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Paleotectonic Investigations of the Mississippian System in the United States

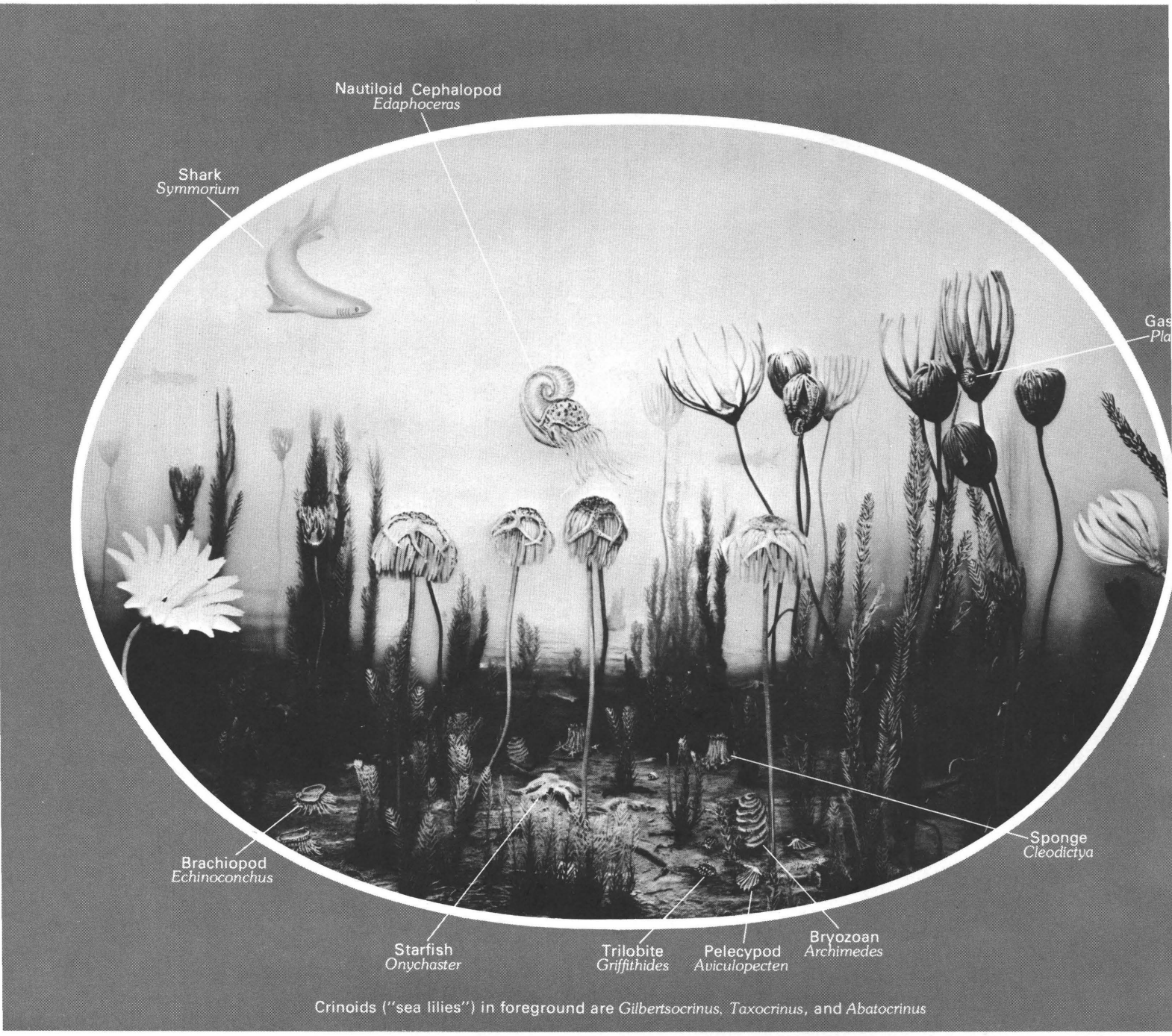
Part I. Introduction and Regional Analyses of the Mississippian System

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010

PART I OF III

PALEOTECTONIC INVESTIGATIONS OF THE MISSISSIPPIAN SYSTEM IN THE UNITED STATES

GENERALIZED RECONSTRUCTION OF SEASCAPE IN NORTH CENTRAL UNITED STATES DURING MIDDLE MISSISSIPPIAN (LATE OSAGE) TIME



Modified from a diorama prepared by the Marchand broth for the National Museum of Natural History, Smithson Institution. Modifications based in large part on Lane (1973). Prepared with the assistance of M. E. Williams Cleveland Museum of Natural History, and several members of the Paleontology and Stratigraphic Branch U.S. Geological Survey.
Artist: Tom Kostick.

Paleotectonic Investigations of the Mississippian System in the United States

Part I. Introduction and Regional Analyses of the Mississippian System

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Chapters A through Q

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PREFACE

This professional paper is the fifth in a series of paleotectonic studies each covering a geologic system in the conterminous United States. Part I provides a region-by-region discussion of data concerning the Mississippian System and an explanation and documentation for the maps and sections contained in part III. Part II of the paper provides a summary of the Mississippian System, presents interregional interpretations permitted by this study, and includes sections on notable features of the system. The maps contained in the separate case as part III may be divided into two groups: (1) a sequence of factual or basic maps that shows, with a minimum of interpretation, the Mississippian System as it occurs today, and (2) interpretive maps that attempt a reasonable reconstruction of the original extent of the system, its tectonics, environment, and geography.

This study of the Mississippian System was made by 23 geologists who were responsible either jointly or individually for the 19 regions into which the United States was divided. These authors are:

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Robert G. Yates

The text and maps were compiled, edited, and assembled by L. C. Craig, C. W. Connor, and K. L. Varnes. Chapters dealing with special features of the Mississippian System were prepared by invited geologists or were coordinated from contributions provided by many of the authors in the above list.

The preceding paleotectonic studies of the geologic system of the conterminous United States are —

Jurassic System

McKee, E. D., and others, 1956, Paleotectonic maps of the Jurassic System: U.S. Geol. Survey Misc. Geol. Inv. Map I–175, 6 p., 9 pls.

Triassic System

McKee, E. D., and others, 1959, Paleotectonic maps of the Triassic System: U.S. Geol. Survey Misc. Geol. Inv. Map I–300, 33 p., 9 pls.

Permian System

McKee, E.D., Oriel, S. S., and others, 1967, Paleotectonic investigations of the Permian System in the United States: U.S. Geol. Survey Prof. Paper 515, 271 p., 6 pls.

McKee, E. D., Oriel, S. S., and others, 1967, Paleotectonic maps of the Permian System: U.S. Geol. Survey Misc. Geol. Inv. Map I–450, 164 p., 20 pls.

Pennsylvanian System

McKee, E. D., Crosby, E. J., and others, 1975, Paleotectonic investigations of the Pennsylvanian System in the United States: U.S. Geol. Survey Prof. Paper 853, 541 p., 39 pls [1976].

---

[1] Nebraska Geological Survey.
[2] Kansas Geological Survey.
[3] Oklahoma Geological Survey.

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[Letters in parentheses designate separate chapters]

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Frontispiece. Generalized reconstruction of seascape in north-central United States during Middle Mississippian time.

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Introduction, Acknowledgments, and Methods

By LAWRENCE C. CRAIG

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–A

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

INTRODUCTION, ACKNOWLEDGMENTS, AND METHODS

By LAWRENCE C. CRAIG

INTRODUCTION

Paleotectonic map compilation was begun by the U.S. Geological Survey in 1952, and a project to compile the Mississippian System was initiated in late 1961. The objective of the program is to prepare publications that document rock thicknesses, generalized lithology, and other regional rock relations for each of the geologic systems on a nationwide basis and to interpret these data in terms of tectonic, environmental, and geographic evolution. The first publications emphasized the data record and included a minimum of interpretation; later publications included more interpretative maps, and in this report the authors of each region have been encouraged to discuss their interpretations in the text as well as to show the interpretations on maps. It is hoped that the factual and interpretive materials together will provide a summary for the reader, as well as a basis for modifications and new concepts as additional data become available.

ACKNOWLEDGMENTS

A compilation of this magnitude must rely on many sources and on the support of many individual geologists. Authors have been dependent on the cooperation of State geological surveys and of university and oil company geologists. In particular, we acknowledge the cooperation of E. C. Reed and V. H. Dreeszen, State Geologists of Nebraska, F. C. Foley and W. W. Hambleton, State Geologists of Kansas, and C. C. Branson and C. J. Mankin, State Geologists of Oklahoma for the author contributions made by members of their staffs: M. P. Carlson of the Nebraska Geological Survey, E. D. Goebel and G. F. Stewart of the Kansas Geological Survey, and Louise Jordan of the Oklahoma Geological Survey. Through the services of the Government Liaison Committee of the American Petroleum Institute, some critical data were obtained that assisted the construction of the maps; these services are gratefully acknowledged.

Special acknowledgment must be extended for assistance provided during the preparation of the frontispiece. Dr. Michael E. Williams, The Cleveland Museum of Natural History, provided original data and guidance for the restoration of the shark, *Symmorium*. Many of the smaller invertebrates were added to the photograph of the Smithsonian diorama based on the published assemblage study by N. Gary Lane (1973). Members of the Paleontology and Stratigraphy Branch, U.S. Geological Survey, were helpful in providing suggestions and guidance during the preparation of the frontispiece.

