICATIONS

Although interval C represents a time of maximum transgressions of the seas and maximum accumulation of sediment in the Pennsylvanian of Arizona, its pattern of deposition is similar to that of all other intervals. Centers of sinking and sedimentation, as well as major positive elements, are in virtually the same positions as before. Only the Piute positive element seems to have been elevated above sea level more than at other times, as it furnished considerable sand to nearby areas of deposition.

INTERVAL D
FORMATIONS INCLUDED

In northeastern Arizona, the lower part of the Upper Hermosa is arbitrarily referred to interval D (pl. 13), though evidence has not yet been obtained to firmly establish its age as Missouri. In southeastern and east-central Arizona, parts of the Horquilla and Naco Limestones, respectively, are classified as interval D, largely on the basis of Missouri faunas.

STRATIGRAPHIC RELATIONS

Rocks containing fossils of definite Missouri age and therefore assignable to interval D have not been recognized in northwestern Arizona. Probably the interval is represented by a hiatus comparable to the one in nearby parts of Nevada, where dating based on fusulinids shows that rocks of Virgil age (interval E) rest directly on rocks of Des Moines age (interval C) (Welsh, 1959, p. 56a).

On plate 8A, interval D is shown extending into the northeastern corner of Arizona from adjoining parts of Utah, Colorado, and New Mexico. This correlation is by no means definite, because fossils of Missouri age have not yet been reported from wells in either Arizona or New Mexico near the Four Corners; nor have Missouri fossils been unequivocally identified from the extreme southeastern corner of Utah (John Chronic, written commun., 1964). The basis for projecting interval D into Arizona is the recognition of fusulinids of Missouri age in samples from several wells in the central and southern part of San Juan County, Utah (table 33).

The presence of interval D south of the Mogollon Rim (fig. 66) in east-central Arizona is established by fossils of Missouri age in various places, between White River Canyon, Ariz. (loc. 414) on the east, to Fossil Creek, Ariz. (loc. 396) on the west, a distance of 100 miles (H. R. Wanless, written commun., 1960). The extent of interval D northward and northeastward in the subsurface is not known, for as yet fossils of Missouri age have not been found in that area.

Fossils considered definitely or probably of Missouri age have been recorded from many parts of southeastern Arizona (for example, Gilluly and others, 1954, p. 34; Epis, 1956; Tyrrell, 1957), but faunal studies in the Chiricahua and Dos Cabezas Mountains in the extreme eastern part of the area (Ross and Sabins, 1965, p. 175) indicate that Missouri rocks are absent there.

UPPER BOUNDARY OF INTERVAL D

In parts of southeastern Utah where the upper part of the Hermosa contains both Missouri and Virgil fossils, the lower one-third of the formation has been assigned to interval D and the upper two-thirds, to interval E. This ratio (1:2) is used southward in Arizona for subdividing the Upper Hermosa, as neither Missouri nor Virgil fossils are reported from sections in that area. Rocks of Missouri age, however, may pinch out southward before reaching the Arizona line, in which case all the Upper Hermosa is Virgil in age. Possibly, therefore, isopachs should not be shown in northeastern Arizona for interval D.

North of the Mogollon Rim in east-central Arizona, the thickness of interval D has been determined for the isopach map (pl. 8A) by projection from an area farther south where boundaries have been established on the basis of fossils. The thickness of Missouri rocks in the fossiliferous area represents about 50 percent of the strata of Pennsylvanian age younger than interval C and this proportion is used in the northern area; possibly, however, rocks of Missouri age are absent or nearly absent there.

THICKNESS TRENDS

Because Missouri rocks are not known from the Grand Canyon area in northwestern Arizona, it seems probable that interval D was restricted to the east half of the State. Furthermore, depositional areas in the northeastern and southeastern parts of the State probably did not connect, as indicated by the apparent rate of thinning of deposits within each of these areas toward the Black Mesa basin area (fig. 66) that separated them.

In northeastern Arizona, strata of interval D are at the
the eastern margin of the Paradox basin and thicken northeastward into the basin. They are 200 feet thick near Four Corners and approach a zero isopach not far to the south and west (pl. 8A).

In southeastern Arizona, the area indicated by isopachs of interval D is similar to but somewhat more restricted than that of the underlying interval C (pl. 8A, 7A). The maximum recorded thickness is 536 feet in the Gunnison Hills (H. R. Wanless, written commun., 1949; Nations, 1963, p. 1255). The axis of deposition trends northwest and north; the zero isopach diagonally crosses the approximate center of the State. The position of the western limit of deposition south of the Mogollon Rim is entirely speculative; but, because rocks of Missourian age have not been recognized in northwestern Arizona, a connection with the Cordilleran geosyncline seems extremely improbable.

The zero line and other isopachs along the eastern margin of the depositional area in southeastern Arizona seem to be more closely spaced than elsewhere, possibly indicating a nearby area with positive tendencies. This constriction of isopachs occurs near the Chiricahua Mountains where there is an absence of Missourian fossils (Ross and Sabins, 1965, p. 175); presumably the paleogeographic features in that area were distinctly different from those of the other Pennsylvanian intervals, all of which are represented in the Chiricahua section.

LITHOFACIES TRENDS

As was true for other intervals of the Pennsylvanian in Arizona, interval D was deposited in parts of geosynclines or basins that lie mostly in adjoining States. The interval is dominantly mudstone shoreward near the zero edge and dominantly limestone seaward where the isopach is thicker (pl. 8B). Between these extremes are combinations of mudstone and limestone in various proportions.

Interval D deposits of the northeastern seaway in Arizona are dominantly mudstone near the zero isopach but contain some limestone; near the Four Corners, limestone is dominant. Farther north, in Colorado and Utah, the proportion of limestone is still greater.

In southeastern Arizona, the lateral succession of lithofacies is similar to, but more extensive than, that in the northeast. It includes an area of nearly pure limestone near the borders of Mexico and New Mexico and, near the northwestern part of this depositional belt, an area of virtually pure mudstone.

SOURCES AND ENVIRONMENTS OF DEPOSITION

The environments of deposition represented by interval D seem to be similar to and a continuation of those represented in interval C, except in the northwest where the sea withdrew. Seaways were somewhat restricted, also, in eastern Arizona; a withdrawal began near the end of Des Moines time, after the progressive deepening of the seaways during early and middle Des Moines time, as described by Ross and Sabins (1965, p. 176).

A very high percentage of mudstone along the northwest margin of those deposits in the eastern and southeastern parts of the State and a progressive decrease in proportion of mudstone to limestone toward southeastern Arizona seem to indicate that a large amount of mud was coming from the northwest.

PALEOTECTONIC IMPLICATIONS

During the time of interval D, an embayment extending northwestward from the Sonoran geosyncline still occupied most of southeastern Arizona. Likewise, at this time, basin sinking and entrapment of sediment continued in northeastern Arizona along the margin of the Paradox basin. In northwestern Arizona, however, sediments no longer accumulated and a broad area stood above the base level of deposition. This part of a former sinking shelf apparently became relatively stable or slightly positive; it formed a southward extension of the Piute positive element, which continued to exist as a land area farther north.

INTERVAL E

FORMATIONS INCLUDED

Rocks assigned to interval E in Arizona consist of the upper middle part of the Supai Formation in northwestern Arizona and the Upper Hermosa in northeastern Arizona (pl. 13). Farther south, in east-central Arizona, interval E is composed of the lower member of the Supai Formation to the west and the intertonguing uppermost part of the Naco farther east. In southeastern Arizona, it consists of the uppermost part of the Horquilla Limestone and the lower part of the Earp Formation, as indicated by the presence of Virgil fossils (H. R. Wanless, written commun., 1949; Epis, 1956; Sabins and Ross, 1963, fig. 4).

STRATIGRAPHIC RELATIONS

Recent faunal evidence establishes the presence of interval E, together with interval C, within a largely unfossiliferous middle part of the Supai Formation in northwestern Arizona. The extent and limits of interval E within this rock sequence, however, have not been determined accurately. Because intervals C and E in nearby parts of southern Nevada have been shown on the basis of fusulinid data to occur in the ratio of 3 to 2, these proportions are arbitrarily applied in Arizona in calculating the interval thicknesses within the Supai strata of questionable age.

In northeastern Arizona, the upper two-thirds of the Upper Hermosa has been assigned to interval E because
this proportion of the formation is of Virgil age in parts of San Juan County, Utah, to the north, where datable fusulinids have been obtained from it. Although the lower one-third of the formation is assigned to interval D on the basis of a few fossil determinations in the same area (table 33), evidence for the projection is admittedly weak and a Virgil age for all of the Upper Hermosa in northeastern Arizona is certainly possible. The thicknesses shown on plate 9A, therefore, possibly should be considerably increased.

Rocks of Virgil age, assigned to interval E, are widely distributed throughout southeastern Arizona, in what was probably an extension of the Sonoran geosyncline. They thin progressively northward toward a zero isopach north of east-central Arizona and they probably once connected across the middle of the State with thin deposits of corresponding age in the northwest (pl. 9A). The westward extent in Arizona has not been determined.

Evidence that interval E extends from the southern part of Arizona northward across east-central Arizona in a form (pl. 9A) similar to that of the two underlying intervals is found in the wide distribution of fusulinids of Virgil age in the upper part of the Naco Formation in northeastern Arizona. The zone of Cucullodiscus culomensis, which in outcrop at Salt River Canyon (Arizona loc. 412) is 800 feet above the base of the Pennsylvanian, has been recognized by Løkke (1962, p. 85) in the subsurface both 53 miles to the northeast (loc. 281) and 53 miles to the north-northeast (loc. 371). Additional fossils of Virgil age have been observed in the Christmas quadrangle (Willden, 1964, p. E25) and at numerous other localities in this part of Arizona.

**UPPER BOUNDARY OF INTERVAL E**

The upper limit of interval E in northwestern Arizona is clearly defined by a surface of erosion marked by channels as much as 30 feet deep and overlain in most places by a conspicuous conglomerate. Strata above the erosion surface are of Wolfcamp age as indicated by fusulinids, corals, and brachiopods from several localities in western Grand Canyon.

Throughout the eastern half of Arizona, interval E, composed of limestones and other rocks of Virgil age, is overlain with apparent conformity by Permian strata considered to be of Wolfcamp age. Lithologic changes between formations are generally used to mark the boundary. In well logs of northeastern Arizona, the contact is at the base of the Cutler evaporites that overlie limestone in the Upper Hermosa. In east-central Arizona, the boundary is the base of the "middle Supai member" as used by Huddle and Dobrovolny (1945) or the bottom of the "Supai salt," "Supai Big A," and other units that overlie limestone of the Naco, as used by the American Stratigraphic Company.

The upper boundary of interval E in southeastern Arizona has been determined on the basis of fusulinids. It is placed in most localities within limestones of the Earp Formation (Tyrrell, 1957; Sabin, 1957, p. 490; Ross and Tyrrell, 1965, p. 620) and is located rather closely by diagnostic zones of Triticites. A red chert conglomerate on an erosion surface at this stratigraphic position in the Whetstone Mountains may indicate a hiatus in that area (Ross and Tyrrell, 1965, p. 620).

In the Gunnison Hills of southern Arizona, a conspicuous conglomerate in the Earp separates a lower unit of light-colored limestone that contains abundant fusulinids from an upper unit of dark limestone that contains a gastropod-cephalopod fauna. On the basis of this difference in lithology, Cooper (in Gilluly and others, 1954, p. 21) subdivides the formation into two parts and suggests that there may be represented "a time-break of possible significance." Despite uncertainty regarding age assignment of the fossils, the lower part of the Earp, discussed above, is mapped in this publication as a part of the Pennsylvanian interval E and the upper part as a part of the Permian interval A.

**THICKNESS TRENDS**

Rocks probably of Virgil age in northwestern Arizona, assigned to interval E, have a maximum thickness of 200 to 300 feet. They thin progressively eastward along the State boundary toward the Piute positive element, which extends south from the middle part of Utah. Isopachs of interval E swing south and east around the end of this element and either closely approach or connect with isopachs of northeastern Arizona. More significant is a trend in isopachs of northwestern Arizona and another in southeastern Arizona which suggests that a negative belt probably once extended diagonally across central Arizona to connect the two areas. The trend of this belt was roughly parallel to and south of the Defiance positive element and its westward extension.

Interval E thins southward and westward from northeastern Arizona where it is widely distributed, as indicated on plate 9A. This trend is well established at places by abundant well-sample coverage, but the position of the zero isopach is very uncertain. Trends where control is good suggest that the interval was limited by the Defiance positive element to the south and that it probably did not extend across the Black Mesa area to the southwest. Thus, although Virgil strata of the Four Corners area may have connected with those of similar age in the Grand Canyon area along the Arizona-Utah boundary, it is equally likely that strata in the two places were confined to separate depositional areas.

Interval E in southeastern Arizona is thickest along a northwest-trending axis, and follows the pattern of interval D (pl. 8A) that preceded it. The maximum recorded thicknesses of this unit, which are in the extreme eastern
part of the State in the Chiricahua and Pedregosa Mountains (fig. 66), are between 500 and 600 feet.

In many parts of east-central Arizona, the thickness of interval E, like that of intervals C and D, is difficult to determine. In a few outcrops along the Mogollon Rim, however, faunal zones are sufficient to permit reliable local subdivision of the Naco Formation into units of Atoka, Des Moines, Missouri, and Virgil age. These units are extended northward into the subsurface by extrapolation, maintaining their approximate proportions. Such procedure involves numerous uncertainties, but seems necessary where lithologic subdivisions have not been recognized and where diagnostic fossils are scarce or lacking. It results in an isopach pattern that thins significantly northward for each interval — a situation that has some support from information on thinning of the entire system and on the extent of certain fossil zones.

LITHOFACIES TRENDS

The lithofacies pattern of interval E (pl. 9B) in Arizona resembles the patterns of underlying Pennsylvanian intervals in showing high percentages of limestone in the northwest and southeast corners of the State, and substantial amounts in the northeastern part — areas closest to the Cordilleran geosyncline, Sonoran geosyncline, and Paradox basin, respectively. The pattern differs markedly from the patterns of other intervals, however, because of the large proportion of sandstone in the northwestern quadrant of the State. The amount of sandstone progressively decreases and the amount of mudstone increases eastward toward the Defiance positive element and northward toward the Plute positive element in middle-southern Utah. Along the western margins of the Defiance positive element, mudstone only is present.

SOURCES AND ENVIRONMENT OF DEPOSITION

In most of eastern Arizona, depositional environments that were established early in the Pennsylvanian Period apparently were maintained with only minor interruptions through Virgil time. The major seaways expanded or contracted, and continental or shallow-water marine deposits accumulated on their floors or along their margins. Either terrigenous mud or calcium carbonate dominated locally according to the distance from shore.

The high percentage of sandstone in interval E in northwestern Arizona suggests that for the first time in the Pennsylvanian a source for such material developed somewhere west or southwest of the State line. Whatever the source, an eastward movement of sand and mud into the State began in the Late Pennsylvanian and continued into Early Permian time.

PALEOTECTONIC IMPLICATIONS

Latest Pennsylvanian time in Arizona was marked by continued sinking of depositional area in the northwest, northeast, and southeast corners of the State. The Defiance positive area and a similar low uplift (Plute positive element) that extended southward into Arizona from the middle of southern Utah were at least slightly positive, and they thus restricted the seaways. An uplift of an area outside the State to the west or southwest of the Grand Canyon area seems to be indicated by an influx of sand from that direction.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

THICKNESS TRENDS

In Arizona, three areas of thick sediment accumulation apparently were centers for transgressing and regressing seas during Pennsylvanian time. These areas are the northwest, northeast, and southeast corners of the State (pl. 11). Positive elements, delineated by zero isopachs, persisted in the middle part of the northern border area and along parts of the eastern border area of the State. A thin deposit of sediment accumulated somewhat north of the center of the State; west and southwest of this area the Pennsylvanian record is largely missing because of erosion and metamorphism.

Throughout most of Arizona, present-day thicknesses of Pennsylvanian rocks probably approximate original thicknesses of deposition, despite the several widespread hiatuses that have been recognized in various areas. The lack of rocks of Atoka and Missouri age in northwestern Arizona and the absence of rocks of Morrow age from the northeastern and middle-eastern parts of the State indicate appreciable time gaps in the record, but the apparent lack of a disconformity suggests that appreciable erosion did not take place during these intervals.

PALEOTECTONIC IMPLICATIONS

Isopach trends suggest that the northwestern area that has relatively thick deposits was structurally a shelf on the east margin of the Cordilleran geosynclinal belt of Nevada; the northeastern area was part of the fast-sinking Paradox basin of Colorado; and the southeastern area was structurally part of the Sonoran geosyncline of Mexico. Isopach trends further suggest that, at least briefly, a shallow, linear depressed area developed across northern Arizona south of Grand Canyon and connected the deeper, geosynclinal belts of the northwest and southeast; another shallow, indefinitely located depression formed near Black Mesa on the Navajo Indian reservation and temporarily joined the Paradox basin with both geosynclines. How far west in Arizona the Sonoran geosyncline originally extended is not known, but thickness trends in areas where isopachs are well controlled suggest a considerable probability that southwestern Arizona or adjacent parts of California were tectonically positive.
The lack of any apparent alinement of the positive or negative elements in Arizona during Pennsylvanian time seems to be responsible for a corresponding irregularity in facies trends of the Pennsylvanian rocks. Marine deposits occur exclusively in the thicker sections. Terrigenous red beds, in general, make up the thin sections lapping landward onto the positive elements and interfingering with marine deposits seaward.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

**UNITS OVERLYING PENNSYLVANIAN**

Throughout most of northern Arizona, the boundary between the Pennsylvanian and Permian Systems lies within the Supai Formation, as currently defined. In the extreme northwestern corner of the State, a carbonate rock unit or tongue, referred to as the Pakoon Limestone by McNair (1951, p. 524-525), is of Wolfcamp age and overlies a rock sequence, largely red beds, of probable Virgil age that is assigned to interval E of the Pennsylvanian. At the base of the Pakoon is a widespread conglomerate. Farther east, along the Grand Canyon and its tributaries and in the subsurface of Black Mesa basin, the lowest Permian is represented by red sandstones of the Supai that have been referred to, for many years, as the Esplanade Sandstone Member. In this area underlying red beds, presumably of Pennsylvanian age, are everywhere capped by a conglomerate, continuous with that in the Pakoon and forming the base of the Esplanade (Permian).

In northeastern Arizona where deposition was marginal to the Paradox basin of Utah and Colorado, terminology of the Four Corners area, rather than that of the Grand Canyon, commonly is used. Thus, strata overlying the highest rocks of definite Pennsylvanian age in this area are referred to as the Cutler Formation, and, in the subsurface near Four Corners, they are represented mostly by Cutler evaporites believed to be of Wolfcamp age. The Pennsylvanian-Permian boundary in this area seems to be gradational and a time of general regression is represented (Peterson and Ohlen, 1963, p. 70). As a result of this regression, which was from south to north, the boundary currently recognized between Virgil and Wolfcamp rocks is at the top of the marine Upper Hermosa near the basin, but entirely within red beds of the Supai or Cutler Formation farther south (Peterson, 1959; Peterson and Ohlen, correlation chart, 1963).

In the Fort Apache area of middle-eastern Arizona, the Pennsylvanian-Permian systemic boundary has, in this publication, arbitrarily been placed at the contact of the Naco and Supai Formations (Winters, 1963, p. 15; Brew, 1965, p. 83). This placement is based on the Pennsylvanian aspect of Naco fossils, although, admittedly, diagnostic fossils are absent in the Fort Apache area, both in the lower part of the Supai and in transitional beds that have been assigned to the Naco (Winters, 1963, p. 8; Brew, 1965, p. 81). Further complications are introduced regionally by differences in age of the lithologic boundary at different places — the result of regressive conditions in which continental detrital deposits from the north and northwest advanced progressively south and southeast across the dominantly marine beds of the Naco. Thus, at Fossil Creek in central Arizona northwest of Fort Apache, the upper boundary of the Naco may be as old as late Des Moines (Jackson, 1951a, b).

The same seaway that is represented by the Naco of east-central Arizona apparently covered most of southern and southeastern Arizona during the closing stages of Pennsylvanian time; and, in that area, which was close to the Sonoran geosyncline, the sea remained well into Permian time. As a result, the Pennsylvanian-Permian boundary occurs within the marine Earp Formation (Gilluly and others, 1954, p. 38). Farther west, it is within the lower part of the Andrada Formation, considered, in part, a facies equivalent of the Earp (Bryant, 1955).

The much greater proportion of clastic deposits in the Earp than in the conformably underlying or overlying formations (Gilluly and others, 1954, p. 19) suggests that a maximum stage of regression was attained at the time of its deposition. Thus, a boundary between the Pennsylvanian and Permian Systems in southeastern Arizona roughly coincides with the southernmost line of advance of the Supai delta into the northern end of the Sonoran geosyncline.

In the area of Superior and Globe in south-central Arizona and in various areas farther south and west where rocks of Permian age are absent, several different formations overlie the Pennsylvanian. South of Globe (fig. 66) in the Saddle Mountain quadrangle, Mesozoic (?) sedimentary rocks rest on Naco (Krieger, 1968b); near Superior (fig. 66) the Oligocene Whitetail Conglomerate is above the Naco (Peterson, 1969); farther east, near Christmas, Upper Cretaceous volcanics, in places, and the Upper Cretaceous Pinkard (?) Formation, in other places, overlie the Naco (Willden, 1964, p. E24). Thus, in at least this small part of Arizona, the record of latest Paleozoic deposition and erosion was eliminated by subsequent stripping.

In southern Arizona, the location of the systemic boundary within the Earp Formation and its correlates has been determined only approximately, on the basis of fauna. The exact position is at present speculative, and an arbitrary boundary has been used in this publication. In the Whetstone Mountains, the systemic boundary is believed by Tyrrell (1957) to lie immediately below a jasper conglomerate within the Earp; the overlying part is considered to be of Wolfcamp age and the greater part of the underlying part is believed to be of Virgil age.
PALEOTECTONIC IMPLICATIONS

Apparently, nowhere in Arizona did appreciable tectonism occur at or near the close of Pennsylvanian time. In most of Arizona, uppermost rocks of Pennsylvanian age are overlain conformably by strata of Early Permian age, and continuous deposition from Pennsylvanian into Permian time seems likely. In the Grand Canyon area and in parts of southern Arizona, a disconformable surface at the systemic boundary indicates uplift and erosion; no folding has been detected. The hiatus represented is of unknown, but probably not great, magnitude. In the Four Corners area, east-central Arizona, and much of southern Arizona, evidence of a break in the record has not been detected.

Seas were generally regressing in the region at the close of the Pennsylvanian. Across wide areas, continental detrital sediments were becoming dominant over chemical marine deposits, which suggests that structural conditions were relatively stable. Sediments were accumulating more rapidly than basins and shelves were sinking. Nothing more than gentle warping seems to be represented by the tectonic record of this time.

REFERENCES


Nevada and Southern California

By RICHARD F. WILSON

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-Q
## CONTENTS

<table>
<thead>
<tr>
<th>Abstract</th>
<th>311</th>
</tr>
</thead>
<tbody>
<tr>
<td>Region defined</td>
<td>311</td>
</tr>
<tr>
<td>Paleogeology</td>
<td>313</td>
</tr>
<tr>
<td>Units underlying Pennsylvanian</td>
<td>313</td>
</tr>
<tr>
<td>Lower boundary of the Pennsylvanian</td>
<td>311</td>
</tr>
<tr>
<td>Interval A</td>
<td>314</td>
</tr>
<tr>
<td>Formations included</td>
<td>314</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>315</td>
</tr>
<tr>
<td>Upper boundary of interval A</td>
<td>315</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>316</td>
</tr>
<tr>
<td>Lithofacies trends and sources of sediment</td>
<td>316</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>316</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>317</td>
</tr>
<tr>
<td>Interval B</td>
<td>317</td>
</tr>
<tr>
<td>Formations included</td>
<td>317</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>317</td>
</tr>
<tr>
<td>Upper boundary of interval B</td>
<td>318</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>318</td>
</tr>
<tr>
<td>Lithofacies trends and sources of sediment</td>
<td>318</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>318</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>319</td>
</tr>
<tr>
<td>Interval C</td>
<td>319</td>
</tr>
<tr>
<td>Formations included</td>
<td>319</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>319</td>
</tr>
<tr>
<td>Upper boundary of interval C</td>
<td>319</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>320</td>
</tr>
<tr>
<td>Interval C — Continued</td>
<td></td>
</tr>
<tr>
<td>Lithofacies trends and sources of sediment</td>
<td>320</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>320</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>320</td>
</tr>
<tr>
<td>Interval D</td>
<td>321</td>
</tr>
<tr>
<td>Formations included</td>
<td>321</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>321</td>
</tr>
<tr>
<td>Upper boundary of interval D</td>
<td>321</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>322</td>
</tr>
<tr>
<td>Lithofacies trends and sources of sediment</td>
<td>322</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>322</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>322</td>
</tr>
<tr>
<td>Interval E</td>
<td>322</td>
</tr>
<tr>
<td>Formations included</td>
<td>322</td>
</tr>
<tr>
<td>Stratigraphic relations</td>
<td>323</td>
</tr>
<tr>
<td>Upper boundary of interval E</td>
<td>323</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>324</td>
</tr>
<tr>
<td>Lithofacies trends and sources of sediment</td>
<td>324</td>
</tr>
<tr>
<td>Environment of deposition</td>
<td>324</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>325</td>
</tr>
<tr>
<td>Total thickness of Pennsylvanian rocks</td>
<td>325</td>
</tr>
<tr>
<td>Thickness trends</td>
<td>325</td>
</tr>
<tr>
<td>Paleotectonic implications</td>
<td>326</td>
</tr>
<tr>
<td>Geologic units directly above Pennsylvanian System</td>
<td>326</td>
</tr>
<tr>
<td>References</td>
<td>326</td>
</tr>
</tbody>
</table>

## ILLUSTRATIONS

For listing of plates (in separate case) see volume "Contents"

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>67</td>
<td>Map showing geographic features in the Nevada and southern California region referred to in text</td>
<td>312</td>
</tr>
<tr>
<td>68</td>
<td>Palinspastic map of earliest Pennsylvanian paleotectonic elements in the Nevada and southern California region</td>
<td>313</td>
</tr>
<tr>
<td>69</td>
<td>Map showing inferred restoration of control points from present positions to positions before post-Pennsylvanian faulting</td>
<td>313</td>
</tr>
</tbody>
</table>
PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

NEVADA AND SOUTHERN CALIFORNIA

By Richard F. Wilson

ABSTRACT

The region of Nevada and southern California during the Pennsylvanian Period included parts of four major tectonic provinces: (1) a shelf area along the southeastern and southern borders of Nevada; (2) a miogeosynclinal belt extending northeastward across eastern Nevada; (3) a linear positive area, termed the "Antler positive element" (the Antler orogenic belt of earlier Paleozoic time), west of and paralleling the miogeosynclinal belt in Nevada; and (4) a eugeosynclinal area bordering the Antler positive element on the west and extending an unknown distance westward across Nevada and southern California.

In the shelf area rocks of earliest Pennsylvanian age overlie, for the most part, strata of Meramec age; in the miogeosynclinal area, they cover strata of Chester age; and on the flanks and locally on the top of the Antler positive element they overlie lower Paleozoic beds, mostly of Cambrian and Ordovician age. In the eugeosynclinal area the record is less clear, but in places Lower Pennsylvanian rocks may rest upon beds of Chester age; continuous deposition from Mississippian into Pennsylvanian time is inferred.

In eastern Nevada a single, large, elongate basin developed in the miogeosynclinal belt during the time of interval A (roughly equivalent to the Morrow time), but by interval B or Atoka time and until the end of the Pennsylvanian Period, this negative area was occupied by three or more smaller basins. In the elliptical Bird Spring–Ely basin in southeastern Nevada a total of 3,500 feet of sediment accumulated during the period. In the centers of two smaller basins in northeastern Nevada similar thicknesses are recorded. Strata in all the basins consisted of dominantly carbonate rock, but along the margins some non-carbonate detritus accumulated.

Relatively thin shelf deposits of extreme southern Nevada constitute a record of successive transgressions and regressions. The sea advanced a long distance into Utah and northern Arizona during Morrow, Des Moines, and Virgil time (intervals A, C, and E), and it receded greatly during Atoka and Missouri time (intervals B and D).

Parts of the Antler positive element of central Nevada remained emergent during most of the Pennsylvanian Period. However, the southernmost section, which had formed a barrier between the eastern and western seaways in southern California during interval A time, was submerged from Atoka time to the end of the period. Detrital accumulations along the margins of the Antler highland in Nevada record intermittent uplift; deposits overlapped the north end of the positive element and may have formed locally across the central part during intervals B and E. Uplifts may have extended from the north end and formed northeast-trending prongs during Late Pennsylvanian time, as suggested by the configuration of zero isopachs for intervals D and E, but whether these zero isopachs represent the ancient shoreline is questionable, for neither interval contains any terrigenous detritus nearby.

The eugeosyncline that occupied the area west of the Antler positive element during Pennsylvanian time is documented by only a few, scattered outcrops in Nevada and southern California. A considerable thickness of rock within the eugeosyncline, however, is well shown by the more than 6,000 feet of strata in the Pumpernickel Formation and lower part of the Havaillah Formation at Battle Mountain, Nev., and by the approximately 10,000 feet in the lower part of the Garlock Formation in California. The eugeosynclinal sequences include much bedded chert and volcanic material; primary structures characteristic of turbidity currents are locally abundant in upper parts of these sequences. Deposition was nearly continuous throughout the period, but after accumulation of interval C the character of the sediments in the northern part of the eugeosyncline changed with introduction of many pebbly sands and conglomerates, probably as turbidites.

In virtually all parts of the Nevada and southern California region, youngest Pennsylvanian rocks are overlain by strata of Wolfcamp age.

REGION DEFINED

The Nevada and southern California region includes all of the State of Nevada and all of California south of the 40th parallel. Within this region, Pennsylvanian rocks crop out in many mountain ranges in southern, eastern, and north-central Nevada but are not known to be present, or are absent, over much of the western part of the State. Scattered outcrops of Pennsylvania rocks are present in southeastern California, particularly in the Death Valley and northern Mojave Desert areas (fig. 67). Rocks of known Pennsylvanian age are present in outcrop in the central Sierra Nevada area, but no others are reported elsewhere in the Sierras.

During Pennsylvanian time, the region was the site of part of the Cordilleran eugeosyncline in western Nevada and parts of California; the Antler positive element (the Antler orogenic belt of earlier Paleozoic time) in central Nevada; several troughs and basins within the Cordilleran miogeosyncline in eastern Nevada and parts of southern California; and a shelf area at the border of Nevada and Utah (fig. 69). Sparse stratigraphic data, post-Pennsylvanian strike-slip movement of considerable magnitude along several fault systems in southern and western Nevada and California, and eastward displacement of stratigraphic units for dis-
Figure 67. — Geographic features in the Nevada and southern California region referred to in text. Symbols and numbers are localities shown on plate 1.
PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

In Nevada and southern California, Pennsylvanian strata overlie rocks ranging in age from youngest Mississippian (Chester) to Cambrian. The distribution of these pre-Pennsylvanian strata as shown on plate 2 in large part reflects the position of late Paleozoic paleotectonic elements in this area. In southern Nevada and in the eastern Mojave Desert in California, the area occupied by the Mississippian Monte Cristo Formation, which is dated as Meramec age in its upper part, coincides with a shelf bordering the southeast side of the Cordilleran miogeosyncline in Pennsylvania. This shelf rocks mark the limits of several basins and troughs that developed in the Cordilleran miogeosyncline in late Paleozoic time.

In northeastern and north-central Nevada, allochthonous Cambrian, Ordovician, Silurian, and, in one place, Devonian rocks belong to the siliceous and volcanic assemblage originally deposited in the Cordilleran eugeosyncline in western Nevada. This assemblage was displaced eastward during the Antler orogeny in Late Devonian and Early Mississippian time. Patterns generally are not shown on plate 2 for pre-
Pennsylvaniaian rocks in the area of the Cordilleran eugeosyncline west of the Antler positive element in Nevada or for areas of outcrop in southern California (El Paso Mountains, loc. 143; Calico Mountains, loc. 206) where Pennsylvaniaian rocks of the detrital-volcanic assemblage are inferred, with varying degrees of certainty, to be present. However, known or probable Upper Mississippian rocks are present at various localities in northwestern Nevada and presumably were coextensive with the Pennsylvaniaian rocks of eugeosynclinal character. The Goughs Canyon Formation of the Osgood Mountains (Hotz and Willden, 1964, p. 26-28) and the Inskip Formation in the East Range (Roberts and others, 1958, p. 2847) are among these Upper Mississippian units.

In northern Elko County (loc. 305; pl. 2), Upper Mississippian strata have been identified in the lower part of the 9,000-foot sequence of detrital and volcanic rocks assigned to the Schoonover Formation (Fagan, 1962; Churkin and Kay, 1967). Formerly, Pennsylvaniaian and possibly Permian ages were inferred for the upper part of the Schoonover; however, a probable Chester age is now suggested for the entire formation on the basis of conodont studies by J. W. Huddle (Robert Coats, oral commun., 1972).

LOWER BOUNDARY OF THE PENNSYLVANIAN

Where the Pennsylvaniaian rests on rocks of Chester age in eastern or southern Nevada or southeastern California, the contacts between the two systems are conformable in some places (Gordon and Poole, 1968) and apparently conformable elsewhere (Webster and Lane, 1967). Where the Pennsylvaniaian rests on rocks of Meramec age the contact is disconformable.

Along the Antler positive element the Pennsylvaniaian rocks lie with angular unconformity on middle or lower Paleozoic units. In the Antler Peak, Edna Mountain, and Osgood Mountains areas in north-central Nevada (fig. 67), the lower boundary of the Pennsylvaniaian is placed at the unconformity at the base of the Battle Formation, which contains Middle Pennsylvaniaian (Atoka) fossils. The lower part of the Battle Formation is primarily conglomerate containing small amounts of sandstone, and it overlies the Osgood Mountain Quartzite, Preble Formation, Harmony Formation, and Scott Canyon Formation (not shown on pl. 2), all of Cambrian age, and Valmy Formation of Ordovician age. The Osgood Mountain Quartzite is predominantly fine to medium grained (Hotz and Willden, 1964, p. 7); the Preble is composed of shale and limestone (Hotz and Willden, 1964, p. 10); the Harmony is composed of interbedded sandstone, arkose, shale, and limestone (Roberts, 1964, p. A23); the Scott Canyon is predominantly chert, argillite, and greenstone (Roberts, 1964, p. A15); and the Valmy, the most extensively exposed of these lower Paleozoic formations, consists of chert, shale, greenstone, and quartzite (Roberts, 1964, p. A17-A18). Thus, a distinctive lithologic as well as structural and faunal break marks the base of the Pennsylvaniaian in this region.

In the Cortez Mountains (loc. 253), east of the Antler Peak area, the base of the Pennsylvaniaian is placed at the base of the Brock Canyon Formation, which contains possible Late Pennsylvaniaian fossils in its lower part (Muffler, 1964, p. 6, 7). The basal member of the Brock Canyon is chert-pebble conglomerate, lithic arenite, and quartzite, and it rests unconformably on the Vinini Formation (Ordovician), composed predominantly of chert and shale in its upper part, near the type locality of the Brock Canyon (Muffler, 1964, p. 7, 88).

At Pyramid Peak (loc. 269) and elsewhere in the Toquima Range in northern Nye County, the Wildcat Peak Formation of Late Pennsylvaniaian age (Kay and Crawford, 1964, p. 441-443) unconformably overlies the Pinecone and Willow Canyon Formation of Ordovician age (Kay and Crawford, 1964, pls. 1, 6). The basal part of the Wildcat Peak is sandy and silty limestone which contains Pennsylvaniaian (Virgil) fusulinids and rests unconformably on folded argillites and cherts of Ordovician age (Kay and Crawford, 1964, p. 441). The unconformity is an uneven erosion surface with small relief.

INTERVAL A
FORMATIONS INCLUDED

Rocks of Pennsylvaniaian age in Nevada and southern California have been divided into intervals A through E, roughly corresponding to the five provincial series of the Pennsylvaniaian System as recognized in the midcontinent region. Many formational names have been applied. The complexity of the nomenclature in part reflects differences in lithology and fauna from one area to another; to a considerable extent it is attributable to the uncertainties of correlation between widely scattered outcrops and to difficulties in identifying some of the units in areas of structural complexity and metamorphism.

Principal rock units of the region assigned to interval A consist of part or all of three formations in southern Nevada, of two formations in central Nevada, and of three in southern California (pl. 13). Interval A is composed of the lower part of the Calville Limestone in eastern Clark and Lincoln Counties in southern Nevada and part of the Bird Spring Formation farther west. Included in interval A throughout much of the area are detrital beds in the upper part of the Indian Springs Member of the Bird Spring Formation, which, however, are of controversial age and are considered Late Mississippian by some geologists. At a single locality (loc. 162) in southern Nye County, the
name Tippipah Limestone has been used for rocks of Pennsylvanian age. The lowermost part of this formation, considered of Morrow age, is assigned to interval A.

In central Nevada interval A is represented at many localities throughout a wide area by a part of the Ely Limestone. In northern Lincoln County rocks referred to this interval form the lower unnamed member of the Ely and are distinguished from rocks of the upper member (intervals B and C) on the basis of fossils. In much of White Pine County and in Elko County, the two members of the Ely have not been recognized, but interval A consists, nevertheless, of the lowest part of the Ely and, in one locality in White Pine County, the Jensen Member (equivalent to Illipah Formation, of some geologists) of the Chainman Shale. Where the Ely was given group status by Dott (1955) in Elko County, most of the basal unit, the Moleen Formation is in interval A.

The Pumpernickel and Havallah Formations (collectively called the Havallah sequence by Silberling and Roberts, 1962, fig. 2) are difficult to date and identify. Until recently they were believed to be confined primarily to Battle Mountain and adjacent areas. However, rocks of the Havallah sequence have now been identified at widely scattered localities in Lander County by Stewart and McKee (1970), and lithologically similar rocks were recognized in western Nye County, southern Mineral County, and northern Esmeralda County by R. C. Speed (1971; oral commun., 1971).