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CONSTRUCTION OF MISSISSIPPIAN PALEOTECTONIC MAPS[1]

COMPILATION OF DATA

The team of authors for the Mississippian investigations began compilation of data in mid-1962. Data collection was largely completed by the end of 1965 and this is the effective date of most of the data contained in this publication. Until publication, however, some important new data continued to be incorporated in the maps and text during the compilation and editing.

The region of responsibility for each author of this paper is shown in figure 1. The work was directed by Lawrence C. Craig and the maps and cross sections were edited by Carol W. Connor and Katharine L. Varnes. Louise M. Kiteley prepared most of the text figures. Curt A. Mast scribed or drafted all the major plates in this publication.

Sources of data and availability. — Each publication in this series is an effort to synthesize the available information of one geologic system. The sources of information are from the published and unpublished records of the geologic profession and include data provided by geologists in private industry, universities, and government surveys. During compilation the data are summarized on standardized data cards, one or more for each locality. The many thousands of cards accumulated during the paleotectonic studies form a permanent file of stratigraphic data arranged by State. This file, except data obtained in confidence, is available for use at the library of the U.S. Geological Survey, Denver, Colo. The specific sources of data used in the study of the Mississippian System are listed in the "Index to Localities and Sources" (chap. W) and in the "References" lists accompanying each chapter of this report.

Index to localities and sources. — The list of sources (part II, chap. W) provides documentation for this study in an abbreviated form and is keyed to the locality maps (pl. 1) by State and locality number, each State having its own series of numbers. The abbreviations are explained at the beginning of chapter W and entries in the chapter list allow the reader to identify the original source and obtain the data if they are a part of the public record. Complete references both for publications and for unpublished theses are given in the "References" at the end of each chapter.

MAPPING UNITS

In order to analyze the Mississippian System it is necessary to separate it into a number of parts. A lithofacies map and, to a lesser extent, an isopach map of the entire system is difficult to analyze; dividing the system into several mapping units minimizes the combining of geologic conditions, depositional and erosional, that differed from time to time during the geologic period.

The Mississippian System is divided into four parts designated, in ascending order, intervals A, B, C, and D (pl. 15). Most of the maps prepared for this publication are designed to show certain geologic features of an interval. The intervals can be recognized and delimited nearly everywhere where the system occurs in the United States, and, therefore, comparison of genetically related events can be made between regions.

DIVISION OF THE SYSTEM

Use of the informal term "interval" and the means for recognizing interval boundaries are discussed by Oriel (in McKee and others, 1959, p. 5). Essentially, the intervals are rock-stratigraphic units that serve as practical time-stratigraphic units as proposed by Rodgers (1954, p. 658–659). Each interval is composed of an assemblage of members, formations, or groups that lies mainly between recognizable lithologic contacts. The lithologic boundary of an interval in any given area may not and commonly does not represent the same time surface as the corresponding interval boundary elsewhere. In some places age control is lacking and rocks are arbitrarily assigned to intervals on the basis of rock-stratigraphic inferences, or age control may be so sparse that interval boundaries are poorly defined and must be placed arbitrarily at a convenient horizon marking a change in rock type. In some

[1] Inasmuch as this paper is one of a standardized series of publications, much of the material presented here is taken directly from preceding paleotectonic map publications without specific acknowledgment. (See "Preface.")

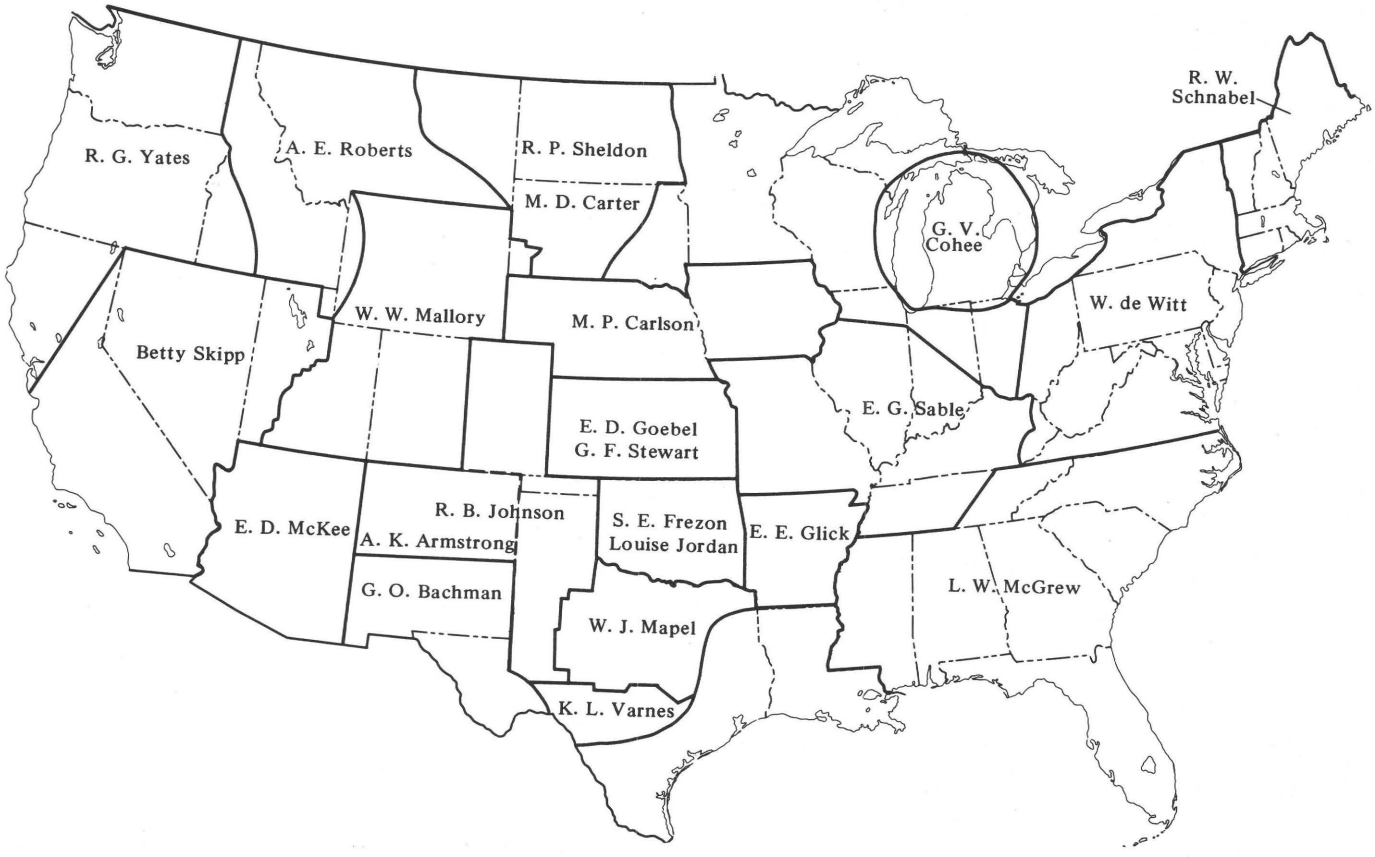

FIGURE 1. — Areas of responsibility.

other areas the effective cutoff date of the work preceded new age information that could not be incoporated into the study. In a few places new or controversial age information has been used in this study that has not yet become a part of official action by the U.S. Geological Survey. For the Mississippian System the four intervals in ascending order correspond approximately to the four provincial series of the Mississippi Valley area: the Kinderhook, Osage, Meramec, and Chester Series.

A two-letter designation, interval A-B, is used in the eastern part of the Appalachian Basin region of this paper. In this area intervals A and B compose a thick sequence of dominantly coarse-grained detrital strata that are poorly known throughout much of the area. The absence of key beds and of well-marked floral and faunal zones precludes subdivision of the mass into discrete meaningful units, and projection of an arbitrary contact was considered misleading. These Lower Mississippian rocks are mapped collectively and are shown conjoined as interval A-B on the isopach and lithofacies maps of both interval A (pls. 3-A, 3-B) and interval B (pls. 4-A, 4-B). This usage is discussed in more detail in chapter C: "Appalachian Basin region."

BASIS FOR CORRELATION

A basis for the assignment of Mississippian rocks to intervals was provided by the Mississippian correlation chart compiled by the Mississippian subcommittee (Weller and others, 1948) of the Committee on Stratigraphy of the National Research Council. At the beginning of the present study of the Mississippian, meetings were held with a number of paleontologists of the U.S. Geological Survey, notably Helen Duncan, J. T. Dutro, Jr., Mackenzie Gordon, Jr., and W. J. Sando, and a preliminary correlation chart was established. In following years, a number of changes were made to the chart as new or different rock or fossil relations were recognized. Particularly helpful in this stage were identifications by B. A. Skipp of endothyrid Foraminifera in some poorly dated sections in Western United States.

Thus, in this paper the assignment of Mississippian stratigraphic units to intervals is the result of both paleontologic and lithologic studies; in some decisions, the placement of units in intervals was quite arbitrary as a result of conflicts of geologic opinion or lack of diagnostic information on age of units. Authors have been at liberty to evaluate, revise source material, and reassign rock units on the basis of new information.