Rocks of the Havallah sequence are dated on the basis of fossils at a few localities. Fusulinids of Atoka age, possibly reworked, occur at Battle Mountain in the lower part of the Havallah Formation; Permian (late Wolfcamp or early Leonard) fusulinids also occur in the Havallah at Battle Mountain (Silberling and Roberts, 1962, p. 17, 18). The Pumpernickel and Havallah may, therefore, range in age from at least Early Pennsylvanian to Permian, but the specific age of these rocks, except in a few localities such as Battle Mountain, is unknown. The Pumpernickel Formation has been assigned in this publication to intervals A, B, and C, and the Havallah (part), to intervals D and E.

During the Sonoma orogeny in Early Triassic time (Speed, 1971; Nichols, 1971), rocks of the Havallah sequence were thrust eastward along the Golconda thrust (pl. 3) over Pennsylvanian and Permian rocks of contrasting lithologic type, such as the Battle Formation and Antler Peak Limestone.

In the Barstow (loc. 203) and El Paso Mountains (loc. 143) areas of southern California, interval A consists of the lower parts of the Oro Grande Formation and the Garlock Formation (members 1-9), respectively. In the Calico Mountains of southern California (loc. 206) it is represented in an unnamed sequence of detrital and metavolcanic rocks, marble, and hornfels of questionable Pennsylvanian age. At Mount Morrison (loc. 171), farther north in California along the east side of the Sierra Nevadas, marble containing brachiopods typical of the lower part of the Ely Limestone is assigned to interval A.

**STRATIGRAPHIC RELATIONS**

The assignment of strata to interval A in the Callville, Bird Spring, and Ely Formations is based largely on the presence of Morrow fossils. In the Pumpernickel, Garlock, and Oro Grande Formations, rocks are arbitrarily assigned to interval A on the basis of thickness as determined in the few places where the rocks have been dated by fossils.

In some areas problems have resulted from differences of opinion concerning certain faunas. In northwestern Clark County, for instance, the Indian Springs Member of the Bird Spring Formation (Indian Springs Formation of Webster and Lane, 1967) was placed entirely in the Mississippian, on the basis of the brachiopod Rhipidomella nevadensis zone, by Webster and Lane (1967, p. 507-510) and by Dunn (1970, p. 2967). Farther east, partial equivalents of the Indian Springs were placed in the Mississippian by Langenheim, Carss, Kennerly, McCutcheon, and Waines (1962, p. 603); however, at Arrow Canyon (loc. 247) the Rhipidomella nevadensis zone was considered to be basal Pennsylvanian by these authors. The Mississippian-Pennsylvanian boundary was placed within the Indian Springs Member on the basis of other brachiopods by Rich (1961, p. 1166) and on the basis of physical criteria by Welsh (1959, p. 57). In the present publication the upper part of the Indian Springs is placed in interval A of the Pennsylvanian, and the lower boundary for rocks of Morrow age as picked by Rich and by Welsh is used as the base of interval A.

A second controversy involves the age assignment of the Jensen Member (Illipah Formation of some geologists) of the Chainman Shale in southwestern White Pine County (loc. 66). This unit was considered by Bissell (1964, p. 571) to be in the “Morrowan-Springeran Series” and by Brill (1963, pl. 1) to be in the Mississippian-Pennsylvanian “transition beds.” In contrast, it has been assigned on the basis of conodonts a Late Mississippian (Chester) age (Dunn, 1970, p. 2967). It is mapped with interval A of the Pennsylvanian in the present publication.

**UPPER BOUNDARY OF INTERVAL A**

The upper boundary of interval A is marked by disconformity in the area of the Callville Limestone of extreme southeastern Nevada; elsewhere in the Nevada-Southern California region the boundary is conformable. The lower part of the Callville contains a fauna of Morrow age (Welsh, 1959, p. 58), and the mid-
dle part has been determined to be of Des Moines age. Thus, a hiatus involving at least all of Atoka time is represented at the top of strata assigned to interval A; no evidence of a physical break has been reported at this horizon.

In most parts of southern Nevada and California where the Bird Spring Formation is recognized and in central Nevada where the Ely Limestone is present, the boundary between intervals A and B is placed at the Morrow-Atoka boundary as determined faunistically. These formations are dominantly carbonate rocks, and the first occurrence of the genus *Profusulinella* marks the interval boundary. In parts of northern Lincoln County, Nev., a marker bed within the zone of *Chaetetes-Profusulinella* has been used as a boundary indicator.

In other areas, largely in western Nevada and southern California, where detrital rocks, volcanics, chert, or metamorphics are dominant and diagnostic fossils are scarce or absent, boundaries between intervals of the Pennsylvanian cannot be located at the series boundaries, so an arbitrary thickness is assigned. Among such sequences are the Pumpernickel Formation of west-central Nevada and the Oro Grande and Garlock Formations of southern California.

**THICKNESS TRENDS**

Isopachs for interval A (pl. 3A) in Nevada and southern California define a north-trending basin that contains rocks more than 1,000 feet thick and that extends from Clark County into southwestern Elko County, Nev. The northern part of this belt of thick rocks roughly parallels the Antler positive element to the west, but the southern part diverges from it. The maximum recorded thickness within the basin is 1,650 feet, in southwestern White Pine County.

No reliable data on thickness of rocks of Morrow age are available for the eugeosynclinal area west of the Antler positive element in western Nevada or southern California.

**LITHOFACIES TRENDS AND SOURCES OF SEDIMENT**

The lithofacies patterns for interval A (pl. 3B) show that the basin in eastern Nevada was filled by dominantly carbonate deposits. Sediments of bordering shelf areas, along both the western margin in Nevada and the eastern margin in Utah, include detrital deposits in progressively larger proportions away from the basin axis. A source of detritus in the Antler positive element to the west is indicated by sand and mud in dominantly carbonate rocks in northern Eureka County and southwestern Elko County; farther south, a source in the Antler positive element is shown by distribution of detritus in carbonate sequences of the Spring Mountains and southern Las Vegas Range.

Rocks of interval A, though sparsely represented and poorly preserved in western Nevada and southern California, are very different lithologically from those to the east. In the unnamed units of the Calico Mountains (loc. 206), in rocks of the Garlock Formation of the El Paso Mountains (loc. 143) in California, and in strata of Battle Mountain and other localities in the upper plate of the Golconda thrust in Nevada, appreciable amounts of volcanic material are included. In stratigraphic sequences of the El Paso Mountains and Battle Mountain much chert is present. About 42 percent of the Pumpernickel Formation is chert in some areas, and the Pumpernickel as a whole resembles the lower part of the Garlock Formation (Roberts, 1964, p. A44). Mud and sand presumably were brought into a eugeosyncline from elevated areas in the Antler uplift to the east or, perhaps, from tectonic faults within the eugeosyncline (Fagan, 1962).

**ENVIRONMENT OF DEPOSITION**

Strata of interval A east of the Antler positive element in Nevada are believed to have been deposited in a normal marine environment. Some of them apparently were formed within a basin, others on bordering shelves. Regardless of location, they consist dominantly of carbonate sediments, and they differ principally in total thickness of accumulation and in amount of associated detritus. The basin deposits as represented by the Ely Limestone of east-central Nevada and the Bird Spring Formation of southern Nevada are much thicker and contain less sandstone than the shelf deposits as illustrated by the Callville Limestone of Utah, Arizona, and southeastern Nevada.

An environment of extensive mud and (or) sand accumulation, associated with carbonate deposits, occurs in southwestern White Pine County (loc. 66) and in northern Clark County (loc. 272) near the west margins of a basin. It is interpreted as a shelf environment where sinking was slow, and it reflects proximity to a nearby area of uplift. Most of the detritus is in the lower part of interval A (*Jensen Member* of Chainman Shale; *Indian Springs Member* of Bird Spring Formation) as defined in this paper. According to some stratigraphers (Webster and Lane, 1967; Dunn, 1970), however, these strata should be considered the product of a Late Missippian rather than an Early Pennsylvanian environment.

The depositional environment in west-central Nevada and southwestern California, which is considered a eugeosynclinal belt, can be interpreted only from a few scattered sections considered to represent interval A. The abundance of mudstone and presence of sandstone in the thick section of the Pumpernickel Formation (approximately 5,000 ft) at Battle Mountain, Nev., suggests that the area was sinking rapidly. Basalts, pillow lavas, and pyroclastics in many places indicate widespread volcanic activity. Bedded chert, abundant in both the
Val forms a part of the lithologically similar Bird Spring Interval is absent in the Callville Limestone of southern Arizona negative areas to the east. The movements on the Antler belt seems to have been relatively great, though irregular, and was accompanied by much volcanism. The rate of sedimentation at times and in some places probably lagged behind the rate of sinking, as testified by some chert deposits interpreted as deep-water types.

In the southern part of the Nevada and southern California region, reconstruction of paleogeographic features has been attempted by constructing a palinspastic map (fig. 68). By restoring physiographic forms a distance of 25 miles to the southeast in the area between the right-lateral faults of the Las Vegas shear zone and the Death Valley–Furnace Creek fault, 75 miles to the southeast in the area west of the Death Valley–Furnace Creek fault, and 60 miles southwest along the left-lateral Garlock fault, a positive element may be shown to extend far south from the Antler belt. In Morrow time, therefore, the eugeosynclinal area west of the Antler belt west probably was completely separated from the Arizona negative areas to the east. The movements on which this restoration is based are shown in terms of estimated displacement of locality points in figure 69.

**INTERVAL B**

**FORMATIONS INCLUDED**

Interval B is represented in the Nevada and southern California region by parts of several formations (pl. 13). Although biostratigraphic evidence indicates that the interval is absent in the Calville Limestone of southern and eastern Clark County, Nev. (Welsh, 1958), the interval forms a part of the lithologically similar Bird Spring Formation to the west and north, where it is identified by fossils of Atoka Age. Likewise, in the Ely Limestone of east-central Nevada and in both the Battle Formation and the Highway Limestone of central Nevada, the interval has been recognized on the basis of Atoka fossils; elsewhere in the region its presence is inferred on lithologic or thickness data.

In the lower part of the Tippipah Limestone at the Nevada Test Site interval B is believed to be present, although supporting biostratigraphic evidence has not been found. The Pumpernickel Formation at Battle Mountain probably includes this interval as well as interval C, indicated by conodont studies (Roberts, 1964, p. A44), and also interval A; but boundaries between the intervals cannot be differentiated in that area. Other rock sequences believed to include Atoka age rocks, assigned to interval B, are the Keeler Canyon Formation of the Inyo Mountains, the Oro Grande Formation of the Barstow area, and the Garlock Formation of the El Paso Mountains— all in southern California. Likewise, a part of the unnamed limestone sequence in the Mount Morrison area of the eastern Sierra Nevada is considered to be interval B.

**STRATIGRAPHIC RELATIONS**

Correlation of interval B across southern Nevada and westward into the Mojave Desert of California seems clear cut, for reasonably definite boundaries can be established in most places on the basis of the distribution of Atoka fossils. The lack of these fossils in the Calville and Supai Formations farther east is believed to be evidence that the seas had regressed northwestward during Atoka time. Correlations can be demonstrated in the largely carbonate rock sections of the Ely Limestone throughout east-central Nevada, but among the several formations containing volcanic rocks, chert, and other materials deposited in the eugeosyncline in west-central Nevada and southern California, fossils are scarce and correlations are necessarily more tenuous.

The largely detrital rocks of the Battle Formation of the Antler Peak area interfinger northwestward with carbonate rocks of the Highway Limestone in Edna Mountain in central Nevada; both formations contain Atoka fossils (Roberts, 1964, p. A32). In the Osgood Mountains, interbedded conglomerates and limestones are similar to those of the Battle and Highway Formations (Roberts, 1964, p. A31) and are assigned to interval B on this basis. Conglomeratic units in central and southern Lander County also are assigned to interval B (pl. 6); these units may not be exact time equivalents of the Battle Formation.

The lowest part of the Keeler Canyon Formation of the Inyo Mountains, which is assigned to interval B, contains not only *Fusulinella* in the basal 200 feet of carbonate rock but also abundant black chert nodules that are correlated with similar nodules in deposits of interval B in the Bird Spring Formation of the Providence Mountains, Calif. (loc. 140).
UPPER BOUNDARY OF INTERVAL B

In the highly fossiliferous carbonate rocks of southern and east-central Nevada, determination of an appropriate upper boundary for interval B is relatively easy. In general, the unit is characterized by the presence of the genus *Fusulinella*, and its upper limit is marked by the first appearance of *Fusulina*. In the Egan Range, in both Lincoln and White Pine Counties, Nev., this faunal change coincides with a lithologic change from thick-bedded micrite and skeletal limestone to thin-bedded silty limestone and cherty limestone, and in White Pine County a similar change has been reported (Bissell, 1964).

In most areas where diagnostic fossils are scarce or absent, accurate determination of the Atoka boundary, which elsewhere approximates the interval boundary, does not seem possible. This situation is encountered in the Pumpernickel Formation of Nevada and in the Garlock and Oro Grande Formations of California. In some other areas, such as central Nevada where the Battle Formation and Highway Limestone crop out, the top of interval B is marked by an erosional surface (Roberts, 1964, p. A27) and by a hiatus.

THICKNESS TRENDS

Intervals A and B have appreciably different thicknesses in various parts of the Nevada and southern California region (pls. 3A, 6A). In southeastern Nevada the Calville Limestone contains no strata of interval B, and therefore the southeastern zero isopach for interval B within Clark County is considerably north and west of that for interval A. In east-central Nevada, where a single elongate basin had occupied the miogeosynclinal area during the time of interval A, the isopachs for interval B delineate two small basins along the east side of the Antler positive element.

No sediments of interval B occur near the Sonoma Range (loc. 311) in west-central Nevada, and rocks of Late Pennsylvanian age rest upon Cambrian strata in that locality. Nearby to the east and north, however, the Battle Formation and the Highway Limestone were deposited during interval B time, leaving a thin sequence of strata where none had accumulated earlier in the period. This record shows an encroachment of interval B on the Antler positive area and a breaching of its north end.

About 200 feet of carbonate rocks of the Keeler Canyon Formation in the Inyo Mountains of southern California (loc. 163) and comparable thicknesses of strata in several other places in that area are considered to represent interval B. Because no deposits of interval A are known in this area, the presence of interval B is interpreted as the record of a transgressing sea.

LITHOFACIES TRENDS AND SOURCES OF SEDIMENT

Lithofacies trends of interval B in the Nevada and southern California region are, in general, similar to those of interval A with a few notable differences (pl. 6B). Most of eastern and southern Nevada was covered by carbonate deposits, and a belt of sandy and muddy carbonate sediment developed along the southeast margin of the area during deposition of both intervals. The belt of maximum detritus, however, had shifted northwestward during the time of interval B from its earlier location, presumably as the result of a regressing sea. Farther north, in east-central Nevada, two small areas of relatively high mud content, within a region of generally pure carbonate rock, coincide with an area between two basins where the interval is thin. The source of the mud is not clear, but the nearest and most likely place was the Antler belt to the west.

A considerable amount of sand accumulated near and across the north end of the Antler belt in north-central Nevada, forming a facies sequence of sandstone adjacent to the Antler uplift that grades northeastward to limy sandstone and sandy limestone. The dominantly detrital sediments, which consist of conglomerates composed of chert, quartzite, and volcanic material, interfinger with carbonate rocks of the Highway Limestone, which contains fusulinids (Roberts, 1964, p. A31). This sequence suggests a source to the southwest, presumably the Antler belt. Similar rocks have been identified in central and southern Lander County; their distribution, as well as that of rocks assigned to interval E, suggests that through much of Pennsylvanian time source areas in the Antler belt may have been islands of low relief rather than a continuous mountain chain. Allochthonous deposits from farther west, including bedded chert, volcanic rocks, greenstone, and other types of material typical of eugeosynclinal deposits, border the Antler belt on the west, and in places they rest on the autochthonous sandstone.

In southern California interval B is represented by marine carbonate rock both in the Keeler Canyon Formation of the Inyo Mountains and along the east side of the Death Valley–Furnace Creek fault in the eastern Mojave Desert (fig. 69). South and west of these deposits are eugeosynclinal rocks, including chert and volcanics in the El Paso Mountains near the Garlock fault and volcanic rocks of the Calico Mountains south of the Garlock fault. These rocks probably are far removed from their original sites, the Calico Mountains volcanic sequence clearly having been moved eastward some tens of miles.

ENVIRONMENT OF DEPOSITION

The environment of deposition throughout most of the Nevada and southern California region during the time of interval B probably was not greatly different from that during the time of interval A. Southern and eastern Nevada continued as a site of widespread carbonate deposition under normal marine conditions. A similar
environment is represented by the carbonate rocks in neighboring parts of southern California; it probably represents either shelf or miogeosynclinal deposits that spread westward during disappearance of the ridge or positive element that formerly extended south from the Antler belt (pl. 3A). Possibly, however, these deposits were transported tectonically from the southeast to their present positions during post-Pennsylvanian time.

A deeper water, offshore environment apparently persisted in Nevada west of the Antler positive element and in parts of southern California during all of interval B time as represented by volcanic rocks and chert deposits in the Pumpernickel Formation at Battle Mountain and in areas to the south in Nevada and in the Garlock Formation in the El Paso Mountains and the Pennsylvanian sequence in the Calico Mountains of southern California.

At the northern end of the Antler positive element in Nevada where deposits of interval A are absent (Battle Mountain, Edna Mountain, Osgood Mountains), lithologic and faunal evidence indicates a marine environment during interval B time. Conglomerates in these formations contain lime lenses with fusulinids and are interbedded with graded beds of sandstone and siltstone that are probably turbidites. Shallow seas also probably surrounded island areas to the south along the Antler trend.

PALEOTECTONIC IMPLICATIONS

In interval B time, continued or renewed movement of the Antler positive element apparently occurred near the north end, where uplift seems to have produced conglomerates of the Battle Formation. These conglomerates finger northeastward into limestones of the Highway Limestone of Edna Mountain and elsewhere (Roberts, 1964, fig. 14, p. A32). Farther south the Antler belt apparently was actively rising, as shown by the coarse detritus that it furnished into the bordering area to the east.

A major tectonic change that may have occurred shortly before the time of interval B was the sinking of the positive area that barred the sea from southeastern California in interval A time at the west edge of the miogeosyncline. (Compare pls. 3A and 6A with respect to zero isopachs and thickness figures in Mojave Desert—southern Nevada area.) This barrier, which presumably extended southward from the Antler positive element, was lost permanently at this time (pl. 15A), and the area became an extension of the miogeosyncline.

INTERVAL C
FORMATIONS INCLUDED

The same formations that contain interval A or B, or both, also include interval C in most parts of Nevada and southern California (pl. 13). In southern Nevada interval C forms a part of the Callville Limestone, the Bird Spring Formation, and the Tippipah Limestone. Farther north it includes the upper part of the Ely Limestone, although some authors (for example, Bissell, 1964) have locally referred strata here assigned to interval C to the Hogan Formation. At Moorman Ranch in White Pine County (loc. 66), where the rocks of Des Moines age are unnamed, Hogan Formation is used in this publication.

In western Nevada interval C is represented by the upper part of the Pumpernickel Formation. In southern California the interval is referred to part of the Oro Grande Formation (loc. 203, Barstow area), part of the Garlock Formation (loc. 143, El Paso Mountains), and part of an unnamed limestone sequence at Mount Morrison (loc. 171, eastern Sierra Nevada).

STRATIGRAPHIC RELATIONS

In the largely carbonate rock sequence of Pennsylvanian age in southern and eastern parts of Nevada, biostratigraphic evidence is available in most places to establish the presence or absence of strata of Des Moines age. Rocks of Des Moines age are assigned to interval C throughout the area.