CHART OF STRATIGRAPHIC TERMS

Plate 15 is a generalized chart showing Mississippian rock-stratigraphic units in the conterminous United States and indicates (by heavy line) the assignment of these units to intervals. In many columns on the chart the interval boundaries do not coincide with the provincial series or time-equivalent boundaries (horizontal dashed lines) by which the interval is ideally defined. These columns represent areas in which a time boundary cannot be identified in the rock sequence or cannot be mapped across a broad area, and the nearest practical lithologic change must be used as a mapping convenience. Thus, in such a place, an interval may include beds older or younger than those included in the ideal interval, or the interval may exclude some beds equivalent in age to part of the ideal interval. Gaps in the stratigraphic record, the result of either erosion or nondeposition, are indicated by shading. Little attempt has been made to indicate details of correlation, such as facies relation, overlap, or intertonguing, and the time span represented by each stratigraphic unit is indicated in only a general way.

On plate 15 and throughout this paper, stratigraphic names that have not been adopted by the U.S. Geological Survey and those for which there has been no occasion for official action are shown in italics. Names are italicized also when usage in an area differs from the accepted usage of the U.S. Geological Survey. Drillers' terms and names of rock units that are defined by economic significance are italicized also. Stratigraphic names that have been adopted by the Geological Survey are not italicized. Where names are applied to units that are definitely not or probably not the same as those of the type area, the names are enclosed in quotation marks.

At the time that this project started, the U.S. Geological Survey did not use -ian endings on series names. In 1972 this was changed officially and now the use of -ian series endings is standard practice. This project was so far advanced at that time that it was deemed inadvisable to change each series name where it occurs on plate 15 or in the text.

TYPES OF MAPS AND SECTIONS

As in previous publications in this series, the basic plates accompanying this paper are of two general types, a group that is essentially factual or "objective" and a group that is interpretive. The factual group includes: (1) the outcrop and control point map, (2) the geologic map of units underlying the Mississippian, (3) the isopach maps of each interval, (4) the lithofacies maps of each interval, (5) the total isopach map of the Mississippian System, (6) the geologic map of units directly above the Mississippian System, and (7) the cross sections of the Mississippian System. The interpretive group includes the following maps of each interval: (1) the paleotectonic maps, (2) the sedimentary environment maps, and (3) the paleogeographic maps. In addition to these basic plates, a few special plates illustrate economic or other special features of the Mississippian System and text figures provide index maps or other illustrations to augment descriptive material.

Comparison of the maps of each interval provides an impression of the dynamic changes which influenced the distribution and nature of the sedimentary rocks of the system and either directly or indirectly indicates the tectonic history of the United States during Mississippian time.

PREPARATION OF MAPS AND SECTIONS

Compilation scale. — Most maps of the factual group (pls. 2–8) were compiled on base maps at a scale of 1:2,500,000 and subsequently were reduced to the publication scale of 1:5,000,000. Cross sections (pl. 9-A–9-G) were compiled at 1:2,500,000 horizontal scale and were published at that scale. Several vertical scales were used, depending on the thickness of units: 1 inch =250 feet, 1 inch=500 feet or 1 inch=1,000 feet. The interpetive map group was compiled on several different scales; the paleotectonic maps were compiled at 1:2,500,000 and reduced to 1:10,000,000 for publication, the sedimentary environment maps were compiled at 1:5,000,000 and reduced to 1:10,000,000 for publication, and the paleogeographic maps were compiled at 1:17,000,000 and were published at the same scale.

Map of control points and generalized outcrop pattern of the Mississippian System (scale 1:2,500,000). — The location of points that were used as controls for the Mississippian maps is shown on plate 1. The "Index to Localities and Sources" provides an abbreviated identification of the source of information for each locality. (See chap. W.) Numbers on the map correspond to those in the index and to those in the punchcard file. All previous paleotectonic map publications have used a continuing series of locality numbers; so that, for example, Texas locality 22 in the Jurassic paleotectonic map publication (McKee and others, 1956) was the same locality and source as Texas locality 22 in the Permian paleotectonic map publication (McKee, Oriel, and others, 1967). This system was abandoned in preparing the Mississippian maps, and locality numbers used here have no relation to numbers used previously.

Purposes of the map of control points and the "Index to Localities and Sources" are to enable the reader to (1) evaluate the relative significance and reliability of the maps and sections, (2) compare data in this paper with data from other sources, and (3) have an oppor-

tunity to prepare other, perhaps more detailed, maps of local areas using supplementary data.

The density of control points on the maps indicates in some degree the relative amount of information available and permits an evaluation of the reliability of isopachs and lithofacies trends. In some areas of dense control more data were available, but all control points could not be shown on the maps at a scale of 1:2,500,000. In addition, some control points represent surface sections or wells that have been described more than once; some of these alternative descriptions are in agreement, some are not. Most of the additional or alternative data in these areas were studied and compared with those selected for map representation, but where there was disagreement only one interpretation could be shown on the principal maps. Some of the other interpretations are shown on special illustrations or are discussed in the text.

Most of the outcrops of the Mississippian System shown on plate 1 have been adapted from large-scale geologic maps, such as State or County geologic maps. The outcrop patterns are necessarily generalized at the scale of publication. For example, single line widths may exaggerate the representation of very narrow belts of outcrop; similarly, in order to show some isolated erosion remnants on the map, their size has been exaggerated, whereas other, generally smaller, erosion remnants have been omitted from the map.

Geologic maps of units directly beneath and directly above the Mississippian System (pls. 2, 8). — These maps provide bracketing information for the Mississippian System. They show only the systemic breakdown of the underlying or overlying rocks, except for the Devonian on the map of the underlying rocks and the Pennsylvania on the map of overlying rocks; these two systems are divided into series. Letter symbols indicate names of selected groups or formations in the general area of their usage but no attempt has been made to draw boundary lines between them.

Thickness maps (pls. 3-A–6-A, 7). — The isopach maps for each interval and the total isopach map of the system are factual maps that indicate thicknesses preserved today. Thus, care must be exercised in interpreting thickness values shown on these maps because they do not necessarily represent original thicknesses; erosion may have removed part of the originally deposited beds of the mapped unit. In several regions of the midcontinent where rocks of the Mississippian System are flat lying and at the surface over broad areas, a convention of placing a line above each posted thickness value is used to indicate that the top of the mapped unit was eroded during Quaternary time. Thicknesses shown at some places are incomplete because basal parts of the sections are not exposed or are not penetrated by drill.

Only a selected, limited number of control points have the thickness values printed by them; all control points shown on each map, however, were used as control in constructing isopachous lines. Although these maps are called "factual," the information provided may be interpretive in some degree; geologic bias, for instance, may be introduced in the assignment of beds to the mapped unit and in places the control may permit alternative construction of isopachs.

In much of the Basin and Range province of Nevada, southern California, southern Arizona, and southern New Mexico, 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 valleys. In only a few places, however, have wells penetrated Mississippian rocks in these areas; thus, the construction of isopachs across the valleys must be regarded as an interpretive reconstruction for the sake of providing as complete a treatment of the conterminous United States as possible.

In the Pacific Coast region, points are so widely spaced that interpolation between them is not warranted. Here information is so sparse and indefinite that only a few tentative thickness values are shown on the total system isopach map; a few symbols indicating presence of rocks are shown on interval isopach maps, and at only a few localities are lithofacies colors shown on the lithofacies maps.

In the course of compiling thickness and lithofacies maps it was desirable to show some faults in order to explain map discontinuities; for example, abrupt offsets in isopachs or marked changes of lithofacies patterns. In areas of known Mississippian or post-Mississippian faulting, such as the Appalachian Mountains, where faults did not produce an abrupt change or where sparse control made delineation of structures uncertain, the faults have been omitted. Selected faults that are known to have moved in Mississippian time are shown in black; those that moved only after the Mississippian are shown in red.

Lithofacies maps (pls. 3-B–6-B). — The lithofacies maps for each interval indicate the distribution of average rock types of the mapped unit as these units are preserved today. Care must be exercised in interpreting the rock types shown on these maps because they do not necessarily represent all of the originally deposited unit; erosion may have removed part of the original beds and the representation on the map of the remaining beds in places may be an incomplete and improper symbolization of the original unit. In places where sections are incomplete because the base of the

unit is not exposed or was not penetrated by drilling, proportions of rock types cannot be computed. In some areas they can be estimated with a reasonable degree of reliability.

As in isopach mapping, lithofacies mapping has 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 basin areas. This projection of lithofacies colors through much of the province should be regarded as an interpretive reconstruction for the sake of providing as complete a treatment of the conterminous United States as possible.

Lithofacies represented on the maps were determined by computation of the proportions of rock types in an interval. The lithofacies colors are used as symbols for 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. The use of the chart has been discussed in detail in "Paleotectonic Maps of the Permian System" (McKee, Oriel, and others, 1967, p. 58).

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 relation one to the other. 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 map pattern representing 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 either a few thick strata or 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 erratically distributed. For this reason, cross sections (pls. 9-A–9-G) have been prepared to show, within the scale limitations, the thickness of rock types and the vertical distribution and stratigraphic relations between rock types. The maps do not agree exactly with the cross sections at some locality points, inasmuch as they are more generalized, owing to greater scale limitation, and as a limited amount of interpretation was permitted to provide coherent lithofacies maps. A special facies overprint is used on these maps to denote areas in which chert content of the mapped unit exceeds 10 percent. The determination of this chert content can only be approximate, for the chert may occur in nodules or in highly irregular beds, and the nature of many of the data records does not permit precise quantitative determinations.

As explained in the discussion of isopach maps, only selected faults have been shown. Faults that are known to have moved in Mississippian time are shown in black; those that moved only after the Mississippian are shown in red.

Cross sections (pls. 9-A–9-G). — Fifty-six cross sections showing thickness, vertical distribution, and lateral relations of rock types were constructed. 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 close of its deposition, and the datum is projected across areas where erosion has removed the uppermost deposits. Position of the top of the remaining rocks below the projected horizontal datum is determined by comparison of remaining strata with complete sections and estimation of the thickness of missing rock. Use of a reconstructed datum results in a cross section that graphically shows the amount and location of regional sinking during deposition of the sediments, so the cross sections are interpretive to the extent that the amount of erosion is an estimate.

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

Most of the cross sections in the eastern part of the country consist of a single interval of the Mississippian System; most of the cross sections in the West consist of combined intervals of the Mississippian.

Paleotectonic maps (pl. 10, figs. 1–4). — The paleotectonic maps of each interval of the Mississippian are an outgrowth of the interpretive maps of preceding publications in this series. They show a considerable amount of information that is derived by the authors of this study from the basic data of the study, as well as from the interpretations of other authors. One feature, the present extent of each interval, is of a factual nature and is taken directly from the interval thickness maps (pls. 3-A–6-A). Restored isopachs indicate the original depositional thickness and extent of the interval and are theoretical reconstructions of the sediments deposited during each interval. The thickness of strata known or inferred to have been eroded since deposition is restored and included in interpretive thicknesses.

The postulated original limits of deposition are indicated by the zero isopach.

Directions of sediment transport are indicated by arrows, and the amount and texture of material transported is indicated by different lengths and widths of the arrows. Superimposed is an arbitrary system of colors indicating the gross tectonics of the interval: areas and relative amounts of subsidence or uplift, and areas of stability during the interval. A scheme of black tectonic symbols is overprinted where appropriate to show structural axes, direction of overthrusting, and slight or severe deformation. Intrusive igneous and volcanic activity are indicated by red symbols.

The paleotectonic maps differ from the sedimentary environment and paleogeographic maps in that the inferred margins of deposition (zero isopachs, pl. 10) do not necessarily represent ancient shorelines of the sea; some strata were deposited in continental environments. Although moderate or great uplift may be postulated for a source area, it does not necessarily imply the presence of mountains. Belts of maximum downwarping may not coincide with areas of maximum water depth in marine environments; the contrary is true in areas where deposition essentially kept up with subsidence.

Sedimentary environment maps (pl. 11, figs. 1–4). — These maps were prepared as an experiment and are highly generalized. Sedimentary environment maps ideally show the distribution of coexisting environments at a particular instant in time. Because environments shift position with time and may sometimes change rapidly, the identification of a common stratigraphic horizon is important. For this reason the preceding paleotectonic map publications included only selected areas of limited extent for mapping sedimentary environments. Generally, these were areas in which a time horizon could be identified in the rocks with some confidence, either with the aid of fossils or by mechanical devices — that is, proportionate division of sections to select an empirical horizon, or identification of facies related to a geologic event, such as a maximum marine transgression. These methods of establishing a time horizon were discussed in more detail by McKee, Oriel, and others (1967, p. 3–4).

In this report a series of four sedimentary environment maps of national scope is included. By agreement among the authors these are constructed on approximate time horizons that can be identified from region to region. Although the construction of the maps seems compatible between regions, great opportunity for error exists at the boundaries of regions. The horizons represented by the maps are upper interval A (late Kinderhook time — approximately mid-Rockford, mid-Chouteau of the midcontinent); mid-interval B (approximately middle Osage); mid-interval C (middle Meramec time — approximately early St. Louis of the midcontinent); and mid-interval D (approximately middle Chester time). Acknowledging the hazards of this national compilation and the problems of interregional as well as local correlation, the authors consider the maps to be fairly reasonable portraits of the national distribution of environments at the selected times.

The classification of the sedimentary environments on these maps is based on and modified from the system used in "Paleotectonic Maps of the Jurassic System" (McKee and others, 1956, pl. 9).

Paleogeographic maps (pl. 12, figs. 1–4). — Paleogeographic maps were prepared for each interval of the Mississippian System. They graphically present the "average" physical appearance of the conterminous United States during the time of each interval and show, by comparison of maps, the major geographic changes during the Mississippian Period. Like the sedimentary environment maps, a paleogeographic map properly should be constructed for a specific instant in geologic time, for the physical geography of the land is continuously changing. Figures 1–4, plate 12, are generalized for each interval. To some extent the degree of change that may take place during a single interval is shown by comparing figure 4*A* (early interval D) with figures 4*B* and 4*C* (late and end of interval D) in Eastern United States.

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New England Region

By ROBERT W. SCHNABEL

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

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

NEW ENGLAND REGION

By ROBERT W. SCHNABEL

ABSTRACT

Mississippian rocks, identifiable by fossils, are unknown in New England. In Maine a few small outcrops of sedimentary rocks have been correlated, on the basis of lithologic similarity, with fossiliferous Mississippian rocks in New Brunswick. Volcanoclastic rocks in the Boston basin of Massachusetts are post-Devonian and pre-Cretaceous and thus may be, in part, Mississippian in age. Volcanic rocks just west of the Narragansett basin of Rhode Island are post-Devonian and pre-Pennsylvanian in age as determined on the basis of structural discontinuities.

ROCKS OF POSSIBLE MISSISSIPPIAN AGE

Sedimentary rocks of possible Mississippian age crop out in four small downfaulted blocks in east-central Maine (fig. 2) (Larrabee and others, 1965). They have been designated Devonian through Pennsylvanian in age on the basis of lithologic similarity to rocks described in adjacent New Brunswick by Clark (1961). Some of the New Brunswick rocks have been dated as Early Mississippian by miospores (Belt, 1968).

A series of plutonic and volcanic rocks exposed just west of the Narragansett basin of Rhode Island, some plutonic rocks in eastern Massachusetts, and some volcanic rocks in the Boston basin are possibly of Mississippian age. The East Greenwich Group of Rhode Island, which includes the Spencer Hill Volcanics, the Cowesett Granite, a granite porphyry, and small masses of fine-grained granitic rocks, has been designated as Mississippian(?) by Quinn (1952). The Quincy Granite of northeastern Rhode Island has been assigned a Mississippian age both on geologic inferences and on a Rb-Sr whole rock radiometric age of 325±15 m.y. (Quinn and Moore, 1968; Bottino, 1963). However, work by R. E. Zartman (in Quinn and Moore, 1968) indicates a minimum age of Ordovician for the Quincy Granite. These discrepancies in age may be the result of miscorrelation of lithologically similar intrusive bodies as suggested by recent work in eastern Massachusetts (K. G. Bell, oral commun. 1970). A thick sequence of layered volcanoclastic rocks in the Boston basin may be, in part, Mississippian in age, but no fossil data have been found to corroborate the age.

No fossils have been reported from any of these possible Mississippian rocks in New England.

THICKNESS AND LITHOFACIES

Thicknesses of the sedimentary rocks of possible Mississippian age in east-central Maine cannot be estimated from their limited exposure, and the thickness of the Spencer Hill Volcanics in Rhode Island has not been recorded; therefore, thickness values have not been shown on the isopach maps (pls. 3-A–6-A, pl. 7).

The dominant rock types in east-central Maine are soft, red, pink, and gray locally carbonate-bearing siltstone; hard gray-green quartzite conglomerate; soft to hard poorly sorted coarse red conglomerate; interbedded red and light-gray quartzite conglomerate; and sandstone (Larrabee and others, 1965). In Rhode Island the Spencer Hill Volcanics are rhyolite flows with interbedded pyroclastic rocks and conglomerate; associated plutonic rocks in the East Greenwich Group are granite and granite porphyry. In the Boston basin possible Mississippian rocks are mostly pyroclastics.

In Maine, Massachusetts, and Rhode Island the areas that are underlain by these rocks are limited in size, but they are represented on the lithofacies maps (pls. 3-B–6-B) by stylized colors and symbols. Inasmuch as the interval assignment of the rocks is unknown, the colors and symbols have arbitrarily been shown on all the interval lithofacies maps.

OVERLYING AND UNDERLYING ROCKS

Because the sedimentary rocks in Maine are exposed as part of isolated fault slices, neither overlying nor underlying rocks were observed. The rocks in Canada with which they are correlated are apparently conformably