In the eugeosynclinal belt of western Nevada and southern California, faunas of Des Moines age have not been reported from the Garlock Formation in the El Paso Mountains, Oro Grande Formation in the Barstow area, Keeler Canyon Formation in the Inyo Mountains, and unnamed limestone of the Mount Morrison sequence. Precise correlations are not possible, and the presence of rocks of Des Moines age assigned to interval C is inferred on the basis of rock types, of fossils of younger or older Pennsylvanian age in the same sequence, or of other suggestive criteria. At Battle Mountain, Nev., however, Des Moines fossils were recorded from the upper part of the Pumpernickel Formation (Roberts, 1964, p. A44).

UPPER BOUNDARY OF INTERVAL C

Where biostratigraphic evidence is available, as in the dominantly carbonate rocks of southern and eastern Nevada, the upper boundary of interval C can be determined at the top of the Des Moines with a considerable degree of assurance. The first appearance of Triticites is used to mark this boundary in the Callville and Bird Spring Formations and the Ely Limestone in Nevada.

Physical criteria for mapping the interval boundary are also available at numerous places. In the Frenchman Mountains (loc. 244) a chert-pebble conglomerate of the Callville Limestone occurs at the contact of intervals C and D, and rocks of Missouri age, representing interval D, are sandier and more crossbedded than those of Des Moines age in interval C (Welsh, 1959). In the Muddy Mountains (loc. 243) the upper boundary of interval C is indicated by a change from limestone to fine-grained sandstone in the Bird Spring Formation; in the Mormon Mountains (loc. 242), by a change to sandy cherty
limestone; and south of Las Vegas, by an increase in siltstone and fine-grained sandstone. In the Ely Limestone of Elko County the boundary is an unconformity, above which is an increase in the proportion of detrital sediment and a chert-pebble conglomerate. At Moorman Ranch (loc. 66) in White Pine County, Nev., the Hogan Formation of Bissell (1964, p. 578) is unconformably overlain by the Riepe Spring Limestone of Pennsylvania (Wolfcamp) age.

No information is available concerning the precise ages of intervals C and D in the Tippipah Limestone of the Nevada Test Site (loc. 162) and in the Bird Spring Formation of the Mojave Desert in California. Likewise, a good basis for picking interval boundaries is not available in most of the volcanic and chert sequences of western Nevada and southern California. An exception is at Battle Mountain, Nev., where the Pumpernickel Formation that includes interval C is overlain by the lithologically distinct Havallah Formation assigned to intervals D and E. The upper boundary of C at the top of the Pumpernickel is a “possible disconformity” (Roberts, 1964, p. A45), and strata above the contact are mainly sandstone, pebbly sandstone, conglomerate, chert, and mudstone interpreted as turbidites.

In the Garlock Formation of the El Paso Mountains, Calif., the top of member 9 of Dibblee (1952, p. 15-19) is used as the upper contact of interval C because it is marked by a lithologic change from mudstone and bedded chert to quartzite that is believed to correlate with the basal Havallah in Nevada (Roberts, 1964, p. A44). The validity of this boundary in Nevada is uncertain, however, for the Pumpernickel and Havallah cannot be consistently separated in some areas, and their status as distinguishable formations has been questioned (J. H. Stewart, written commun., 1972).

THICKNESS TRENDS

During the time of interval C, sedimentation in eastern Nevada was greatest in a north-trending series of basins. The Bird Spring–Ely basin, extending from the southern Las Vegas Range (loc. 250) northward to Cave Valley (loc. 262), remained as an elongate feature paralleling the Antler Belt lying to the west, but the axis of the basin was farther east than during the time of interval B. In deep parts of this basin, sediment in excess of 800 feet accumulated.

In east-central and northeastern Nevada three small elliptical basins containing deposits as much as 1,000 feet thick replaced the single basin of earlier Pennsylvanian time. Moorman Ranch (loc. 66), Cherry Creek Mountains (loc. 127), and the central Pequop Mountains (loc. 168) are located near the axes of these basins. A northeast-trending area, between the southern large basin and the northern smaller basins of the miogeosynclinal belt, is delineated by a zero isopach on the interval C isopach map (pl. 7A). Thinning in this area may be the result of post-Pennsylvanian uplift and erosion, but more likely the area was a positive element during the time of interval C, for sand is present within carbonate deposits along its southeastern margin (pl. 7B).

The Antler belt through interval C time probably was a low linear positive element west of the miogeosynclinal belt; no overlapping rocks comparable to those of interval B have been identified at the northern end.

LITHOFACIES TRENDS AND SOURCES OF SEDIMENT

Largely carbonate rocks in eastern Nevada have a distribution similar to that of interval B; however, the mud content is appreciable along the east margin of the Antler positive element in western and southern Elko Counties, western White Pine County, and northwestern Lincoln County, and sand is a notable constituent of the limestone in southwestern Elko County and northwestern White Pine County. These relations probably indicate a source of the detritus within the Antler positive area to the west. A local source area in east-central Nevada is suggested by detrital sediments along the southeast margin of an area defined by a zero isopach in the vicinity of the Snake Range in east-central Nevada and Utah.

In extreme southeastern Nevada, a broad northeast-trending belt of sandstone and carbonate rock and a narrow parallel belt of mudstone and carbonate rock to the northwest indicate by their positions, relative to those of similar rocks of interval B, a northwestward expansion of detrital sediments during interval C time. The distribution of detrital sediments and the southeastward extension of interval C (pl. 7) beyond the probable depositional limit of interval B (pl. 6) suggest that the sediment supply from an eastern source, believed to be in southwestern Utah, increased during C time and was delivered into a transgressing sea.

Deposits of the eugeosynclinal belt of Nevada and California include volcanics, chert, and other rocks like those deposited earlier in the Pennsylvanian in this region.

ENVIRONMENT OF DEPOSITION

About the same environments of deposition prevailed in Nevada and southern California in interval C time as in interval B time. Newly uplifted areas along the Antler positive element in central Nevada and across the Nevada-Utah State line near the present Snake Range (loc. 264) apparently caused local influxes of detrital sediment in widespread local areas; however, a shallow clear-water environment conducive to the deposition of calcium carbonate still prevailed throughout most of eastern and southern Nevada. Sand and some mud continued to accumulate along the nearshore areas, which extended farther south than before. These deposits merged seaward into the dominantly carbonate facies.
The western part of the region apparently continued as a source of volcanism and a repository for volcanic materials, bedded cherts, and some types of detrital sediment. Turbidity currents seemingly were not particularly active in Nevada during this part of the Pennsylvanian.

**PALEOTECTONIC IMPLICATIONS**

Principal paleotectonic elements in Nevada and southern California during the time of interval C consisted of local uplifts along the north-trending Antler belt in central Nevada and on a relatively small positive element in east-central Nevada; other uplifts that affected the region were in adjacent parts of west-central Utah (pl. 7A) on the Piute positive element.

Principal areas of depression were several small basins in the broad miogeosynclinal area east of the Antler belt and in the eugeosyncline west of the Antler belt. Tectonic movements continued at about the same rate as before, as indicated by thicknesses of sediments accumulated. Volcanism continued in parts of the western seaway.

**INTERVAL D**

**FORMATIONS INCLUDED**

Interval D is represented by part of the Bird Spring Formation, some of the Tippipah Limestone, and the lower half of the *Strathearn Formation*, Etchart Limestone, and Havallah Formation in Nevada; it forms part of the Bird Spring, Keeler Canyon, Oro Grande, and Garlock Formations in southern California (pl. 13). In the Egan and Golden Gate Ranges of Lincoln County, Nev., interval D is represented in the lower part of an unnamed rock sequence overlying the Ely Limestone.

**STRATIGRAPHIC RELATIONS**

In stratigraphic sections of the miogeosynclinal belt of eastern and southern Nevada interval D is generally recognized by the presence of a Missouri fauna. Diagnostic fossils have been found in the Bird Spring Formation and the *Strathearn Formation* at numerous localities, and rocks of Missouri age are inferred to be present throughout a wide area occupied by these formations. On the basis of similar lithologies, correlation of interval D has been extended into the Mojave Desert area of southern California.

Within the Callville Limestone of southern Nevada a hiatus occurs between intervals C and E; the absence of rocks of Missouri age (interval D) has been determined from the faunal record (Welsh, 1959). Farther north, in the Cortez Mountains, Shoshone Range, Battle Mountain, and Edna Mountain of north-central Nevada, the lower parts of the Brock Canyon Formation and Antler Peak Limestone are mapped entirely as interval E, although basal units probably are partly of Missouri age.

The eugeosynclinal belt in western Nevada and adjoining parts of California includes interval D throughout much of the area, but in most places definite evidence of a Missouri age for rocks assigned to the interval is difficult to establish. Part of the Keeler Canyon Formation in the Inyo Mountains contains *Triticit*es of Upper Pennsylvanian types (Merriam and Hall, 1957, p. 7); these strata are considered, in part, of Missouri age and are assigned to interval D. Elsewhere within the eugeosyncline, as in the Oro Grande Formation of the Barstow area and in the Garlock Formation of the El Paso Mountains, Calif., and in other sequences, the interval is only in part dated by fossils. Correlation by lithology has been resorted to in some places; thus, the Jory Member of the Havallah Formation has been correlated with member 10 of the Garlock Formation (Roberts, 1964, p. A49). Fossils of Atoka age have been found near the base of the Havallah, but they are, in part, abraded and are considered to be reworked (Roberts, 1964, p. A48).

**UPPER BOUNDARY OF INTERVAL D**

Neither a physical break nor a marked change in lithology between intervals D and E is recorded in the Nevada and southern California region. Biostratigraphic evidence is used in most places to locate the upper boundary of interval D. In the Bird Spring Formation of northern Clark County, Nev., the first occurrence of *Triticit*es of Virgil age is considered the base of overlying interval E. This horizon can be recognized from south to north between Apex (loc. 248) and Arrow Canyon (loc. 247) and westward to the Spring Mountains area (locs. 184, 93). In the Tippipah Limestone at the Nevada Test Site to the northwest (loc. 162) a Late Pennsylvanian fauna containing *Triticit*es has been recorded at 1,625 feet above the formation base (Johnson and Hibbard, 1957, p. 360-363), but the position of the Missouri-Virgil boundary is indeterminate there.

In east-central Nevada where the interval boundary is within the lower part of the *Strathearn Formation*, the boundary at most places is placed below the lowest Virgil species of *Triticit*es, as is done farther south. In some parts of that area, however, including northeastern Elko County, most or all of interval D is missing. In the unnamed Pennsylvanian sequence of the Egan Range (loc. 230) and of Cave Valley (loc. 262), in Lincoln County, biostratigraphic evidence is available for locating the boundary at the Missouri-Virgil contact (Bissell, 1964, p. 593; Brill, 1963, pl. 1, fig. 7).

Where diagnostic fossils are absent, the series boundary, normally used as the interval boundary, cannot be accurately located. In many sequences of the eugeosynclinal belt of western Nevada and southern California, including the Oro Grande Formation, the Garlock Formation, and others, this lack of a con-
sistent boundary makes meaningless any thickness determinations for interval D.

THICKNESS TRENDS

Basins of deposition in eastern Nevada (pl. 8A) were much more restricted and received considerably less sediment during the time of interval D than during any of the earlier Pennsylvanian intervals. The maximum thickness of sediments (500-600 ft) deposited in the Bird Spring–Ely basin was little more than half the amount accumulated in that area during interval C time. Farther north, interval D is absent, which suggests that at least two basin areas of interval C no longer were subsiding.

In the northeast corner of Nevada, interval D is missing within a large area trending from the Antler positive element northeastward (pl. 8).

LITHOFACIES TRENDS AND SOURCES OF SEDIMENT

Limestone similar to that of earlier Pennsylvanian times was deposited throughout the basins of southern and east-central Nevada during the time of interval D. In the marginal parts of these much restricted basins, accumulations of terrigenous detritus were smaller than in previous intervals. Only in a few places did sand or mud form appreciable parts of the sediment.

The principal area where sand accumulated was off the north end of the Antler positive element in western Elko County, Nev. The presence of sandy limestone of the Strathearn Formation suggests that nearby land areas in this part of the Antler region were still being actively eroded; farther south along the belt, no similar evidence of eroding land is recorded.

On the west side of the Antler positive element at Battle Mountain and in adjacent areas, detrital rocks of the allochthonous Golconda plate occur in interval D. Conglomeratic sandstones, conglomerates, and other rocks of the Havallah Formation contain sole markings and related structures and are interpreted as turbidites (Roberts, 1964, p. A45); no crossbedding or ripple marking is recorded.

In the southern California part of the eugeosyncline, bedded chert is included in the Oro Grande Formation of the Barstow area, and volcanics are major constituents of the Calico Mountains sequence, as in older intervals of the Pennsylvanian. A major change in lithology is recorded by rocks assigned to interval D in the Garlock Formation of the El Paso Mountains, where essentially pure sandstone (member 10) overlies beds dominantly of mudstone (Dibblee, 1952) of interval C. Farther north, in the Keeler Canyon Formation of the Inyo Mountains, interval D is represented by a mixture of mudstone and limestone.

ENVIRONMENT OF DEPOSITION

The depositional environment represented by interval D in the miogeosynclinal belt of eastern Nevada is considered to have been normal marine. This environment is indicated by nearly pure limestone within the basins, only local areas of terrigenous detritus along margins of the basins, and a fauna of brachiopods, fusulinids, and other marine forms.

The environment of parts of the eugeosynclinal area west of the Antler positive element contrasted strongly with the environment of the carbonate-depositing seas farther east. At Battle Mountain, Nev., interval D sediments consisted of sands containing quartz, chert, and some feldspar grains, deposited at least in part by turbidity currents which suggest appreciable local relief of the sea floor and deep water. In the Calico Mountains and in other localities, volcanic rocks of interval D indicate an area of active volcanism nearby. Elsewhere in the western part of the region, as in the Inyo Mountains, detrital deposits and their sedimentary structures imply shallower water in which current action helped distribute the sediment. These strata include silty to pebbly limestone, chert-pebble conglomerates, and pink to maroon shaly mudstone.

PALEOTECTONIC IMPLICATIONS

Interval D probably represents one of the tectonically least active parts of Pennsylvanian time in the region of Nevada and southern California. Basins of the eastern seaway sank at a reduced rate and were gradually filled with mostly carbonate sediment. The entire seaway was much reduced in size, and relatively little terrigenous detritus accumulated in or around the basins. Probably the slow rate of deposition reflects a general lack of uplift and erosion in surrounding areas.

At the north end of the Antler positive element, renewed or continued uplift may have been the cause of relatively greater detrital sedimentation in that area. Possibly, a northeast-trending prong of the Antler element in Elko County was uplifted. Whether this area was a positive element subject to erosion or was uplifted and eroded later is not known. The absence of the interval along much of the Nevada-Utah border may be due to withdrawal of the sea westward, which resulted in nondeposition over wide areas where rocks of earlier Pennsylvanian age were deposited.

Turbidites in the Battle Mountain allochthon are believed to indicate uplift of a source area northwest of the Antler belt, but the sediments moved mostly along the north-trending axis of the eugeosyncline, as shown by flute casts and other directional indicators.

INTERVAL E

FORMATIONS INCLUDED

Interval E in the area extending from southern Nevada to the eastern Mojave Desert of California is part of the Callville Limestone and Bird Spring Formation and of the Tippipah Limestone (pl. 13). Farther
north in Nevada it consists of parts of the *Strathearn Formation* in Elko County, the *Wildcat Peak Formation* in the Toquima Range, and the lower part of the Brock Canyon Formation in Eureka County and the Antler Peak Limestone in Battle Mountain and Edna Mountain. It also constitutes a part of the Etchert Limestone of the Osgood Mountains.

On the west side of the north-trending Antler belt in central Nevada interval E forms the upper part of the Jory Member of the Havallah Formation in Battle Mountain, Nev. In California interval E is within the Keeler Canyon Formation of the Inyo Mountains and is part of the Oro Grande Formation in the Barstow area and of the Garlock Formation in the El Paso Mountains. It is believed to be absent at Mount Morrison in the eastern Sierra Nevada.

**STRATIGRAPHIC RELATIONS**

Fossiliferous rocks of Virgil age occur in most sequences of Pennsylvanian rocks in Nevada and are assigned to interval E. *Triticites* of Virgil type and other diagnostic fossils have been reported from the following formations and places in Nevada:

1. Callville Limestone, southeastern Nevada.
2. Bird Spring Formation, northwestern Clark County.
5. Antler Peak Limestone, Osgood Mountains.

Elsewhere in both Nevada and California, interval E has been mapped on the basis of stratigraphic position beneath rocks containing fossils of Wolfcamp age, as follows:

1. Bird Spring Formation, eastern Mojave Desert, Calif.
2. Tippipah Limestone, Nevada Test Site.
3. Jory Member of Havallah Formation, Battle Mountain, Nev.
5. Rocks in western Lander and Nye Counties and Esmeralda County, Nev., that are correlated with the Havallah.

At a few places rocks assigned to interval E probably include some strata of Missouri age as well as strata of Virgil age (pl. 13), as discussed for interval D. These places are in the Shoshone Range, Battle Mountain, and Edna Mountain, Nev., where physical and faunal criteria for separating the intervals are lacking.

In the Garlock Formation of the El Paso Mountains in California, member 10 has been correlated on lithologic grounds with the Jory Member of the Havallah Formation (intervals D and E) in central Nevada. In the Oro Grande Formation of the Barstow area, California, interval E is inferred to be represented in the upper part. Fossils in the middle of the formation are of Pennsylvanian (?) age and the formation is overlain unconformably by the Fairview Valley Formation of Permian and Permian (?) age (Bowen, 1954).

**UPPER BOUNDARY OF INTERVAL E**

In most rock sequences of late Paleozoic age in the Nevada and southern California region, strata of Pennsylvanian or inferred Pennsylvanian age are overlain by Permian strata. The contact between these systems, which is also the upper boundary of interval E, is marked by physical change in some places but must be determined by biostratigraphic evidence elsewhere.

In southeastern Nevada, the Callville Limestone terminates upward at a disconformity; fossils of latest Virgil age are absent in this area (Welsh, 1959). The position of the hiatus is marked by a change from limestone of the Callville to dolomite of the Pakoon Formation; the interval boundary can be approximately indicated also by the position of the highest *Triticites* of Virgil age or the first occurrence of a Wolfcamp fauna.

In the area west and north of that occupied by the Callville Limestone, the Bird Spring includes both Pennsylvanian and Permian strata. Interval E in the Bird Spring Formation is overlain in places such as Mountain Springs (loc. 252) and northern Bird Spring Range (Welsh, 1960, p. 59) by a limestone-pebble conglomerate which probably is the same marker unit that occurs at this horizon throughout Grand Canyon, Ariz. The overlying Permian strata are carbonate rocks, in general much like the Pennsylvanian, but they locally include much siltstone near the contact. Fossils of Wolfcamp age occur above the contact in many places.

In the Bird Spring Formation of the Mojave Desert of California (Providence Mountains, loc. 140) the lowest occurrence of a Wolfcamp fauna (Thompson and others, 1946) is used to locate the interval boundary, and in the Tippipah Limestone at the Nevada Test Site (loc. 162) similar biostratigraphic data are employed (Johnson and Hibbard, 1957, p. 362).