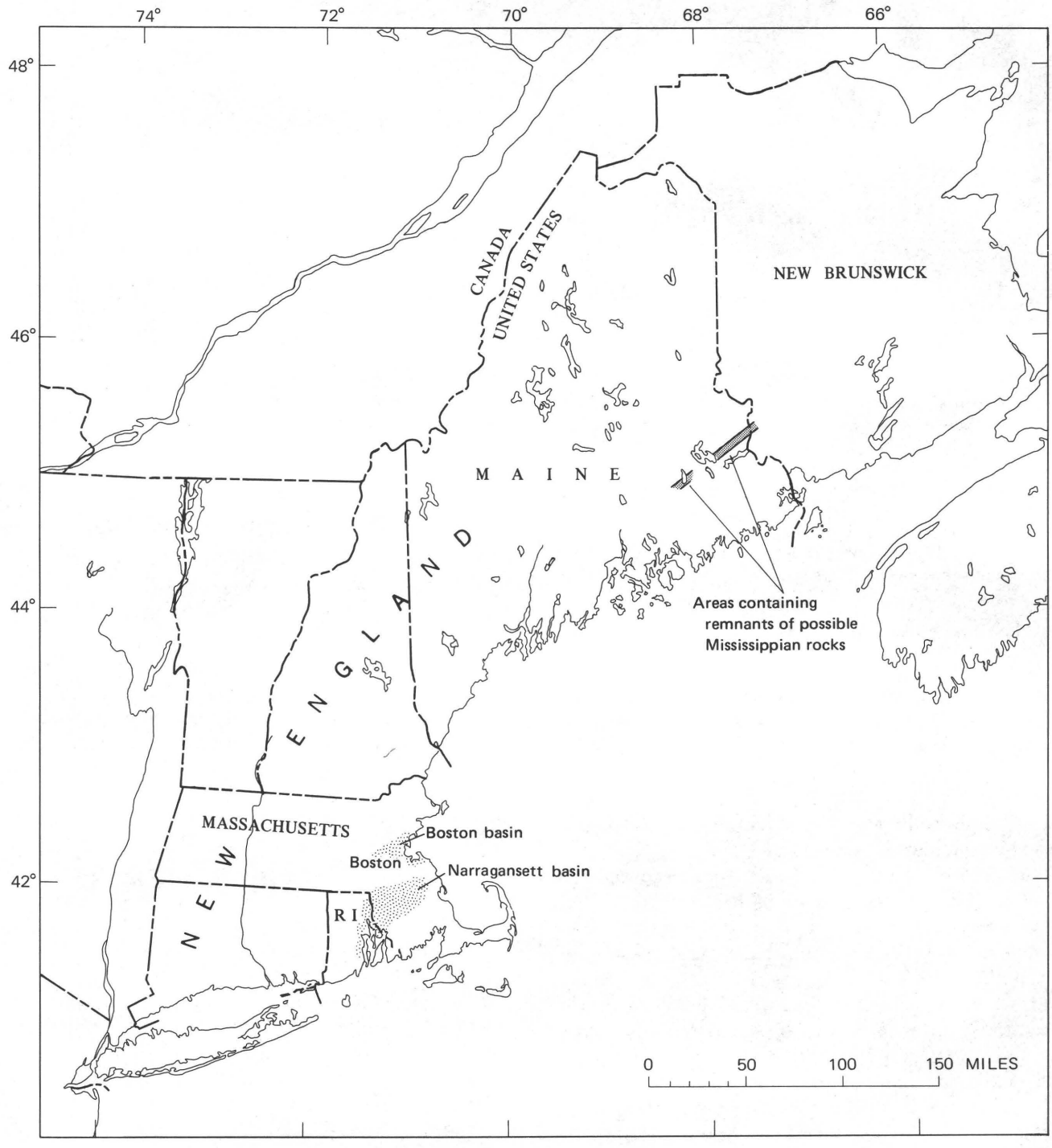

FIGURE 2. — New England and adjacent areas and geographic and tectonic features mentioned in the text.

overlain and underlain by sedimentary sequences of Pennsylvanian and Devoniage age (Belt, 1968). These relations are indicated in column 1 of plate 15 even though they are not evident in the outcrops in Maine. This projection of New Brunswick relations has not been shown on the maps of overlying and underlying rocks (pls. 2, 8).

Igneous rocks adjacent to the edge of the Narragansett basin include layered volcanics that were tilted to steep angles and were exposed by erosion before deposition of the overlying Pennsylvanian rocks (pl. 8) (Quinn and Moore, 1968). Although Pennsylvanian rocks are widespread in the Narragansett basin and, locally, they rest in angular unconformity on possible Mississippian rocks, the subsurface extent of the rocks of possible Mississippian age in the basin is unknown; thus, a symbol for overlying Pennsylvanian rocks is not shown on the map of overlying units (pl. 8). Nowhere in eastern Massachusetts or Rhode Island are intrusive rocks seen to cut fossiliferous strata; therefore, none are unquestionably Mississippian in age. Plutonic and metamorphic rocks of older Paleozoic age are present in the Narragansett basin area and are indicated in column 1 of plate 15 as underlying the East Greenwich Group; these are arbitrarily shown on the map of underlying rocks (pl. 2) as undifferentiated Devonian(?).

The thick sequence of rocks in the Boston basin is undated, and the age of rocks overlying possible Mississippain rocks is unknown; for this reason, on plate 8 no rocks are shown to overlie the Mississippian rocks in the Boston basin.

PALEOTECTONIC IMPLICATIONS

During Mississippian time nearly all of New England was probably an area of high rugged mountains (pl. 12, figs. 1–4) that formed during the Acadian orogeny. Off the present southeast coast, in the Boston-Rhode Island area, there may have been active volcanoes. Faulting and the development of rift valleys was occurring in New Brunswick (Belt, 1968) and these faults probably extended into east-central Maine. Only slight uplift is shown on the paleotectonic maps (pl. 10, figs. 1–4) for all the region except for diagrammatically represented subsiding fault trough areas in east-central Maine and neutral areas in Rhode Island and Massachusetts that received volcanic deposits.

SEDIMENTARY ENVIRONMENTS

In the New Brunswick-Newfoundland area of eastern Canada, Mississippian rocks consist of sediments derived from source areas along faults that were active during the time of deposition. The sedimentary rocks are coarse fanglomerates and conglomerates close to the fault scarps, and they grade into fine-grained fluvial and lacustrine sandstones and shales away from the faults (Belt, 1968). Presumably, the sedimentary environment in eastern Maine was similar to that in New Brunswick, and this is shown in a generalized manner as alluvial fans on the sedimentary environment maps (pl. 11, figs. 1–4) in this area. Volcanics in Rhode Island and Massachusetts may have accumulated as flows and ash falls in an alluvial or swamp environment.

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Appalachian Basin Region

By WALLACE DE WITT, JR. *and* LAURA W. MCGREW

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

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

THE APPALACHIAN BASIN REGION

By WALLACE DE WITT, JR., and LAURA W. MCGREW

ABSTRACT

The tectonic framework of the Appalachian basin during the Mississippian Period, which was inherited with little modification from the Devonian, consisted of an eastern arcuately elongate miogeosynclinal segment extending from eastern Pennsylvania to northern Mississippi and a western subparallel unstable shelf segment of slightly smaller size. Upland source areas of sedimentary, igneous, and metamorphic rocks bordered the basin on the north, east, and southeast. Repeated rejuvenation of these source areas released vast floods of terrigenous detritus, which accumulated as delta complexes and delta fluvial plains in the adjacent geosynclinal segment of the basin. The finer grained components accumulated as prodelta silts and muds in the deeper waters which covered the unstable shelf segment. Mississippian rocks in the Appalachian basin are locally more than 7,000 feet thick.

The northeastern source areas were the site of greatest uplift and folding during the Mississippian Period. Although modified locally, the position of the major elements of the tectonic framework of the Appalachian basin underwent little change during Mississippian time.

Generally, the basal boundary of the Mississippian System is not well marked in the Appalachian basin, and at many places deposition was continuous across the systemic boundary. Locally, in the Anthracite district of Pennsylvania, an episode of uplift and gentle open folding occurred early in Kinderhook time. Uplift of a small part of eastern West Virginia may have taken place concurrently. During Kinderhook and Osage time, when intervals A and B accumulated, wedges of terrigenous clastics filled the northern half of the basin to a maximum thickness of about 3,000 feet. Subaerially deposited deltaic sediments covered much of the geosynclinal segment at times and spread onto and across the shelf segment of the basin. Evaporites accumulated locally in a small tectonically controlled faulted basin in southwestern Virginia late in Osage time. In contrast, interval A is very thin or locally absent in the southern half of the basin. The overlying sheet of interval B carbonate strata is more than 200 feet thick.

In Meramec time, during which interval C accumulated, a marked flooding of the geosynclinal segment of the basin and a decreased supply of clastic detritus from the low-lying source areas adjacent to the southern two-thirds of the Appalachian basin produced a thick sequence of dominantly carbonate strata in all but the extreme northeastern part of the area. Shallow-water marine carbonate sediment accumulated in high- and low-energy environments to a thickness of more than 2,600 feet in southwestern Virginia, whereas a thinner sequence of red beds was deposited concurrently in parts of eastern Pennsylvania and extreme northeastern West Virginia. A sheet of calcareous sediment, ranging in thickness from 100 to 300 feet and in composition from quartzose sand in western Pennsylvania to nonclastic carbonate mud in Tennessee and northern Mississippi, blanketed the shelf segment of the basin in Meramec time. Local warping of the basin exposed the recently deposited strata in parts of central Ohio, and some shallow scour channels trenched the recently exposed rocks. Elsewhere, the basin was quiescent.

Interval D shows a more complex depositional pattern. Widespread deposition of carbonate sediment continued throughout much of the Appalachian basin in early Chester time. However, strong uplift and rejuvenation of source areas peripheral to the basin produced a great influx of terrigenous debris which filled the geosynclinal segment of the basin by late middle Chester time and displaced the sea from all but the western distal extremities of the shelf segment before the close of Chester time. Chester-age sediments accumulated in a subaerial environment in eastern Pennsylvania to a thickness of more than 5,000 feet; to the south a mixed sequence of marine and nonmarine sediments in the Greendale syncline of southwestern Virginia exceeded 4,500 feet in thickness; and from Tennessee southwestward into northern Mississippi interval D rocks are as much as 2,000 feet thick. They are predominantly shale with some interbedded marine sandstone and limestone.

Conspicuous disconformities at the base of the Lower Pennsylvanian rocks in outcrops in northeastern Kentucky, eastern Ohio, and northwestern Pennsylvania pass into paraconformities downdip into the basin and are not present in the axial part of the geosynclinal segment of the Appalachian basin. Although some tilting, warping, and erosion of the northern edge of the basin occurred near the end of Chester time, deposition was continuous from Late Mississippian to Early Pennsylvanian along the axis of the geosyncline.