Throughout most of east-central Nevada the upper boundary of interval E is located on faunal evidence. In the *Strathearn, Wildcat Peak*, Brock Canyon and other formations, either Late Pennsylvanian or Early Permian faunas or both serve to identify the contact, but physical criteria are generally lacking.

In western Nevada and southern California Permian Wolfcamp faunas have been reported in the upper part of the Jory Member of the Havallah Formation of Battle Mountain (Roberts, 1964, p. A48) and in the Keeler Canyon Formation of the Inyo Mountains (Merriam and Hall, 1957, p. 4-7), but no physical break is apparent. The upper boundary of interval E has been placed at the top of the Jory and within the Keeler Canyon. In the Garlock Formation of the El Paso Mountains, Calif., the
boundary is determined by lithologic correlation with the Jory Member of the Havallah Formation in Nevada (Roberts, 1964). In the Oro Grande Formation of the Barstow area, California, an unconformable surface overlain by pebble conglomerate at the base of the Fairview Valley Formation (Bowen, 1954) is considered the boundary.

THICKNESS TRENDS

The thickness trends in interval E (pl. 9A) are much like those in interval D, but, in general, the basins are more expanded and the sediments are thicker. The eastern and southern limits of the Bird Spring–Ely basin, for example, shifted into Utah and Arizona, and beginning late in Missouri time, the opposite margin extended a considerable distance northwestward of its earlier position. Maximum thickness of deposits in the Bird Spring–Ely basin was about 800 feet for interval E as compared with 500 feet for interval D. The gap shown between occurrences of interval E in northern Nye County may be erosional rather than depositional in origin.

Along the Antler positive element of central Nevada, marked differences between isopach patterns of interval D and interval E record notable overlap of interval E. This overlap is illustrated in the Toquima Range of northern Nye County where marine and nonmarine beds of the Wildcat Peak Formation cover older rocks northwestward. At the north end of the Antler element the marine Antler Peak Limestone and the mixed continental and marine Brock Canyon Formation (Muffler, 1964, p. 12) overlap southeastward and generally thin in that direction.

A prong outlined by the zero isopach continued to extend from the northern part of the Antler positive element northeastward as during the time of interval D. This prong had a narrow depositional trench or basin paralleling it along the southeast side. South of the depositional trough was a second prong, nearly parallel to the one at the northern end of the Antler positive element.

LITHOFACIES TRENDS AND SOURCES OF SEDIMENT

The lithofacies of interval E (pl. 9B) in Nevada have a distribution resembling more closely the facies relations of interval C (pl. 7B) than those of interval D (pl. 8B). The similarities and differences apply mainly to the miogeosynclinal belt; they are attributed partly to transgressions and regressions of the sea and partly to influx or lack of influx of detritus resulting from uplift or stable conditions in surrounding landmasses.

The interval E lithofacies map shows a broad belt of sandy limestone and calcareous sandstone along the southern and southeastern margins of the Bird Spring–Ely basin in Nevada and adjacent parts of Arizona. An area of argillaceous carbonate rock borders this belt along its northwestern or seaward side. Trends from sand to mud suggest a northwesterly direction of detritus transport in that area. A sandy carbonate belt in the Wildcat Peak Formation on the opposite or northern side of the basin suggests, at the same time, transport of detrital material southeastward from the Antler positive element of central Nevada. Detritus in northern Nye County includes gravel in chert and limestone conglomerates; some limestone boulders contain Carboniferous fossils.

In the northern half of Nevada, carbonate rocks occupy an elongate basin paralleling the northeast-trending prong from the Antler belt. The lack of appreciable detritus in this basin indicates that the adjacent area of the prong, where interval E is absent, probably was not a source of detritus. Farther west in north-central Nevada, however, a sequence of facies ranging from muddy limestone, to sandy limestone, to limy sandstone, to pure sandstone suggests that detritus in that area was being shed to the northeast from the north end of the Antler positive element.

Lithofacies of interval E in the eugeosynclinal belt seem to be similar to those of interval D. In the Inyo Mountains area of California, muddy limestone, silty to sandy limestone, and conglomerate containing chert and limestone pebbles continued to accumulate. In the Garlock Formation of the El Paso Mountains, Calif., fine-grained sandstone or quartzite and interbedded chert were formed, and in the Oro Grande Formation of the Barstow area, California, chert was formed. The region of Calico Mountains continued as an area of volcanism.

ENVIRONMENT OF DEPOSITION

During the time of interval E, clear relatively shallow seas containing an abundant marine fauna covered the basin areas in the miogeosynclinal belt of Nevada. Accumulating sediment was almost exclusively carbonate. In shelf areas bordering these basins much terrigeneous sand and mud derived from adjacent lands also was deposited — probably the result of high energy of very shallow waters.

In eastern Nevada the lithofacies patterns for each of the Pennsylvanian intervals are generally similar, although the positions of marginal belts of detrital sediment shifted back and forth with time. Minor fluctuations, showing cyclic character, are also recognizable throughout most of the sequence, but because they developed almost exclusively within a normal marine environment, they are less pronounced than cyclothems of the Pennsylvanian in Eastern United States, where marine and nonmarine deposits alternate.

An alternation of relatively thin bedded calcarenite with finer grained but thicker bedded nonclastic limestone in northeastern Nevada is thought to represent cyclic changes from shallow to moderately deep water in which the wave or current agitation increased or
decreased according to the water depth. Chert-rich and nearly chert-free zones seem unrelated to the major variations in the textures of the limestone and to the inferred depth changes (Dott, 1958). Thirty cycles have been recognized in the thickest parts of the Bird Spring Formation of southern Nevada (Welsh, 1959, p. 11-12).

Near the southern margin of the Bird Spring–Ely basin a northwesterly increase in bioclastic limestone and a corresponding decrease in oolitic limestone was interpreted by Welsh (1959, p. 75-78) as a change from shallow to deeper water deposition and is used, in part, to differentiate between the Callville and Bird Spring Formations. Along the so-called Las Vegas hinge line, which marks the position of an abrupt increase in thickness northwestward, a high percentage of sand and a large amount of crossbedding were believed by Welsh to indicate the former presence of currents and shoals in the vicinity of barrier islands.

In the Toquima Range of south-central Nevada, the Wildcat Peak Formation (Kay and Crawford, 1964, p. 441) contains fusulinid-bearing carbonate rock of Late Pennsylvanian age1 interbedded with conglomerates that include boulders as much as 2 feet in diameter. These conglomerates may have been deposited above sea level but, according to some geologists, "more likely are orogenic clastics deposited in the sea" (F. J. Kleinhampl, written commun., 1972). Whatever their genesis, the place of deposition was on the flanks of the Antler belt. Thus, an environment adjacent to a positive element is implied for this sequence of rocks on the northwest margin of the Bird Spring–Ely basin.

Farther north, in the Cortez Mountains of Eureka County, the Brock Canyon Formation contains much conglomerate, some dolomite, and some black shale containing plant fossils. Its depositional environment is interpreted as having alternated between continental and lagoonal or littoral (Muffler, 1964, p. 12). The coarse detritus probably was deposited subaerially; the siltstone and dolomite were possibly lagoonal.

The environment of deposition in the eugeosynclinal belt during the time of interval E probably was much like that of preceding Pennsylvanian time. Turbidity flows are recorded in the upper part of the Jory Member of the Havallah Formation in west-central Nevada, and some of the volcanic rocks in the Calico Mountains of southern California are believed to be this age also.

PALEOTECTONIC IMPLICATIONS

Renewed uplift in parts of the north-trending Antler positive element of central Nevada during the time of interval E is documented by considerable accumulations of detrital sediment, including some coarse con-glomerates, in the Wildcat Peak Formation along the east flank and in the Brock Canyon Formation at the north end. Basins in eastern Nevada subsided contemporaneously, and the seas transgressed along the east southeast and northwest margins of the Bird Spring–Ely basin.

Northeastern Nevada probably continued as a largely negative area despite the presence, as shown on the isopach map (pl. 9A), of areas where Upper Pennsylvanian deposits are absent. Uplift and erosion of these areas probably occurred in post-Pennsylvanian time, for deposits of nearly pure limestone surround them. The areas in easternmost Nevada, having no deposits, apparently remained much as they had been in interval D time.

Downwarping and volcanism in the eugeosyncline of western Nevada and southern California probably continued as in earlier Pennsylvanian time, but precise details in terms of time and place are sparse, both because of problems in dating the rock and because of few exposures and complex structure.

TOTAL THICKNESS OF PENNSYLVANIAN ROCKS

THICKNESS TRENDS

The total thickness pattern of Pennsylvanian rocks in Nevada and southern California reflects accumulation of deposits for most of the period in two mainly negative belts, separated by a north-trending belt called the Antler positive element (pl. 11). The subsiding belt to the east was a miogeosyncline; included within it were the elliptical Bird Spring–Ely basin in the south, a small subcircular basin near Moorman Ranch (loc. 66) in east-central Nevada, and a northwest-trending linear basin in western Elko County. Pennsylvanian rocks in the Bird Spring–Ely basin have a maximum thickness of 3,500 feet in the southern Egan Range (loc. 230), 2,950 feet in east-central Nevada, and 3,500 feet in western Elko County.

The belt on the west was a eugeosyncline whose isopach configuration cannot be defined, because as yet few reliable thickness measurements are known from this belt. Incomplete thicknesses have a magnitude of some thousands of feet. For example, the combined Pumpernickel and lower part of the Havallah Formations on Battle Mountain, Nev., are more than 6,000 feet thick (Roberts, 1964, p. A38-A46), and the lower part of the Garlock Formation in the El Paso Mountains of California is calculated to be about 10,000 feet thick (Dibblee, 1952, p. 15-19).

The Antler belt, which extends northward across central Nevada, apparently was largely positive throughout most of Pennsylvanian time, but the edge of the land area is not well defined, because of the sparse control available for various intervals (pl. 11).

1A fossil assemblage from near the base of the Wildcat Peak Formation collected by E. H. McKee and determined by C. H. Stevens suggests that Atoka, as well as Virgil, fossils are represented in the formation (E. H. McKee, written commun., 1972).
The absence of Pennsylvanian rocks in a small circular area that includes the Snake Range, near the east border of central Nevada, is interpreted as the result of post-Pennsylvanian erosion rather than Pennsylvanian uplift, because adjacent deposits of Pennsylvanian age are largely nondetrital.

**PALEOTECTONIC IMPLICATIONS**

Tectonic activities that produced the present distribution and thickness of Pennsylvanian rocks can be summarized as follows:

1. Moderate uplift of the Antler positive element occurred periodically along its length but especially near its north end. The most northern part, in northeastern Nevada, was submerged and covered by sediments during the last part of Pennsylvanian time. Periodically, the central and southern parts of the positive element may have been largely under shallow seas dotted with low islands.

2. The eastern seaway from time to time expanded and contracted southeastward and eastward as the result of periodic epigenic uplift and subsidence or of eustatic changes in sea level. Notable retreats of the sea were in pre-Morrow, Atoka, and Missouri times.

3. Deposits of the eugeosynclinal area to the west were notably coarser in the late part of the Pennsylvanian than in the early or middle parts. This increase in textural size probably resulted from uplift that caused greater relief of land areas, but it possibly also reflects a rapid sinking of the sea floor, as suggested by much evidence of deposition by turbidity currents in the Late Pennsylvanian. In the western part of the region, volcanism apparently was active at many times throughout the period.

**GEOLOGIC UNITS DIRECTLY ABOVE PENNSYLVANIAN SYSTEM**

Pennsylvanian rocks in the Nevada and southern California region are nearly everywhere overlain by rocks of Early Permian (Wolfcamp) age except in east-central Elko County, where Permian rocks of Leonard age were deposited on Pennsylvanian rocks of Atoka and possibly Des Moines age. Tertiary volcanics overlie rocks assigned to interval E of the Pennsylvanian in several areas in central and northern Nevada. Some of the complexities on the map of overlying rocks are the result of the Golconda thrust and other post-Paleozoic faults.

**REFERENCES**


Fritz, W. H., 1960, Structure and stratigraphy of the northern Egan


Pacific Northwest Region

By WILLIAM P. IRWIN

PALEOTECTONIC INVESTIGATIONS OF THE PENNSYLVANIAN SYSTEM IN THE UNITED STATES, PART I: INTRODUCTION AND REGIONAL ANALYSES OF THE PENNSYLVANIAN SYSTEM

GEOLOGICAL SURVEY PROFESSIONAL PAPER 853-R
CONTENTS

<table>
<thead>
<tr>
<th>Page</th>
<th>Intervals A-E — Continued</th>
</tr>
</thead>
<tbody>
<tr>
<td>329</td>
<td>Thickness trends</td>
</tr>
<tr>
<td></td>
<td>Lithofacies trends</td>
</tr>
<tr>
<td></td>
<td>Sources, environments of deposition, and paleotectonic implications</td>
</tr>
<tr>
<td>331</td>
<td>References</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**ILLUSTRATION**

[For listing of plates (in separate case) see volume “Contents”]

**FIGURE 70.** Map showing counties and other geographic features in Pacific Northwest region mentioned in text ............ 330
ABSTRACT

The Pennsylvanian of the Pacific Northwest region is sparsely represented at a few widely spaced localities by an assemblage of marine detrital and volcanic rocks, ribbon chert, and lenses of limestone. Locally, deposition probably was continuous from Late Mississippian into Early Pennsylvanian, as indicated by the upper part of the Baird Formation in northern California and by the upper part of the Coffee Creek Formation in east-central Oregon. In northern Washington, the Early — and perhaps younger — Pennsylvanian is represented by part of the long-ranging Chilliwack Group. Middle or Late Pennsylvanian deposition is indicated in Oregon where the fossil-plant-bearing Spotted Ridge Formation overlies the Coffee Creek Formation with slight angular discordance. The presence of sparse fossil plants in some of the Chilliwack Group and Mount Roberts Formation suggests that some strata in northern Washington may be broadly correlative in age with the Spotted Ridge Formation. Most of the Pennsylvanian is overlain by Lower Permian eugeosynclinal strata.

The area of deposition of Pennsylvanian rocks in the Pacific Northwest region probably was a broad mobile belt that lay between the Antler orogenic belt on the east and the deep ocean basin on the west. Detrital sediment may have been derived in part from actively emergent lands along the Antler orogenic belt, but much of the detritus probably was derived through local cannibalism of both preexisting and contemporaneous eugeosynclinal rocks within the mobile belt. Volcanic rocks accumulated at various places within the mobile belt during intermittent episodes of submarine, and perhaps subaerial, volcanism.

REGION DEFINED

The Pacific Northwest region consists of northern California, Oregon, and Washington. Within this region, data on Pennsylvanian rocks are available only from widely separated areas of outcrop in the Klamath Mountains of California, the Suplee area of east-central Oregon, and northern Washington.

PALEOGEOLOGY

UNITS UNDERLYING PENNSYLVANIAN

Pennsylvanian rocks of the Pacific Northwest region probably are underlain at most, if not all, places by Mississippian rocks, but boundary relations between the two systems are generally not known. In the Klamath Mountains of California (fig. 70), Mississippian strata crop out along strike in a belt about 40 miles long and are part of an eastward-dipping homoclinal Paleozoic sequence that locally includes Pennsylvanian strata. The Upper Mississippian (Visean) is represented by the Baird Formation (pl. 2), which chiefly consists of interlayered mudstone and pyroclastics and which ranges in thickness from 3,000 to 5,000 feet (Albers and Robertson, 1961). The Visean age is based on the presence of gigantoproductid brachiopods and other fossils in the middle part of the formation. Although the Baird formerly was thought to be entirely Late Mississippian, the presence of certain fusulinids strongly suggests that its upper part may be Early Pennsylvanian (Skinner and Wilde, 1965).

The Coffee Creek Formation, a correlative of the Baird, crops out in a few square miles in the Suplee area of east-central Oregon. Like the Baird, part of the Coffee Creek Formation contains the gigantoproductid brachiopod of Late Mississippian age, and part contains fusulinids of probable Early Pennsylvanian age (Skinner and Wilde, 1965). Deposition of both the Baird and Coffee Creek Formations presumably was uninterrupted from Late Mississippian to Early Pennsylvanian, as judged from the apparent lack of an obvious physical break between the middle and upper parts of the two formations.

In northern Washington, scattered limestone lenses that contain fossils of probable Pennsylvanian age occur in the Chilliwack Group, a terrane that generally is sparsely fossiliferous and structurally complex. Some of the Chilliwack Group may also be of Late Mississippian age, but the boundary is not known.

INTERVALS A-E

FORMATIONS INCLUDED

Pennsylvanian rocks of the Pacific Northwest region (pl. 11) are for the most part too poorly known to be subdivided precisely into separate intervals. Those in the Klamath Mountains of California are restricted to the
uppermost part of the Baird Formation (pl. 13). At a few localities the uppermost part of the Baird contains the fusulinids *Eostaffella* and *Pseudostaffella* that strongly suggest an Early Pennsylvanian (Morrow or Derryan) age, according to Skinner and Wilde (1965). These geologists believe that the Middle and Upper Pennsylvanian probably is missing, but that it may be represented by a few feet of unfossiliferous strata remaining at the top of the Baird. The fusulinid-bearing upper part of the Baird is tentatively considered to represent interval A.

In Oregon, Pennsylvanian rocks are found only in the Suplee area. There the Pennsylvanian rocks comprise the upper part of the Coffee Creek Formation and the Spotted Ridge Formation. Certain other rocks in eastern Oregon were formerly thought to be Pennsylvanian (Gilluly, 1937; Pardee, 1941) but are now considered to be Permian (Taubeneck, 1955; Bostwick and Koch, 1962). Part of the Coffee Creek Formation contains a fusulinid fauna similar to that of the upper part of the Baird (Skinner and Wilde, 1965), and thus it may represent interval A. The Spotted Ridge Formation overlies the Coffee Creek Formation with slight angular discordance (Merriam and Berthiaume, 1943). Fossil plants in the Spotted Ridge were once considered to be Early Pennsylvanian in age (Read and Merriam, 1940; Merriam and Berthiaume, 1943), but upon reexamination, an age assignment more precise than simply Pennsylvanian has been discounted (Mamay and Read, 1956). The interval or intervals represented by the Spotted Ridge Formation presumably are younger than interval A. Placement of the interval A-interval B boundary at the Coffee Creek-Spotted Ridge contact is arbitrary.