REGION DEFINED

As commonly defined, the Appalachian basin consists of the elongate area underlain by the large mass of downwarped Paleozoic rocks between the Blue Ridge anticlinorium on the east and the crest of the Cincinnati arch on the west (fig. 3). As here defined, the basin is about four times as long as it is wide. Its axial trend is roughly N. 40° E. although local segments may depart considerably from the average trend. The basin contains two distinct structural segments: an eastern segment composed of complexly folded and faulted miogeosynclinal rocks, and a western segment composed of gently folded to undeformed rocks in the foreland or

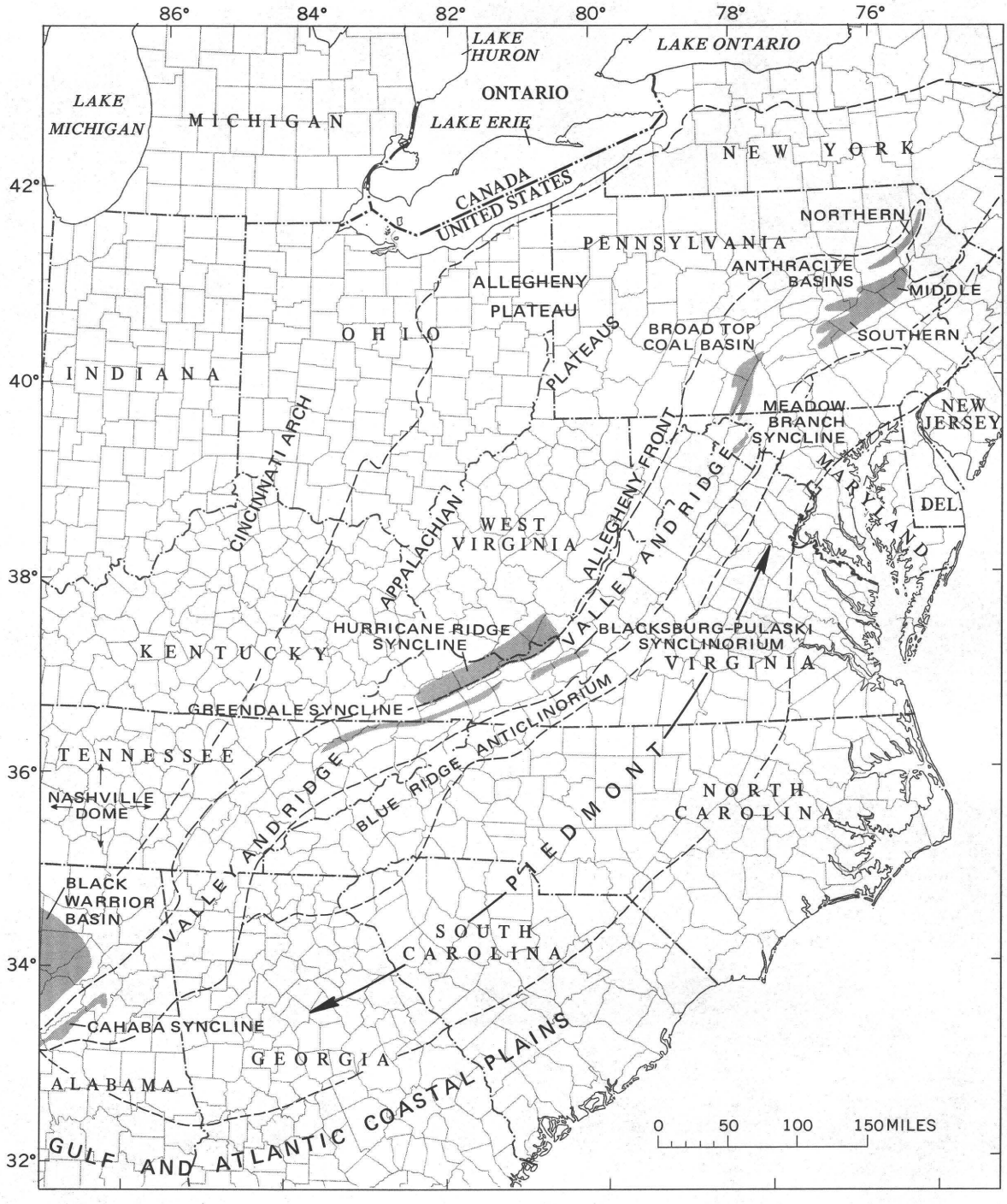


FIGURE 3. — Some physiographic and structural elements within and adjacent to the Appalachian basin showing some features mentioned in text.

slightly unstable shelf part of the basin. Throughout the basin the two segments are separated at the Allegheny Front, which marks the change from strongly deformed to gently deformed rocks. The strongly folded segment of the basin corresponds areally to the Valley and Ridge physiographic province and the slightly deformed western segment corresponds to the Appalachian Plateaus province.

Most of the preserved Mississippian rocks in the basin underlie the Appalachian Plateaus province; although locally, Mississippian strata remain infolded in the structurally deeper synclines of the Valley and Ridge province. The Anthracite basins of eastern Pennsylvania, the Greendale syncline of southwestern Virginia and northeastern Tennessee, and the Cahaba syncline of central Alabama are examples of synclines

which contain infolded Mississippian rocks in the Valley and Ridge province. The Allegheny Front marks the eastern edge of the main sheet of Mississippian age rocks. The northern boundary of these strata is a ragged line extending west from the Northern Anthracite basin of Pennsylvania to Huron County (fig. 4) in north-central Ohio. The western side of this roughly triangular patch of rocks extends south along the east flank of the Cincinnati arch into south-central Kentucky, east-central Tennessee, and northern Alabama where it is buried by coastal plain rocks in the vicinity of the Black Warrior basin of northwestern Alabama and northern Mississippi (fig. 3).

PALEOGEOLOGY

UNITS UNDERLYING MISSISSIPPIAN

Throughout most of the Appalachian basin, Upper Devonian strata (pl. 2) underlie the Mississippian rocks. Paleontologic data indicate that the Devonian rocks are for the most part in the Bradford stage, and at many places deposition appears to have been continuous across the systemic boundary. However, rocks older than Bradfordian underlie the Mississippian sequence in the following four areas in Alabama: in the Coosa River valley southeast of Birmingham, the Frog Mountain Sandstone of Early or Middle Devonian age; about 12 miles northwest of Birmingham, the Red Mountain Formation of Silurian age; in the northwestern corner of the State, undifferentiated Silurian rocks; and in the north-central part of the State, undifferentiated Ordovician rocks. In Polk County, Ga., the Rockmart Slate of Ordovician age underlies Mississippian rocks.

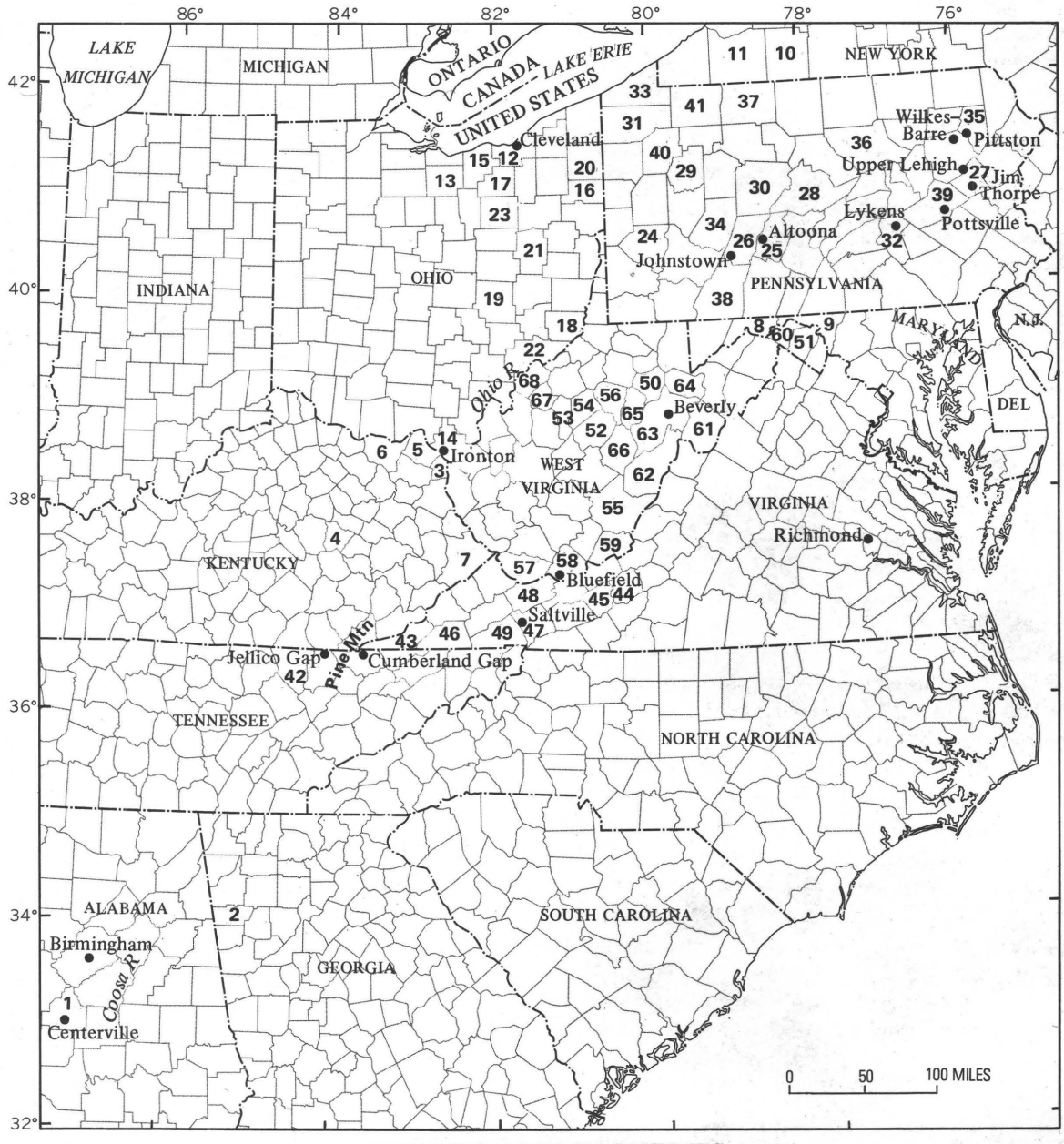
Upper Devonian strata exhibit a diversity of rock types ranging from conglomerate and coarse-grained sandstone in the continental red bed sequence of the Catskill and Hampshire Formations (pl. 15) to fine-grained black shale in the Ohio and the Chattanooga Shales. A largely continental sequence of red and purple rocks makes up the Catskill Formation in the Anthracite region of Pennsylvania and south along the Allegheny Front to Altoona in Blair County. From Blair County south into the Virginias the name Hampshire Formation is frequently used for this sequence of red beds. The names Catskill and Hampshire have been used interchangeably for the red-bed facies at the discretion of individual geologists. The red beds of the Catskill or Hampshire intergrade westward with greenish-gray and olive-drab marine sandstone, siltstone, and shale. Many of these strata accumulated in a well-aerated shallow marine environment near the shore of the Late Devonian sea, and the local abundance of fossils in these strata attests to the large and diverse benthonic fauna that existed throughout much of the Late Devonian. To the west and south of these shallow-water deposits the rocks are dominantly dark gray, dark brown, or black; strongly indicative of their euxinic environment of accumulation. Fossils are relatively scarce in the dark rocks excepting conodonts which are locally abundant.

LOWER BOUNDARY OF MISSISSIPPIAN

The stratigraphic position of the lower boundary of the Mississippian System in the Appalachian basin has provided much lively controversy during the last 100 years. Several recent studies have greatly aided in establishing locally acceptable boundaries and have considerably reduced the area in which boundary problems exist.

In the western part of the basin throughout much of Ohio, western West Virginia, eastern Kentucky, and a part of southwestern Virginia, the boundary between the Mississippian and Devonian Systems is generally placed at the base of a zone of marine fossils in the basal part of the Bedford Shale. The underlying Cleveland Member of the Ohio Shale contains definite Late Devonian fauna and is equivalent to the Gassaway Member of the Chattanooga Shale (Hass, 1956, p. 23). The age of the fauna in the Bedford is equivocal and has been considered as Devonian or Mississippian by previous geologists. However, the age of most of the Bedford Shale excepting the basal fossil bed is Mississippian (de Witt, 1970, p. G10). The gray fossiliferous Bedford contrasts strongly with the underlying black Ohio or Chattanooga Shale, and the boundary can be readily located in both surface and subsurface sections (pl. 15). In Kentucky the Bedford Shale feathers out of the sequence along a line between Estill and Pike Counties (pl. 9-B, sec. *j–j'*, interval A). South of this line the base of the Mississippian System lies within the uppermost few feet of black shale of the Chattanooga Formation, and the systemic boundary must be located by paleontologic criteria. The Chattanooga contains at most only a few feet of beds of Mississippian age and has arbitrarily been excluded from this study throughout its extent. The lithologic contact at the top of the Chattanooga is the closest feasible contact to the systemic boundary.

In eastern Ohio, western Pennsylvania, and a part of northern West Virginia, the Bedford Shale is absent or unrecognizable, and the basal boundary of the system is here arbitrarily placed at the base of the *Murrysville sand* in Pennsylvania or its equivalent in West Virginia which may be identified as *Murrysville* or "*Gantz*" by well drillers. In Erie and Crawford Counties, Pa., the basal boundary of the system is the base of the Cussewago Sandstone, the surface equivalent of the *Murrysville sand*. The Cussewago Sandstone is absent to the east in Warren and McKean Counties, and the



IDENTIFICATION OF COUNTIES

ALABAMA
1. Bibb

GEORGIA
2. Polk

KENTUCKY
3. Boyd
4. Estill
5. Greenup
6. Lewis
7. Pike

MARYLAND
8. Allegany
9. Washington

NEW YORK
10. Allegany
11. Cattaraugus

OHIO
12. Cuyahoga
13. Huron
14. Lawrence
15. Lorain
16. Mahoning
17. Medina
18. Monroe
19. Muskingum
20. Trumbull
21. Tuscarawas
22. Washington
23. Wayne

PENNSYLVANIA
24. Allegheny
25. Blair
26. Cambria
27. Carbon
28. Centre
29. Clarion
30. Clearfield
31. Crawford
32. Dauphin
33. Erie
34. Indiana
35. Lackawanna
36. Lycoming
37. McKean
38. Somerset
39. Schuylkill
40. Venango
41. Warren

TENNESSEE
42. Scott

VIRGINIA
43. Lee
44. Montgomery
45. Pulaski
46. Scott
47. Smyth
48. Tazewell
49. Washington

WEST VIRGINIA
50. Barbour
51. Berkeley
52. Braxton
53. Calhoun
54. Gilmer
55. Greenbrier
56. Lewis
57. McDowell
58. Mercer
59. Monroe
60. Morgan
61. Pendleton
62. Pocahontas
63. Randolph
64. Tucker
65. Upshur
66. Webster
67. Wirt
68. Wood

FIGURE 4. — Locality map of the Appalachian basin region.

base of the Mississippian is the base of the fine-grained Corry Sandstone or its coarse-grained equivalent, the Knapp Formation. The presence of a conodont fauna in the *Kushequa Shale Member* in the basal part of the Knapp Formation similar to the fauna in the base of the Bedford Shale of Ohio indicates the equivalency of the basal Bedford and the Knapp. The beds below the Knapp contain a typical Late Devonian brachiopod fauna.

Location of the basal Mississippian boundary in outcrops along the east face of the Allegheny Plateau and in the outliers of the Valley and Ridge province — in the triangular area between McKean and Carbon Counties, Pa., and Greenbrier County, W. Va. — is complicated by the fact that the systemic boundary lies within the red-bed facies of the Catskill or Hampshire Formations. Most workers in the Appalachians during the second half of the 19th and the first half of the 20th centuries commonly drew the base of the Mississippian (Carboniferous) at the base of the stratigraphically oldest white sandstone or quartzite overlying the red beds of the Catskill or Hampshire. However, I. C. White (1881–83) showed that a transition zone several hundred feet thick separated the dominantly red Catskill below from the massive conglomeratic quartzites of the Pocono Formation above. The transition zone contains intercalations of rocks characteristic of both the Catskill and the Pocono. White believed that the rocks of the transition zone were of Mississippian (Carboniferous) age; however, his conclusion was largely ignored by other workers.

Studies by Read (1955), Dally (1956), Trexler, Wood, and Arndt (1962), and Read and Mamay (1964) have demonstrated that White was correct and that in much of the northeastern part of the Appalachian basin the systemic boundary is within the Catskill or Hampshire red-bed sequence. Both floral evidence in Pennsylvania and faunal evidence in West Virginia indicate that in several places the upper part of the red-bed sequence is of Early Mississippian age. In this paper, the basal Mississippian boundary has been arbitrarily placed at the base of the massive sandstone in the red beds that most closely approximates the time plane indicated by paleontologic control. At some places where data are scant, the boundary is the top of the red beds; elsewhere, the boundary has been drawn within the sequence where control existed.

In the Valley and Ridge province southeast of Greenbrier County, W. Va., the base of the Mississippian is difficult to locate because the red beds of the Hampshire feather out of the section and the clastic rocks of Late Devonian and Early Mississippian age are gradational. However, Glover (1953) has demonstrated the Mississippian age of the *Parrott Formation*, the basal unit of the Mississippian in Pulaski County, Va., and he has presented correlations which showed that Reger's *Broad Ford Sandstone* is of Late Devonian age at the type locality. The *Broad Ford Sandstone* in parts of southwestern Virginia contains a *Cyrtospirifer* fauna which links it with the so-called "Chemung" Formation — a catch-all name for non-red marine clastic rocks of Late Devonian age which underlie the oldest Mississippian rocks west and south of the red beds of the Catskill or Hampshire. The name Chemung Formation or Chemung Group has fallen into disrepute and has been abandoned in central New York which was the type area for the unit.

In the southern part of the Appalachian basin from Tennessee south, the lower boundary of the Mississippian System has, in this report, been placed at the base of the Maury Formation. The assignment is based largely on the conclusions of Hass (1956, p. 23–24) and Conant and Swanson (1961, p. 16–19, 67) who summarized the controversy over the age of the Maury Formation and Chattanooga Shale. The change in color from the black of the Chattanooga Shale to the green of the Maury Formation and the phosphate nodules generally present in the lower part of the Maury Formation are the physical criteria which have been used for picking the lower boundary of the Mississippian System.

PALEOTECTONIC IMPLICATIONS

The similarity of many of the oldest Mississippian rocks to the youngest Devonian strata below them, the vertical gradation of units across the systemic boundary at many places, and the absence of marked changes in the regional pattern of sedimentation at the beginning of the Mississippian indicate that the tectonic framework of the basin and adjacent source areas did not change abruptly at the advent of the period. However, the intercalation of coarser grained strata in the basal Mississippian in the northeastern part of the basin indicates an increased uplift of source areas east and northeast of the Appalachian basin.