In northern Washington, small limestone lenses on Orcas Island (loc. 21) in San Juan County are thought likely to be Early Pennsylvanian in age, owing to the presence of the fusulinid *Eostaffella* and the foraminifer *Tetratax* (Danner, 1966). A Pennsylvanian age for limestone on San Juan Island (loc. 12) in San Juan County is suggested by the presence of certain poorly preserved Foraminifera (Danner, 1966). To the east on the mainland, in the Red Mountain and Black Mountain areas (loc. 10) in Whatcom County limestone thought likely to be Pennsylvanian contains the fusulinids *Eostaffella* and *Ozawainella*, and the foraminifer *Tetratax*, as well as a variety of other fossils (Danner, 1966). Other Pennsylvanian limestone lenses (locs. 25, 26, and 28) are reported along a northwest-trending belt in Snohomish and Skagit Counties (Danner, 1966) and at Twin Lakes (loc. 24; Misch, 1952) and Ridley Creek (loc. 31; W. R. Danner, written commun., 1970) in Whatcom County; all the mainland localities noted are referred to the *Chilliwack Group*. Pennsylvanian fossils have been reported from an unnamed limestone near Springdale (loc. 30) in southeastern Stevens County (Enbyysk, 1956). Limestone lenses and associated graywacke and argillite in northern Stevens County (loc. 29) are a southward continuation of fossiliferous rocks of the Mount Roberts Formation (R. G. Yates, oral commun., 1966) which, near Paterson in British Columbia, within a mile of the International Boundary, are considered to be Pennsylvanian(?) by Little (1960).
UPPER BOUNDARY OF THE PENNSYLVANIAN SYSTEM

The upper boundary of the Pennsylvanian is clearly defined nowhere in the Pacific Northwest region. In the Klamath Mountains the only probable Pennsylvanian is the upper part of the Baird Formation. The next youngest formation is the McCloud Limestone, which is chiefly, if not entirely, Early Permian in age and which crops out to the east of, and parallel to, the belt of Baird. The Baird Formation and McCloud Limestone have been mapped in detail only in the southern part of the belt, where most of the McCloud is separated from the Baird by a linear diorite intrusive mass within which most of the McCloud occurs as engulfed isolated masses (Albers and Robertson, 1961). In at least one locality, however, the intrusive mass apparently is absent between the two formations, inasmuch as Lower Pennsylvanian fusulinid-bearing limestone in the upper part of the Baird is separated from the McCloud by not more than 100 feet of tuffaceous sandstone. The lower four and most of the fifth of eight faunal zones in the McCloud are also recognized in beds adjacent to the west that are lithically indistinguishable from the Baird (Skinner and Wilde, 1965). The lowest part of some occurrences of the McCloud is not clearly Permian and may be Pennsylvanian (Skinner and Wilde, 1965). Problems arising from these data cannot be resolved now, and the upper boundary of the Pennsylvanian in the Klamath Mountains will herein be tentatively considered to be the base of the McCloud Limestone.

In the Suplee area of Oregon, the upper boundary of the Pennsylvanian is tentatively considered to be the top of the fossil-plant-bearing Spotted Ridge Formation, which is overlain disconformably by the Coyote Butte Formation of Early Permian age (Merriam and Berthiaume, 1943), but the interval represented by the uppermost Spotted Ridge is not known. In northwestern Washington, the limestone lenses of questionable Early Pennsylvanian age are thought to occur disconformably below strata that contain plant fragments; higher in the section these strata include Lower Permian limestone (Danner, 1966). It is not known whether these plant-bearing strata are equivalent to the Spotted Ridge Formation of Oregon or whether they are Pennsylvanian in age.

THICKNESS TRENDS

The Pennsylvanian rocks of the Pacific Northwest region constitute only a small part of the complexly deformed eugeosynclinal terrane of Paleozoic and Mesozoic rocks. Most of this terrane is concealed by a blanket of uppermost Mesozoic and Cenozoic rocks. Erosional windows in this blanket of younger rocks expose only a small part of the total extent of Paleozoic rocks and, within these windows, the rocks of Pennsylvanian age are recognized only in three general areas: the Klamath Mountains of northwestern California, the Suplee area of east-central Oregon, and northern Washington. However, regional lithic and structural trends suggest that the Pennsylvanian rocks are part of a belt of Paleozoic rocks that is virtually continuous beneath the blanket from window to window. These trends suggest that the belt continues southeast from the Klamath Mountains into the northern Sierra Nevada, and northward from Washington into Canada.

The thickness and continuity of the Pennsylvanian within this belt of Paleozoic rocks are not clear. Recognition of Pennsylvanian rocks within the Paleozoic terrane is difficult, owing to a lack of distinctive lithology, the scarcity of fossils and the questionable age assignment of those that have been found, and the structural complexity. Microfossils thought likely to be Early Pennsylvanian in age occur in limestone lenses in the upper part of the thick section of detrital sedimentary and volcanic rocks at some localities, and in the upper part of a thick carbonate section at another locality; lower parts of these sections contain Late Mississippian mega fossils. The position of the time boundary between the rocks of the two ages is not known. Lower Permian limestone lenses commonly are found in the vicinity of the limestone of probable Early Pennsylvanian age, but the two generally are separated by detrital sedimentary and volcanic rocks. At some localities these intervening rocks contain fossil plants and are thought to lie disconformably on the strata that include the limestone lenses of probable Early Pennsylvanian age. The age of these intervening strata is not known at most localities; but in the Suplee area, where they are most fully developed, the fossil plants are considered to be Pennsylvanian. Owing to the uncertainties in position of time boundaries and to lack of precise age data, the thickness of the Pennsylvanian can be only roughly estimated.

In the Klamath Mountains, where only the strata thought to be of interval A are recognized, the total thickness of the Pennsylvanian probably is no more than a few hundred feet. In the Suplee area, the thickness of strata thought to be of interval A is not known even within wide limits. The Pennsylvanian plant-bearing formation that overlies the beds of interval A in the Suplee area is as much as 1,500 feet thick (Merriam and Berthiaume, 1943), but whether the formation is entirely Pennsylvanian is not clear. In northwestern Washington, the Pennsylvanian rocks on Orcas Island in San Juan County are thought to be between 500 and 1,000 feet in thickness, and to the east, in the foothills of the northern Cascade Mountains, between 1,000 and 2,000 feet in thickness (Danner, 1966, p. 66).

LITHOFACIES TRENDS

The Pennsylvanian of the Pacific Northwest region is dominantly of the eugeosynclinal facies. In the Klamath
Mountains, the uppermost part of the Baird, tentatively assigned to interval A, consists of interbedded mudstone, volcanic rocks, and minor limestone, whereas strata of equivalent age (part of Coffee Creek Formation) in the Suplee area are chiefly limestone. The Spotted Ridge Formation in the Suplee area consists of mudstone, sandstone, and conglomerate, thin chert beds and minor amounts of limestone. In Washington, most of the limestone lenses of probable Pennsylvanian age are in detrital sedimentary sequence with associated volcanic rocks.

**SOURCES, ENVIRONMENTS OF DEPOSITION, AND PALEOTECTONIC IMPLICATIONS**

Rocks thought to represent interval A probably were deposited in a sea bordering the western continental margin, and although specific evidence is lacking, the predominant source of detrital sediments probably was tectonic lands to the east. Abundant andesitic pyroclastic rocks in the Klamath Mountains area indicate a proximity to vigorously active volcanoes, and volcanic activity is indicated in Washington. A lack of volcanic rocks in the dominantly carbonate section of the Suplee area indicates local volcanic quiescence. That the sea probably was warm and shallow is suggested by the presence of corals among many other varieties of fossils in the strata that closely underlie and that are transitional into interval A in this area. A specific source area for the detrital rocks of interval A is not indicated by lithologic evidence; but the source area is inferred to have lain to the east, and may have been in highlands generated during the Antler orogenic event. Scattered volcanic islands may have been local contributors of pyroclastic detritus. Uplift, probably mild, may have terminated interval A. In the Klamath Mountains area, local erosion of interval A and a hiatus in deposition is suggested between interval A and the base of the McCloud Limestone. In the Suplee area, the presence of plant fossils of Pennsylvanian age in strata that disconformably overlie beds of probable interval A might indicate uplift and the nearby presence of subaerial conditions. Similar indication of subaerial condition is found in northeastern Washington. A return to shallow marine conditions near the end of the Pennsylvanian is indicated by the widespread occurrence of Lower Permian limestone in the Pacific Northwest region.

**REFERENCES**


INDEX

Arizona—Continued

Ardmore Basin. See Texas Panhandle and Oklahoma.
Arkoma basin. See Texas Panhandle and Oklahoma.
Arrodeo Limestone Member 121
Arkansas, abstract 357
Bloomer syncline 158
Broken Bow-Benton uplift 158, 172
Camplin dome 164
Gulf Coast Plain 159, 165, 172, 173
interval A. environments 163
formations 163
lithofacies trends 164
paleoecology 164
sources 163
stratigraphic relations 164
thickness Trends 165
upper boundary 165
interval C. environments 170
formations 168
lithofacies trends 169
paleoecology 172
sources 170
stratigraphic relations 169
thickness trends 169
upper boundary 169
Maudislevem embayment 159, 162, 165
Mulberry fault zone 159, 166
Oachita Mountains, 158, 160, 161, 165, 171, 172, 173
Oachita trough 169
overlying units 173
paleoecology 173
Outarka area 159, 161, 165, 172
Paucarc arch 172
region defined 157
Ti Valley fault 158
total thickness, paleoecology 172
trends 172
underlying units 159
paleoecologics 160
Arkansas Valley, sediment sources, Bannier, J. C., quoted 170
See also Arkansas and northern Louisiana.
Arkoma basin. See Texas Panhandle and Oklahoma.
Arremdisio Group 255
Arrce Formation 235
Ashland Mica Schist, graphite 56
Atoke Age. See Interval B.
Atoke Formation 164, 184
coal 166
conglomerates 189
 fauna 167
paleoecology 171
sandstone, Scull, B. J., quoted 168
Atoke Series 165
Atrasado Member, lower 238
upper 297
Auburn Shale Formation 109, 123, 181
Aulacocrea campbelli 64

Page

Arizona Subgroup 82, 104
Arbuckle Group 129, 248
Arbuckle Group 110, 111
Arbuckle uplift. See Texas Panhandle and Oklahoma.
Armore basin. See Texas Panhandle and Oklahoma.
Argentine Limestone Member 121
Arkansas, abstract 295
Arkansas Valley 302, 303, 305, 307
Black Mesa area 302, 303, 305, 307
Black River 300
Bute fault 297
Chiricahua Mountains 296, 303, 304, 306
Cooglide dam 500
Conchlerian geosyncline 295, 299, 301, 302, 306
Defiance positive element 295, 302, 305, 306
Don Cabezas Mountains 306
East Kaibab monocline 296, 297
Escalante cliff 501
Fort Apache area 501
Fossil Creek 503, 507
Four Corners area 298, 300, 301, 302
Globe 507
Grand Canyon 296, 302, 303, 305, 307, 308
Bright Angel trail 297
Hiddend Canyon 299, 301
Quatermayer Canyon 298, 299
Tanner trail 297, 300
Twin Springs Canyon 298
Guano Hills 302, 304, 955
Holy Joe Peak 500
interval A. environments 299
formations 297
lithofacies trends 299
paleoecology 300
sources 299
stratigraphic relations 297
thickness trends 299
upper boundary 299
interval B. environments 301
formations 300
lithofacies trends 301
paleoecology 301
sources 301
stratigraphic relations 300
thickness trends 300
upper boundary 300
interval C. environments 302
formations 302
lithofacies trends 302
paleoecology 302
sources 302
stratigraphic relations 302
thickness trends 302
upper boundary 301
interval D. environments 304
formations 304
lithofacies trends 304
paleoecology 304
sources 304
stratigraphic relations 304
thickness trends 304
upper boundary 304
interval E. environments 306
formations 306
lithofacies trends 306
paleoecology 306
sources 306
stratigraphic relations 306
thickness trends 305
upper boundary 305
karst topography 295
Little Dragon Mountains 296
Mogollon Rim 301, 303, 304, 306
overlying units 307
paleoecology 308
Patrick basin 297, 300, 302, 304, 305
Pedregosa Mountains 300, 306
Pennsylvanian units, undefined 297

Page

Aulina, Mississippian 297
Avis Sandstone Member 224
Avoa Limestone Member 123
B
Babb Rock formation 75
Baldwin coal 161, 165, 166
Baldwin coal group 25
Bandera Shale formation 105, 151
Bandera Quarzite Member 105
Bar B Formation, upper 239
Barnett Shale 290
Barren Measure, Lower 46
Upper 50
Barren measures, defined 49
Barnsdall formation 189
Basin of eastern Colorado. See Eastern Colorado.
Battery Rock coal 75
Battery Rock Sandstone Member 77
Battle Creek coal 32, 34
Battle Formation 315, 317
Battle Mountain alluvial fan. See Nevada and southern California.
Bear Gulch Limestone Tongue 283
Beverhead Mountains. See Montana and Dakota region.
Bexley coal 32, 34
Bedford coal 39
Bee Rock Sandstone Member 25, 32
Berdeena, defined 42
(Desa Mazon) 119
Berman Formation, lower 238
Bell Limestone Member 123
Belden Shale 259
Bell lower 256
Bell coal 80
Bell sand 117
Belle City formation 189
Bellingham Conglomerate 12
Bend arch. See central and west Texas.
Bend flexure 199
Bentontie, Aka Formation 166
Berino Member 207
Bern Limestone formation 109, 123, 151
Bernadote Sandstone Member 80
Bethany Falls Limestone Member 121, 145
Big Mounds 145
Big Llalh Limestone 247
Bievier Formation 104
Big Belt Mountains. See Montana and Dakota region.
Big Lime 187
Big Saline formation 201, 206
Big Snowy Mountains. See Montana and Dakota region.
Big Springs Limestone Member 123
Big Thompson Canyon, Colo 250
Bighorn basin. See Rocky Mountain region.
Bighorn Mountains. See Montana and Dakota region.
Bioherm, algal 272
Wahuansee Group 152
Bird Sprang-Ev Basin. See Nevada and southern California.
Bird Spring Formation, interval A 314
interval B 317
interval C 319
interval D 321
interval E 322
Bishop Cap Member 221
Black Creek coal 54
Black Creek coal group 30
Black Hills. See Montana and Dakota region.
Black Hills uplift. See Nebraska region.
Black Knob Ridge area 189
Black Mesa area. See Arizona.
Black Prince Limestone 296, 297
INDEX

<table>
<thead>
<tr>
<th>Page</th>
<th>Index</th>
</tr>
</thead>
<tbody>
<tr>
<td>Doniphan Shale Member</td>
<td>123</td>
</tr>
<tr>
<td>Dorchester Slake Member</td>
<td>12</td>
</tr>
<tr>
<td>Dormick Hills Formation, interval A</td>
<td>180, 183</td>
</tr>
<tr>
<td>Dormick Hills Formation, interval B</td>
<td>184, 207</td>
</tr>
<tr>
<td>Dormick Hills Formation, interval C</td>
<td>211</td>
</tr>
<tr>
<td>Dos Cabezas Mountains. See Arizona.</td>
<td></td>
</tr>
<tr>
<td>Douglas Group</td>
<td>265</td>
</tr>
<tr>
<td>Eagle basin, evaporites</td>
<td>272</td>
</tr>
<tr>
<td>Eagle coal</td>
<td>39</td>
</tr>
<tr>
<td>Eagle Shale Member</td>
<td>89</td>
</tr>
<tr>
<td>Eagle Valley Evaporite</td>
<td>271</td>
</tr>
<tr>
<td>Earp Formation, division</td>
<td>305</td>
</tr>
<tr>
<td>interval E</td>
<td>240, 204, 307</td>
</tr>
<tr>
<td>Dye Member</td>
<td>161, 164</td>
</tr>
<tr>
<td>Duyfken monoclinal. See Arizona basin.</td>
<td></td>
</tr>
<tr>
<td>Eastern Mountain Shale Formation</td>
<td>211</td>
</tr>
<tr>
<td>Eastern Colorado, abstract</td>
<td>245</td>
</tr>
<tr>
<td>Enamado basin, Hogoton embayment</td>
<td>247, 254, 259</td>
</tr>
<tr>
<td>Apishapa-Sierra Grande uplift</td>
<td>245, 247, 250, 252</td>
</tr>
<tr>
<td>Apishapa uplift</td>
<td>254, 256, 259</td>
</tr>
<tr>
<td>Basin of eastern Colorado</td>
<td>247</td>
</tr>
<tr>
<td>Canons City embayment</td>
<td>245, 251, 254</td>
</tr>
<tr>
<td>channels, interval A</td>
<td>256, 258, 260, 261, 263</td>
</tr>
<tr>
<td>Fereenout Creek fault</td>
<td>247, 251, 254</td>
</tr>
<tr>
<td>Frontrange uplift</td>
<td>249, 250, 251, 253</td>
</tr>
<tr>
<td>glauconite, interval A</td>
<td>250</td>
</tr>
<tr>
<td>interval A, environments</td>
<td>251</td>
</tr>
<tr>
<td>formations</td>
<td>249</td>
</tr>
<tr>
<td>lithofacies trends</td>
<td>250</td>
</tr>
<tr>
<td>paleotectonic sources</td>
<td>251</td>
</tr>
<tr>
<td>thickness trends</td>
<td>250</td>
</tr>
<tr>
<td>upper boundary</td>
<td>249</td>
</tr>
<tr>
<td>interval B, environments</td>
<td>253</td>
</tr>
<tr>
<td>formations</td>
<td>253</td>
</tr>
<tr>
<td>lithofacies trends</td>
<td>253</td>
</tr>
<tr>
<td>paleotectonic sources</td>
<td>253</td>
</tr>
<tr>
<td>thickness trends</td>
<td>253</td>
</tr>
<tr>
<td>interval C, environments</td>
<td>254</td>
</tr>
<tr>
<td>formations</td>
<td>254</td>
</tr>
<tr>
<td>lithofacies trends</td>
<td>254</td>
</tr>
<tr>
<td>paleotectonic sources</td>
<td>254</td>
</tr>
<tr>
<td>thickness trends</td>
<td>254</td>
</tr>
<tr>
<td>interval D, environments</td>
<td>258</td>
</tr>
<tr>
<td>formations</td>
<td>258</td>
</tr>
<tr>
<td>lithofacies trends</td>
<td>258</td>
</tr>
<tr>
<td>paleotectonic sources</td>
<td>258</td>
</tr>
<tr>
<td>thickness trends</td>
<td>258</td>
</tr>
<tr>
<td>interval E, environments</td>
<td>259</td>
</tr>
<tr>
<td>formations</td>
<td>259</td>
</tr>
<tr>
<td>lithofacies trends</td>
<td>259</td>
</tr>
</tbody>
</table>

Eastern Colorado—Continued

<table>
<thead>
<tr>
<th>Page</th>
<th>Index</th>
</tr>
</thead>
<tbody>
<tr>
<td>interval E—Continued</td>
<td>261</td>
</tr>
<tr>
<td>paleotectonic sources</td>
<td>261</td>
</tr>
<tr>
<td>thickness trends</td>
<td>260</td>
</tr>
<tr>
<td>upper boundary</td>
<td>259</td>
</tr>
<tr>
<td>Keys dome</td>
<td>250</td>
</tr>
<tr>
<td>overlying units</td>
<td>262</td>
</tr>
<tr>
<td>paleotectonic trends</td>
<td>263</td>
</tr>
<tr>
<td>Red Creek arch</td>
<td>248, 249, 262</td>
</tr>
<tr>
<td>references</td>
<td>265</td>
</tr>
<tr>
<td>region defined</td>
<td>245</td>
</tr>
<tr>
<td>sandstone dikes</td>
<td>147</td>
</tr>
<tr>
<td>structural features</td>
<td>247</td>
</tr>
<tr>
<td>total thickness, paleotectonic trends</td>
<td>262</td>
</tr>
<tr>
<td>underlying units</td>
<td>247</td>
</tr>
<tr>
<td>lower boundary</td>
<td>248</td>
</tr>
<tr>
<td>interval E—Continued</td>
<td>249</td>
</tr>
<tr>
<td>Eastern Interior coal basin. See Illinois basin region.</td>
<td></td>
</tr>
<tr>
<td>Effingham Lime Member</td>
<td>321</td>
</tr>
<tr>
<td>Egan Range, unnamed sequence, interval D</td>
<td>321</td>
</tr>
<tr>
<td>Elephant Butte Formation</td>
<td>263</td>
</tr>
<tr>
<td>Elk horn coal</td>
<td>104</td>
</tr>
<tr>
<td>Edlin Limestone Member</td>
<td>123</td>
</tr>
<tr>
<td>Elk Cove</td>
<td>263</td>
</tr>
<tr>
<td>Elmo Limestone Member</td>
<td>123</td>
</tr>
<tr>
<td>Elly basin. See Nevada and southern California.</td>
<td></td>
</tr>
<tr>
<td>Ely Lime Member, interval A</td>
<td>266, 211</td>
</tr>
<tr>
<td>interval B</td>
<td>269, 217</td>
</tr>
<tr>
<td>interval C</td>
<td>271</td>
</tr>
<tr>
<td>interval E</td>
<td>275</td>
</tr>
<tr>
<td>Emery River fault zone. See Appalachian area, central</td>
<td></td>
</tr>
<tr>
<td>Emery Member</td>
<td>109, 123, 131</td>
</tr>
<tr>
<td>Engleval Member</td>
<td>105</td>
</tr>
<tr>
<td>Envine Creek Lime Member</td>
<td>123</td>
</tr>
<tr>
<td>Eosactella, interval A</td>
<td>330</td>
</tr>
<tr>
<td>Esowaringella ultimata</td>
<td>257</td>
</tr>
<tr>
<td>Erin Shale</td>
<td>25, 27, 31</td>
</tr>
<tr>
<td>Esplanade cliff. See Arizona.</td>
<td></td>
</tr>
<tr>
<td>Esplanade Sandstone Member</td>
<td>307</td>
</tr>
<tr>
<td>Esranita trough. See New Mexico.</td>
<td></td>
</tr>
<tr>
<td>Euchet Limestone, interval D</td>
<td>321</td>
</tr>
<tr>
<td>interval E</td>
<td>275</td>
</tr>
<tr>
<td>Fadora Shale Member</td>
<td>123</td>
</tr>
<tr>
<td>Evansville valley</td>
<td>75</td>
</tr>
<tr>
<td>See also Illinois basin region, La Salle anticline</td>
<td></td>
</tr>
<tr>
<td>Evaporites, interval B, Nebraska region</td>
<td>119</td>
</tr>
<tr>
<td>interval C, Eagle basin</td>
<td>272</td>
</tr>
<tr>
<td>Nebraska basin</td>
<td>120</td>
</tr>
<tr>
<td>Paradox basin</td>
<td>271</td>
</tr>
<tr>
<td>interval D, Nebraska region</td>
<td>122</td>
</tr>
<tr>
<td>interval E, central and west Texas</td>
<td>227</td>
</tr>
<tr>
<td>Excelsior Formation</td>
<td>104</td>
</tr>
<tr>
<td>Elxine Limestone Member</td>
<td>107, 109, 123, 131</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Page</th>
<th>Index</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fairbank Member</td>
<td>117</td>
</tr>
<tr>
<td>Fairfield basin. See Illinois basin region.</td>
<td></td>
</tr>
<tr>
<td>Fairley Lime Member</td>
<td>121</td>
</tr>
<tr>
<td>Faulting, intraformational</td>
<td>164</td>
</tr>
<tr>
<td>Fauna. See Fossil.</td>
<td></td>
</tr>
<tr>
<td>Fayetteville Shale</td>
<td>161</td>
</tr>
<tr>
<td>Ferdinand Lime Member</td>
<td>80</td>
</tr>
<tr>
<td>Finnie Sandstone Member</td>
<td>75, 79</td>
</tr>
<tr>
<td>Fire clay. See Clay, refractory</td>
<td></td>
</tr>
<tr>
<td>Fire Clay section</td>
<td>39, 40</td>
</tr>
<tr>
<td>Fire Clay coal</td>
<td>32, 34</td>
</tr>
<tr>
<td>Five Point Lime Member</td>
<td>145</td>
</tr>
<tr>
<td>Flechado Formation</td>
<td>254</td>
</tr>
<tr>
<td>Fleming Formation</td>
<td>104</td>
</tr>
<tr>
<td>Flint, black, Flint Ridge</td>
<td>45</td>
</tr>
<tr>
<td>Kanawha Black</td>
<td>45</td>
</tr>
<tr>
<td>Kilgore</td>
<td>45</td>
</tr>
<tr>
<td>Zaleski</td>
<td>45</td>
</tr>
<tr>
<td>Flint Hill Member</td>
<td>105</td>
</tr>
</tbody>
</table>
Il':D · X

33
Pag
45
30, 32

Flim Rid~ Flint. ................................................ .
F il , Anermrtes ~pp ., interval ...................... ..
64
Aul4co1Mca campbellr .................................. .
50
Calliptens conferl(J ......................................... .
50
Daruurtes, interva l E...................................... ..
169
Densosporrtes , mten I ................................ .
Drplo thmema obrusrloba .............................. .
169
Lanngatosporrtes obscurus ................. .. .......... .
169
thiessenrr .. .. ............................. .. ............... .
L~prdodmron, rnter. I
............................... .
56
31
~p .. interval A ......................................... ..
50
Lescuroptens, imerva l E ................................ .
pp. , rmerv I 0 ....................................... .
47
Manopterrs pottsvrllea .................. .................. . 30, 32
pygmaea ........ ......................................... .. 25. 32
Megalopten s ~pp . , interval B ........................ ..
36
europteris flexuosa ...................................... .. 42. 47
rarrnervrs .................................................. . 42.
sagrnawensrs ......... ................................ ... .
64
chlehanr ....................... ... ........................ .
64
tennesseeana ............................................. . 25. 32
tenurfolro ... ........ ..................................... ..
!16
47
Pecoptens, int rva l 0 .................................... ..
42
spp., interval ....................................... ..
Pun ctatrspontes obliquus ................... ............ .
169
hernopteru artemuraefoliordes ..... ................ .
64
phenoph yllum emargrnatum ....................... .
saxrfragaefolrum .. .................................. ..
trgmana, interval B ..................................... ..
tr trunk~ ....................................................... .
40
Florida Isla nd uplift. ee • ew M xi o.
Flut ca LS , e da and south •m C'.ali fomia ........ .
322
Fontana • hale Forma ti o n ..................................... ..
121
Fore t Cir ba in . ee Ka no;as.
u also Missouri a nd Iowa .
u also Nebra ka region .
Fon Apa h area . u Ari zona .
Fon Chadbourn · fault~ rem. u Centra l a nd 1ve>r
Texas.
112
f on Dodge G ypsum ...............................................
Fo n . 011 Limc•sro ne Forma ti o n .................... l31, 139. I 7
Fort . 0 11 Subgroup ....................................... .. 2. 104 . 105
Fo n Wo nh basin . u C'£nrral a nd wesr T exas.
Fo~sil
reek. ee Ariwna.
Foss ih . Au/rna, Mi sis ippia n ........... .. ........... ... .... .
29?
Beedeina , (Des Moine ).. .. .... .. ......................... .
119
Bo 1 n basin .... ........ ........ .. ................ ... .... ......
II
Chaetetes, imerval B........................... .. .... .... ...
316
imerva l ............................................ 83. 139. 254
mrlleporaceus ............................................
11 7
o t losau r bone .......................................... .. ..
275
Eostaflella, imcrva I A ......................................
3 0
owaerrngdla ult rmata ............ ........ ............. ..
257
Fusulrna , interva l B..................................... ... .
36
int rval ....................................... 42. 2.
leer................................................... ......... .
nouamexr cana ................... ........................
I 7
Fusulrnella , imcrval ..................................... 65. 234
25 1. 299. 317
inrerva l C ..................................... ..... .. ......
186
rowms rs ............. .... ....................... .. .. 36. 66. 7 . 79
stoutr .......... .. .............. ........................ 36. 79
~p .. in rerva l B........................................... 66. 79
giga nr o produ r rid brachi op<Xi .. ........................
329
Kan san~l/a, inrcr I 0 ......................... 46, 86. 107 , 189
I ~ wrcnce Form:uio n .......................................
110
J.rrrgula , inwrval ..........................................
2
int erva l ............................ ......................
52
Lmo productus, inrcrva l B ............................ ..
m;u11w , AI ka f o rma uon , ll cndricks. . A.,
and Pa rk, , B. C. . quored ........... .
Me o lo bus, lnl<'rvll l ..................................... ..
meso lob us ..... ...................................... ... ..
11 7
Mrller~lla aduma ................................. .... ........
285
inrcrv;tl ........................... 130, 180. 234. 248. 299
marblensr .................................................
285
M , linid pelt' poet ........................................
Ill
Nan. n~riln 'K' II b;u in .........................................
II

p, ge
1-o II - C:o nunued
330
Owwarntlla .................................................. ·
167
Para ltgouras (Drabolou ras) vancostatum .... .
Paramrl/erella sp .. imerval ........................ .. &1 .
275
pel cosaur ne ............................................ ..
Prrsmopora, rnre
.................................. ..
Profusu/r nd/4 , adva nced , rnterval
inten I B ................. 36. 79. 99. I 4. 25 1.29 . 316
P eudofusulrna , inr n al E ..............................
145
P eudoschwagerrna , rnten I £.. ...................... 14 5, 22?
Pseudostaffella, nHen I A...............................
330
Rhrprdomella nevadensu ............. .......... .........
~ 15
chr w phorra oltl4homae ................. ................
2~5
ch wagenna, inrerval E.. ............................... ..
275
S prro rbu, inrenoal E ......... .. ............. .........
52
taurod rmidae , int n oa l ..............................
169
330
Tetrataxu ............... ....... ... ............................. ...
257
Trrtrcrtes burgessae ........................................ ..
305
cu llomms rs .............. .. .. .... ...... ............. .
319
interval ................................................. .
interval 0 ....... .................................... . 46, 23 . 32 1
inrerval E............................. . 191 . 238. 321. 323
rrregularis ........... .................... ........ ..........
257
Wede!tmdellma , int n • I C...................... 119. I 6. 254
36
matura .................. ...................... .. ............ .
\ orcc rer localir .............................. ......... ..... II , I ~
Fo untai n Formarion ................ .. ............................ 245, 24
a nh ydrite ..........................................................
253
in rerval ........ ......................... ............. ........ ...
249
inrerva l 8 ................ ... ................... ................. .. 25 1. 2 9
inrrrval C ..... .. ............................. .... ............ .. .. 254. 271
int erva l 0 ... ....................... .. ..... ................ .. .... 256, 273
interva l E ........... ......... .. ......... ...... .......... .... ...... 259. 275
low r. de cribed ................................ .... ...........
252
and tone dik ................................ ..... ...........
260
upper, d scribed ......... ........................ ......... .. ...
252
Four C',omrr area. u Ari zona.
Franci oal ...... .. ......... .... ....... ....... ................. .........
35
Francis Cr ek ha l M mber. .................................
85
189
Franci Forma tion ................................................. .
Franklin M untain . u C'.enrral and we 1 Texa .
I
Fra nk Conglom ra re ............ .................... ..
42
Fr rpon coa l ...... ............................... ...... .. ............. .
44
Freepon a ndsrone Member ........................... .. .... .
Freezeour Cr ek fault . ee Ea rem C',olorado.
Fremonr Lim esrone.......... ....... ... ... .. .. ...... .......... .. ... .
248
French Creek hale Member .... .. ................... .. .. ..... .
123
Fren h Li k. Indi ana. u Ill inois basin regio n.
Fren. lr , Lime rone Member..... ..... ........ ...... .. .... .. ..
I 4
Fre nal Croup ................. ...... ............................
235
Friedri ch ha le Member ...... .................... ............ ...
123
Fri bic Limwon Member .. ....... .. ..........................
121
Fr m ra ng . defined..................... .. ......... ...........
245
Fronrrangc uplih. u
rem Colorado.
ee also . ebra ka r gion.
ee also Rock Mou ntai n region.
Fulda Lime tone Member.......... ...... ...... ... .. ........... .
0
Fusulma , interva l B......... ... .. ...... ......... .. .................
~6
inrerva l ................. 42, 82 . 83. 186. 212, 237. 254. 31
leer .................................. .. .. .. .......... ...... .. ..... .. ... 36. 79
nouamexicana ................................ .. .. ... ...........
187
u also Beedeina .
Fusulinella, imen '3 1 A ............................................ 65. 234
interval 8 .... ....................... . 99. 119. I 4. 25 1. 299.317
I
inrerva l . ...................................... ..... ....... .. ....
10Wt:11S I .................................................... 36. 66. i , 79
SIOUII ................ . ............... .. .......................
36, 79
prol rfrca................. ........... .. .............................
I 6
~ p .. int en 'll l B....... ........ ...... .........
66. 79
G

alesburg ha le Formarion ........................... 10 . 121.
apra nk Formatio n ....... .. .......... .. ......................... .. 211 ,
Permia n.. ..... .............. ................... .. ... .............. .
Garcia Forma ti o n ...... .. ............................ .......... .. .. .
Ca rlock fau lt. u evada and uthem California.
Garlock Form ario n . inr erval
inten•a l B .................. .... ................. .... ........... .. .
inren •al C ..................... .................. .. ...... ..... ... .
inrerva l 0 .............................................. .......... .

13 1
214
225
235
315
3 17
319
321

Page
Carner Formarion .. .................................................
211
nglo mera te ................. .................................
213
.a · pon L ime ro ne ...... ..... .......... .. ....................... 46, 49, 50
Centr ' coal.. ............................................................
75
,corge Creek coa l fi eld . u Appa lac hia n a rl'a ,
cemral.
Geo rge . ree k )'n line. ee ppa l, ch ian a re:t,
r nr ral.
.eorgia . u Appa la hi a n a rea, urh m .
ila Member ....................................................
91
Cinard Formatio n ..... .. ........................................ 25 . 29. 32
G ladevi ll e Forn1a rio n ............................................ ..
Cl:l Mou ntai n . u
ntral a nd w t Tr a .
Clauconire . int cn•a l A ...................... ........ ...... I I. 205 . 250
la u oniri lime tone .. ...... .. ...... ........ ..
204
Cla u o niri sa nd 1 nc .......................................... ..
134
Gl en E rie Forma rion ........................................... ..
24
Glen E ric ha le Member ......... .. .... ...................... ..
249
coa l ................ ..... ......... ... ................................. .
251
Glen haw Fo rma tion ....................................... ... .. .. 46, so
Cobbler Fonna rion ............................... .. ............. ..
234
lo wer .............................................................. ..
236
upper ........... ... ............... .... .............. ... .. .... ....... .
237
Coen Lime ron e Member ...................................... .
211
.olconda rhnr 1. ee 'evada a nd urher n ( ::t li fornia .
.o lden Carr Rang . unn a med cqucn t , inre rva l D 32 1
Golf Cour Fo rma tion .. .... ... ..... .... .......... ....... .......
1
.. ........ ........................... .. . 9, 109

224

c
n on Formario n ................................... .
raford Forma rion .......... .... ......... ......................... .
Graford Group .............. .. ..................... ................. .
Graham F rma tion ............ ... .... ................. ... .... ..... .
chann el ...... ...... ........ .... .... .............................. .
lcs .. ...... .................................... ................. .

314
217
217
224
225
225
67.

79,

Greene Forma tion ............... .. .................. .... ...........
Greenhom fauh . ee Mo ntana a nd Da kota region .
Greenland a nd ton e M mber.. .. .... ............. ...... .....

123
235
51
161

90
75, 79
211
a nd no nhcm

.un ight Lime: 1 ne M m ber.................................
•psum , Forr Dodge ...... .. ......................................
inr erva l B ... .... ............... ..... .............. ........... .. ...
inrerval ......................................
inrerv<j l 0 ................ .....................

224
Ill
67
69
2 9

H

Hale Forrnarion ...................................... 160. 179, I 0. I 3
Hale • B. R .. quored , coa l. Atoka ......... .................
166
Ha ll Can o n Member .... .. .... ... ............ .......... ..........
266
Ha hon Tu ff Lentil ........ ............ ..... .. ... ................. ..
I I
Ha nce Forma rio n ................ .... ..... ........ ....... ... .........
35
Han onburg C roup ......................... ........
235
H appy Holl ow Lime tone Member ...... ........ .........
123
Hardeman ba in . ee Central and west Tcx.a .
See also Texas Panhandl e a nd Okla hom a.
Harding a nd rone .. ... ............................ ...
24
H a rd~ cr.a bbl e Lime rone ........ .................................
247
Ha rlan and tone .. ... ....... .. ................. .. .... .. ............
42
Ha rtford Lime tone Member. .... ... .. .. .......... ...........
123
Ha nridgr hal ................................ .......................
34
101
........ .. ............................... . 104 , 165. I . I
17 1


<table>
<thead>
<tr>
<th>INDEX</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand Mountain. See Appalachian area, central, Walden Ridge.</td>
<td>236</td>
</tr>
<tr>
<td>Sandia Formation.</td>
<td>236</td>
</tr>
<tr>
<td>Sandia Mountains. See New Mexico.</td>
<td>260</td>
</tr>
<tr>
<td>Sandstone dikes, Ingleside Formation.</td>
<td>200</td>
</tr>
<tr>
<td>Sanger de Cristo Formation, interval D.</td>
<td>238</td>
</tr>
<tr>
<td>Sanger de Cristo Mountains. See New Mexico.</td>
<td>285, 275</td>
</tr>
<tr>
<td>Santo Limestone Member.</td>
<td>211</td>
</tr>
<tr>
<td>Sapelo Limestone Formation.</td>
<td>121</td>
</tr>
<tr>
<td>Savannah Formation, 168, 170, 187.</td>
<td>275</td>
</tr>
<tr>
<td>Sashalt Range. See Rocky Mountain region.</td>
<td>235</td>
</tr>
<tr>
<td>Schioptera oklahomana.</td>
<td>314</td>
</tr>
<tr>
<td>Schooner Formation.</td>
<td>228</td>
</tr>
<tr>
<td>Schuller Member.</td>
<td>25, 32</td>
</tr>
<tr>
<td>Schaum member.</td>
<td>275</td>
</tr>
<tr>
<td>Sellers Limestone Member.</td>
<td>54, 76</td>
</tr>
<tr>
<td>Seminoles formation.</td>
<td>131, 156, 187, 189</td>
</tr>
<tr>
<td>Seneca Forin.</td>
<td>186</td>
</tr>
<tr>
<td>Sequatchie Valley anticline. See Appalachian area, central.</td>
<td>275</td>
</tr>
<tr>
<td>Series, Pennsylvanian.</td>
<td>2</td>
</tr>
<tr>
<td>Shaver Limestone.</td>
<td>109, 129, 131</td>
</tr>
<tr>
<td>Sevier, Missouri.</td>
<td>101</td>
</tr>
<tr>
<td>Sevile Limestone Member.</td>
<td>56, 66, 78, 79, 104</td>
</tr>
<tr>
<td>Sewage coal.</td>
<td>25, 32, 34</td>
</tr>
<tr>
<td>Sewell Conglomerate.</td>
<td>77</td>
</tr>
<tr>
<td>Sewell coal.</td>
<td>25, 32, 34</td>
</tr>
<tr>
<td>Sewell Member.</td>
<td>25, 32</td>
</tr>
<tr>
<td>Seats: licks.</td>
<td>50</td>
</tr>
<tr>
<td>Scammon Formation.</td>
<td>105</td>
</tr>
<tr>
<td>Sharon coal.</td>
<td>25, 32</td>
</tr>
<tr>
<td>Sharon Conglomerate Member.</td>
<td>25</td>
</tr>
<tr>
<td>Sharon iron ore.</td>
<td>34</td>
</tr>
<tr>
<td>Sharon Sandstone.</td>
<td>35</td>
</tr>
<tr>
<td>Sharp Mountain Member.</td>
<td>52, 62</td>
</tr>
<tr>
<td>Shavers Group.</td>
<td>109, 122, 151, 145, 148, 191, 259</td>
</tr>
<tr>
<td>Megacyclothems.</td>
<td>110, 111, 148</td>
</tr>
<tr>
<td>stratigraphy.</td>
<td>149</td>
</tr>
<tr>
<td>Shawntown-Rough Creek fault zone. See Illinois basin region.</td>
<td>276</td>
</tr>
<tr>
<td>Shelburn Formation.</td>
<td>81</td>
</tr>
<tr>
<td>Sheldon Limestone Member.</td>
<td>123</td>
</tr>
<tr>
<td>Sheridan arch. See Montana and Dakota region.</td>
<td>129</td>
</tr>
<tr>
<td>Shool Creek Limestone.</td>
<td>87</td>
</tr>
<tr>
<td>Shouemaker Limestone Member.</td>
<td>123</td>
</tr>
<tr>
<td>Shoshone Basin. See Appalachian area, central.</td>
<td>129</td>
</tr>
<tr>
<td>Short Mountain outlier. See Appalachian area, central.</td>
<td>129</td>
</tr>
<tr>
<td>Shumway Limestone Member.</td>
<td>90</td>
</tr>
<tr>
<td>Siderite, Atoka Formation.</td>
<td>166</td>
</tr>
<tr>
<td>Sidielite limestone, Arkansas.</td>
<td>163</td>
</tr>
<tr>
<td>Sierra Diablo. See Central and west Texas.</td>
<td>129</td>
</tr>
<tr>
<td>Sierra Grande uplift. See New Mexico.</td>
<td>129</td>
</tr>
<tr>
<td>Silver Lake Shale Member.</td>
<td>123</td>
</tr>
<tr>
<td>Sinkholes, Kansas.</td>
<td>129</td>
</tr>
<tr>
<td>Sonoran geosyncline. See Arizona.</td>
<td>129</td>
</tr>
<tr>
<td>South Bend Limestone Member.</td>
<td>121</td>
</tr>
<tr>
<td>South Dakota, northern. See Montana and Dakota region.</td>
<td>122</td>
</tr>
<tr>
<td>Southern. See Nebraska region.</td>
<td>123</td>
</tr>
<tr>
<td>Stanly County. See Montana and Dakota region.</td>
<td>124</td>
</tr>
<tr>
<td>Spanish Limestone Member.</td>
<td>125</td>
</tr>
<tr>
<td>Speci mountain Limestone Member.</td>
<td>224</td>
</tr>
<tr>
<td>Sphenophyllum emarginatum.</td>
<td>68</td>
</tr>
<tr>
<td>Saxifraga foliolosia.</td>
<td>68</td>
</tr>
<tr>
<td>Sphenoporus aramaiswaioloides.</td>
<td>64</td>
</tr>
<tr>
<td>Spotsail, interval F.</td>
<td>92</td>
</tr>
<tr>
<td>Spoon Formation, interval B.</td>
<td>36, 66, 78</td>
</tr>
<tr>
<td>interval C.</td>
<td>81, 187</td>
</tr>
<tr>
<td>Spotted Ridge Formation.</td>
<td>230</td>
</tr>
<tr>
<td>Spotted Ridge Limestone.</td>
<td>531</td>
</tr>
<tr>
<td>Spring Branch Limestone Member.</td>
<td>123</td>
</tr>
<tr>
<td>Spring Hill Limestone Member.</td>
<td>121, 143</td>
</tr>
<tr>
<td>Springer Formation.</td>
<td>530</td>
</tr>
<tr>
<td>Springfield Plateau. See Missouri and Iowa.</td>
<td>179, 180, 183</td>
</tr>
<tr>
<td>Squamitum tillite. See Roxbury conglomerate.</td>
<td>276</td>
</tr>
<tr>
<td>Squire Jim coal bed.</td>
<td>50</td>
</tr>
<tr>
<td>Stalmarke sand.</td>
<td>190</td>
</tr>
<tr>
<td>Stanly County, South Dakota. See Montana and Dakota region.</td>
<td>129</td>
</tr>
<tr>
<td>Stanley Shale Formation.</td>
<td>160, 181, 200</td>
</tr>
<tr>
<td>Stannum Limestone Formation, 107, 121, 151, 164, 144</td>
<td></td>
</tr>
<tr>
<td>Stark Shale Member.</td>
<td>121</td>
</tr>
<tr>
<td>Staunton Formation.</td>
<td>78, 81</td>
</tr>
<tr>
<td>Stauroliths, interval C.</td>
<td>169</td>
</tr>
<tr>
<td>Steussy Member.</td>
<td>211</td>
</tr>
<tr>
<td>Surnigaria, interval B.</td>
<td>80</td>
</tr>
<tr>
<td>Stockton coal.</td>
<td>35</td>
</tr>
<tr>
<td>Stonerock Limestone Member.</td>
<td>187</td>
</tr>
<tr>
<td>Stonehouse Canyon Member.</td>
<td>282</td>
</tr>
<tr>
<td>Stoner Limestone Member.</td>
<td>121</td>
</tr>
<tr>
<td>Story Formation.</td>
<td>275</td>
</tr>
<tr>
<td>Stratoclass sformation.</td>
<td>109, 123, 151</td>
</tr>
<tr>
<td>channel cut.</td>
<td>109</td>
</tr>
<tr>
<td>Stranger Formation.</td>
<td>110, 151</td>
</tr>
<tr>
<td>Strashub Shandon Member.</td>
<td>24</td>
</tr>
<tr>
<td>Strathbroom Formation, interval D.</td>
<td>321</td>
</tr>
<tr>
<td>interval E.</td>
<td>323</td>
</tr>
<tr>
<td>Stratigraphy, interval division.</td>
<td>3</td>
</tr>
<tr>
<td>Strawn Group.</td>
<td>186, 206, 210</td>
</tr>
<tr>
<td>Strawn limestone, lower.</td>
<td>211, 222</td>
</tr>
<tr>
<td>Strawn series.</td>
<td>210</td>
</tr>
<tr>
<td>Stream channels. See Channels.</td>
<td>276</td>
</tr>
<tr>
<td>Structural features.</td>
<td>5</td>
</tr>
<tr>
<td>Appalachian area, central.</td>
<td>19, 23</td>
</tr>
<tr>
<td>Northern.</td>
<td>18, 20</td>
</tr>
<tr>
<td>Southern.</td>
<td>18</td>
</tr>
<tr>
<td>Appalachian region.</td>
<td>19</td>
</tr>
<tr>
<td>Arizona.</td>
<td>296</td>
</tr>
<tr>
<td>Arkansas and northern Louisiana.</td>
<td>158</td>
</tr>
<tr>
<td>Central and west Texas.</td>
<td>199, 200</td>
</tr>
<tr>
<td>eastern Colorado.</td>
<td>247</td>
</tr>
<tr>
<td>Illinois basin region.</td>
<td>74</td>
</tr>
<tr>
<td>Kansas.</td>
<td>128</td>
</tr>
<tr>
<td>Michigan basin region.</td>
<td>64</td>
</tr>
<tr>
<td>Missouri and Iowa.</td>
<td>98</td>
</tr>
<tr>
<td>Montana and Dakota region.</td>
<td>280</td>
</tr>
<tr>
<td>Nebraska region.</td>
<td>116</td>
</tr>
<tr>
<td>Nevada and southern California.</td>
<td>513</td>
</tr>
<tr>
<td>New Mexico.</td>
<td>254</td>
</tr>
<tr>
<td>New England.</td>
<td>10</td>
</tr>
<tr>
<td>Rocky Mountain region.</td>
<td>268</td>
</tr>
<tr>
<td>Texas Panhandle and Oklahoma.</td>
<td>179</td>
</tr>
<tr>
<td>Stuart Shale Formation.</td>
<td>186</td>
</tr>
<tr>
<td>Stull Shale Member.</td>
<td>123</td>
</tr>
<tr>
<td>Subinterval A. See interval A.</td>
<td>297</td>
</tr>
<tr>
<td>Subinterval A. See interval A.</td>
<td>129</td>
</tr>
<tr>
<td>Sulphide ores, Missouri.</td>
<td>102</td>
</tr>
<tr>
<td>Supai Big A.</td>
<td>505</td>
</tr>
<tr>
<td>Supai delta. See Arizona.</td>
<td>505</td>
</tr>
<tr>
<td>Supai Formation.</td>
<td>296</td>
</tr>
<tr>
<td>interval A.</td>
<td>297</td>
</tr>
<tr>
<td>interval C.</td>
<td>501</td>
</tr>
<tr>
<td>interval E.</td>
<td>504</td>
</tr>
</tbody>
</table>
INDEX

Texas Panhandle and Oklahoma—Continued
interval D—Continued
thickened trends
upper boundary
interval E, environments
lithofacies trends
paleoecotones
sources
thickened trends
upper boundary
Keys dome
Marietta basin
Madoc arch
Munsur arch
Nemaha anticline
Osage geosyncline
Osage structural belt
Osage uplift
overlying units
paleoecotones
Osark uplift
Palo Duro basin
Red River arch
references
structural features
total thickness, paleoecotones
trends
unwelding units
lower boundary
paleoecotones
Union County, New Mexico
Wichita uplift
13 finger limestone
13-fingered limestone
Thomson coal
Thrift Formation
Thurber coal bed
Thurman Sandstone Formation
Ti Valley fault, See Arkansas and northern Louisiana.
Tontonose Sandstone Member
Tonkawa sand
Tontonosean facies
Topache Formation, interval A
Topoek Formation
Topoek Limestone Formation
Topography, pre-Pennsylvania, Kansas
Toronto Limestone Member
Trace Creek Shale Member
Traceable S sandstone complexes
Trade Water Formation, interval B
interval C
Transcontinental arch
Tree stumps
Tree trunks, fossil
Trinity Group
Tripollic chert
Triovill Sandstone Member
Trinitites

Val Verde basin. See Central and west Texas.
Vamosa Formation
Vancom Formation
Vanport Limestone Member
Vekol Mountains. See Arizona.
Verdigis Formation
Verdes Group
Verne coal
Verne Limestone Member
Villas Shale Formation
Village Bend Limestone Member
Virgil Age. See Interval E.
Virgil Series
Virginia. See Appalachian area, central.
Volcanics, Blackstone Series
Boston basin
Lyman Complex
Marlboro Formation
Marquette Complex
Narragansett basin
New England

INDEX

Turbidites

Unia Mountains. See Rocky Mountain region.
Uncompaghre trough, evaporites
Uncompaghre uplift. See New Mexico.
See also Rocky Mountain region.
Undersand, Olive Hill Clay Bed
root zones
Sciotoville Clay Member
Underlying units, Anasazi fields area
Appalachian area, central
northern
southern
Appalachian region
Arkansas and northern Louisiana
Arizona
Boston basin
central and west Texas
eastern Colorado
Illinois basin region
Kansas
Michigan basin region
Missouri and Iowa
Montana and Dakota region
Narragansett basin
Nevada region
Nevada and southern California
New Mexico
Pacific Northwest region
Rocky Mountain region
Texas Panhandle and Oklahoma
Union County, New Mexico. See Panhandle and Oklahoma.
Union Valley Sandstone Member
Uniontown coal
Upper Cambrian
Upper Devonian
Upper Paleozoic
Upper unnamed formation, Pleasanton Group
Utah. See Rocky Mountain region.

INDEX

Sapulpa Formation
Supai sand
Supai area. See Pacific Northwest region.
Sweatwater trough. See Rocky Mountain region.
Swepse Limestone Formation
algal mounds
T
Tacket Formation
Talladega Slate
Talladega slate belt. See Appalachian area, southern.
Tallant Formation
Tarkio Lime stone Formation
Tarler coal
Tebo Formation
Tecumseh Shale Formation
Tetley Mountains. See Montana and Dakota region.
Tennessee. See Appalachian area, central.
Tendeply Sandstone, interval C
interval D
sources
Tenas Formation
upper, interval A
Tetra/axs
Tonopah Sandstone Member
Topoek Formation
Topoek Limestone Formation
Topography, pre-Pennsylvania, Kansas
Toronto Limestone Member
Trace Creek Shale Member
Traceable S sandstone complexes
Trade Water Formation, interval B
interval C
Transcontinental arch
Tree stumps
Tree trunks, fossil
Trinity Group
Tripollic chert
Triovill Sandstone Member
Trinitites

Val Verde basin. See Central and west Texas.
Vamosa Formation
Vancom Formation
Vanport Limestone Member
Vekol Mountains. See Arizona.
Verdigis Formation
Verdes Group
Verne coal
Verne Limestone Member
Villas Shale Formation
Village Bend Limestone Member
Virgil Age. See Interval E.
Virgil Series
Virginia. See Appalachian area, central.
Volcanics, Blackstone Series
Boston basin
Lyman Complex
Marlboro Formation
Marquette Complex
Narragansett basin
New England

INDEX

Sapulpa Formation
Supai sand
Supai area. See Pacific Northwest region.
Sweatwater trough. See Rocky Mountain region.
Swepse Limestone Formation
algal mounds
T
Tacket Formation
Talladega Slate
Talladega slate belt. See Appalachian area, southern.
Tallant Formation
Tarkio Lime stone Formation
Tarler coal
Tebo Formation
Tecumseh Shale Formation
Tetley Mountains. See Montana and Dakota region.
Tennessee. See Appalachian area, central.
Tendeply Sandstone, interval C
interval D
sources
Tenas Formation
upper, interval A
Tetra/axs
Tonopah Sandstone Member
Topoek Formation
Topoek Limestone Formation
Topography, pre-Pennsylvania, Kansas
Toronto Limestone Member
Trace Creek Shale Member
Traceable S sandstone complexes
Trade Water Formation, interval B
interval C
Transcontinental arch
Tree stumps
Tree trunks, fossil
Trinity Group
Tripollic chert
Triovill Sandstone Member
Trinitites

Val Verde basin. See Central and west Texas.
Vamosa Formation
Vancom Formation
Vanport Limestone Member
Vekol Mountains. See Arizona.
Verdigis Formation
Verdes Group
Verne coal
Verne Limestone Member
Villas Shale Formation
Village Bend Limestone Member
Virgil Age. See Interval E.
Virgil Series
Virginia. See Appalachian area, central.
Volcanics, Blackstone Series
Boston basin
Lyman Complex
Marlboro Formation
Marquette Complex
Narragansett basin
New England
| INDEX |
|-------------------|----------------------|
| **Wahaunee Group—Continued** | **Page** |
| coal | 152 |
| described | 150 |
| Moore, R. C., quoted | 152 |
| stratigraphy | 151 |
| Washburn coal | 32 |
| Wakarusa Limestone Member | 123 |
| Walden Ridge. See Appalachian area, central. |  
| Waldrip Shale Member | 222 |
| Walter Johnson Member | 105 |
| Wamsutta Formation | 12 |
| Wann Formation | 189 |
| Waywater Limestone | 180, 183 |
| Warnington Limestone Member | 235 |
| Warner Formation | 105 |
| correlation unit | 101 |
| Warrensburg Sandstone Member | 88, 107, 108 |
| Warrior coal field. See Appalachian area, southern. |  
| Warsaw Limestone | 217 |
| Warburg Sandstone | 55 |
| Washington. See Pacific Northwest region. |  
| Washington Formation | 51 |
| Wattsville basin. See Appalachian area, southern. |  
| Wayland Shale Member | 224 |
| Waynesburg coal | 50 |
| Waynesburg Formation | 51 |
| Wea Shale Member | 121 |
| Weber Sandstone, interval D | 273 |
| interval E | 256 |
| Wedekindella, interval C | 119, 186, 254 |
| matura | 36 |
| Weir Formation | 104 |
| Well logs | 3 |
| Wells Formation, interval A | 256 |
| interval B | 269 |
| interval C | 271, 286 |
| interval D | 273 |
| interval E | 275 |
| Wendover Member | 121, 122 |
| **West Franklin Limestone** | **Page** |
| West Virginia. See Appalachian area, central. |  
| Westerly Limestone Member | 121 |
| Weston Formation | 108 |
| Weston Shale Member | 107 |
| West Mountains. See Rocky Mountain region. |  
| Westumma Shale Formation | 186 |
| Watauga County, Washington. See Pacific Northwest region. |  
| Whetstone Mountains. See Arizona. |  
| Whisky Canyon Formation | 235 |
| White Cloud Shale Member | 123 |
| White River. See Arizona. |  
| White River uplift. See Rocky Mountain region. |  
| Whitfield Range, Trail Creek. See Montana and Dakota region, Trail Creek. |  
| Whitt Group | 217 |
| Wichita paleoplain | 228 |
| Wichita uplift. See Texas Panhandle and Oklahoma. |  
| Wildcat Peak Formation | 314, 925 |
| Wiles Limestone Member | 217 |
| Willard Shale Formation | 109, 123, 131 |
| Williams Canyon Limestone | 247 |
| Willis coal | 80 |
| Williston basin. See Montana and Dakota region. |  
| Winchell Limestone Formation | 217 |
| Wind River Range. See Rocky Mountain region. |  
| Winifred Limestone Member | 59 |
| Winslow Formation. See Atoka Formation. |  
| Winnetuk Shale Member | 121 |
| algal mounds | 143 |
| Wise Formation | 35 |
| Wissabickon Schist | 49 |
| Wolf Mountain Shale Formation | 217, 219 |
| Wolfcamp Formation | 228 |
| Wood River basin. See Rocky Mountain region. |  
| Wood River Formation, Amherst orogenic belt | 270 |
| interval A | 266 |
| interval B | 269 |
| interval C | 271, 286 |
| interval D | 275 |
| interval E | 273 |
| Wood Siding Formation | 123, 131, 145, 191 |
| Woodbury Limestone Member | 90 |
| Woody Creek Member | 161, 165, 164 |
| Woosocket basin. See Narragansett basin. |  
| Worcester locality | 11, 15 |
| Wyandotte Limestone Formation | 107, 121, 131 |
| algal mounds | 145 |
| Wynn Limestone Member | 217 |
| Wyoming. See Rocky Mountain region. |  
| Campbell County. See Montana and Dakota region. |  
| Crook County. See Montana and Dakota region. |  
| eastern. See Nebraska region. |  
| Wyoming shelf. See Montana and Dakota region. |  
| See also Rocky Mountain region. |  
| Y, Z |  
| Yakimak Limestone, interval A | 282 |
| Yellow leaf basin. See Appalachian area, southern. |  
| Yellowstone National Park, northern. See Montana and Dakota region. |  
| Yeshick coal | 32, 35 |
| Zaleski Flint | 45 |
| Zanesville, Ohio, escarpment | 45 |
| Zarah Subgroup | 106 |
| Zeandale Limestone Formation | 109, 131, 145 |
| Zuckerman Limestone Member | 190 |
| Zuni positive element. See Arizona, De Soto positive element. |  
| Zuni uplift. See New Mexico. |  

*U.S. GOVERNMENT PRINTING OFFICE: 1975-677-340/20*