In the southern part of the Appalachian basin the Late Devonian was a time of tectonic quiescence. Still, shallow seas covered much of what is now Tennessee, northwestern Georgia, northern Alabama, and northeastern Mississippi. Conant and Swanson (1961) suggested that a land area of low relief existed in western North Carolina with lowlands bordering the Late Devonian sea to the south, southeast, and southwest. These lowlands were a subaerial continuation of a submerged peneplain which formed the bottom of the Chattanooga sea. A landmass may have existed in the Ozark area to

the west and northwest and several islands lay within the sea as indicated by the localities where earliest Mississippian rocks rest directly on rocks older than Chattanooga Shale.

Sandy facies in the Chattanooga Shale in northeastern Mississippi and adjacent parts of Tennessee and near Centerville (fig. 4) in Bibb County, Ala., suggest the proximity of large streams entering the Chattanooga sea at these localities (Conant and Swanson, 1961).

The basal Mississippian Maury Formation, over most of its extent, is entirely of Kinderhook age; but at a few localities in north-central Tennessee, the oldest beds of the formation are probably of very late Devonian age, and in some places the uppermost beds are probably of early Osage age (Hass, 1956, p. 23, 24; Conant and Swanson, 1961, p. 67). Conant and Swanson (1961, p. 68) suggested that it is a transitional unit marking a gradual regional change from the poorly oxygenated waters of Late Devonian time to the well-aerated seas of later Mississippian. They also suggested that, although there is very little physical evidence of a hiatus between the Chattanooga Shale and Maury Formation, deposition may have been interrupted locally in Late Devonian time by shallowing of the sea and by some warping and planation of the sea bottom. Phosphate and glauconite are conspicuous constituents of the Maury Formation. Goldman (1922) cited instances in which the presence of these minerals along with a minimum of detrital material are indicative of important stratigraphic hiatuses.

INTERVAL A

RELATION OF INTERVAL A TO INTERVAL A-B

In most of the western and southern parts of the basin, the rocks of interval A, which are of Kinderhook age, can be readily distinguished from the rocks of interval B, which are of Osage age (pl. 15). On the other hand, in the eastern part of the basin, the fossils and key beds necessary to separate the two intervals are missing at many places, and the thick sequence of clastic rocks cannot be subdivided regionally into component intervals A and B. Local subdivision is possible, but the stratigraphic relations of individual units in the sequence are not well established, and much paleontologic work remains to be done before intervals A and B can be differentiated and delineated with certainty throughout the Appalachian basin. Because of the relative impossibility of distinguishing intervals A and B throughout much of the eastern part of the basin, the strata comprising these intervals in that area here are treated as a single interval combining the rocks of Kinderhook and Osage ages in a unit mapped as interval A-B. Combined interval A-B is shown on both plates 3 and 4 for convenience in comparing with the differentiated intervals.

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Throughout much of eastern Ohio, eastern Kentucky, western West Virginia, and a part of southwestern Virginia, interval A is made up in ascending stratigraphic order of the Bedford Shale, Berea Sandstone, and Sunbury Shale (pl. 15; pl. 9-B, sec. *k–k'*, interval A). In much of eastern Kentucky from Pike County northwest to Lewis County, siltstone and sandstone fill the interval below the Sunbury Shale, and well drillers commonly refer to the sequence as the *Berea sand*, although surface data demonstrate that much of the sequence is of Bedford age (Morris, 1966).

In northeastern Ohio and part of northwestern Pennsylvania, the Cussewago Sandstone is the basal part of interval A, and the top of the interval is not well marked because the Sunbury Shale grades laterally into the basal part of a thicker unit of very dark brownish-gray shale, the Orangeville Shale. The Shellhammer Hollow Formation of Erie and Crawford Counties, Pa., includes equivalents of the Bedford Shale, Berea Sandstone, Corry Sandstone, and the basal part of the Sunbury Shale (de Witt, 1951, pl. 2). The Cussewago Sandstone is absent in Warren and McKean Counties to the east, where the coarse-grained conglomeratic Knapp Formation or its fine-grained equivalent, the Corry Sandstone, composes most of interval A.

In much of western Pennsylvania and adjacent northern West Virginia, the coarse-grained *Murrysville sand*, the subsurface equivalent of the Cussewago Sandstone, occupies the lower two-thirds of interval A. The Bedford Shale and Berea Sandstone are not present, and the Sunbury Shale has lost its identity in a thick sequence of dark-gray silty shale between the *Murrysville sand* below and the *2d gas sand* above. To the east in central Pennsylvania the *Murrysville* and related strata merge into the thick wedge of coarse-grained clastic rocks, the Pocono Formation, that compose interval A-B.

Throughout the southern part of the Appalachian basin from northeastern Tennessee southwestward into northern Mississippi interval A is represented by a single unit, the Maury Formation (pl. 15). It consists of grayish-green mudstone, claystone, shale, siltstone, and some glauconitic sandstone, and it contains abundant calcium phosphate. The phosphate generally occurs as concentrations of nodules in the lower part and as sparse, scattered nodules and plates in the remainder of the formation. Although thin — generally

2–5 feet thick — the Maury Formation is believed to represent deposition during most, if not all, of interval A or Kinderhook time (Conant and Swanson, 1961, p. 67).

UPPER BOUNDARY OF INTERVAL A

The top of the black *Siphonodella*-bearing Sunbury Shale (pl. 15) is here arbitrarily selected as the upper boundary of interval A in the west part of the basin because the horizon is widespread and readily identified in both surface and subsurface sections. Although some of the overlying Cuyahoga Formation of Ohio is of Kinderhook age (Root and others, 1961, p. 7), the lack of paleontologic control, the absence of widespread key beds, and the complex facies relations in the unit prevent its separation into intervals A and B.

The Sunbury grades into the basal part of the Orangeville Shale, a thick unit of dark-gray silty shale in northeast Ohio (pl. 15, col. 9), and in this area and in adjacent northwest Pennsylvania the upper boundary of interval A is lifted slightly to the base of the drillers' *2d gas sand* or its lateral equivalents which the drillers call *Gable, Welch Stray*, or *Weir sand*. Although these sand bodies are younger than the Sunbury and are probably not absolute time equivalents, they are the most widespread group of key beds in this part of the basin. The *2d gas sand* has been traced from the Pittsburgh area eastward to the vicinity of Johnstown, Cambria County, Pa., (Pepper and others, 1954, pl. 14) east of which it cannot be identified with certainty in the thick clastics of interval A-B.

Throughout the southern part of the Appalachian basin the upper boundary of interval A is the top of the Maury Formation, and its upper surface is generally sharply defined by the overlying Fort Payne Formation. Occasionally, the blue-green shaly chips of claystone so characteristic of the upper part of the Maury are also present in the lower 2–3 inches of the Fort Payne. At these places the contact is placed at the lowest occurrence of limestone or chert. However, both the upper and lower boundaries of interval A are commonly difficult to select because the Maury is so thin that at many places it is lost completely in well cuttings and is difficult to find in outcrops.

THICKNESS TRENDS

In much of the western part of the Appalachian basin region, interval A is less than 100 feet thick; however, at two localities in central Ohio it is more than 200 feet thick (pl. 3-A). Locally, it is interrupted by several linear belts of thicker strata in which the interval is more than 100 feet thick. One linear belt extends from north-central Ohio (Lorain County) southward across Ohio and eastern Kentucky. A second extends from south-central Pennsylvania (Somerset County) northwestward to the vicinity of Cleveland in north-central Ohio. These two elongated thick masses are joined by a short cross-trending belt extending from northeastern Ohio (Mahoning County) to central Ohio (Muskingum County). Interval A also attains a thickness of more than 100 feet in a small area in northwestern Pennsylvania (Warren and McKean Counties).

With the exception of one locality in Scott County, northeastern Tennessee, where it is 25 feet thick, interval A throughout the southern part of the Appalachian basin is uniformly a very thin unit, 1–7 feet thick.

LITHOFACIES TRENDS

Commonly in Pennsylvania and adjacent parts of West Virginia, the areas of thick strata correspond to areas of coarse-grained quartzose sandstone. The conglomeratic Knapp Formation makes up a thick interval A in Warren and McKean Counties, and a sheet of pebbly *Murrysville sand* matches the belt of thick strata in western Pennsylvania and northeastern Ohio (pl. 3-B). In Trumbull County, Ohio, the *Murrysville* is overlapped from the northwest by a tongue of finer grained *Berea sand*, which is coarser grained to the northwest near Cleveland. Correspondence of a thick interval A with coarse-grained rocks is less in Ohio, Kentucky, and western West Virginia than in Pennsylvania.

The maximum thickness of the Bedford Shale, representing the Red Bedford delta (fig. 6; Pepper and others, 1954, p. 45) underlies a broad area along the western edge of the preserved rocks in central Ohio (from Huron County to Lawrence County). Rocks in this area are predominantly shale and mudrock. In contrast the Berea Sandstone dominates the lithofacies map (pl. 3-B) in northern Ohio, where it is thickest and coarsest grained. The Berea thins and becomes finger grained to the south as the lithofacies pattern clearly shows. A belt of predominantly sandy rocks extends from Tuscarawas County south across Ohio and West Virginia to the Virginia border in the vicinity of Bluefield. In West Virginia several narrow sinuous tongues of sandy strata project eastward into the adjacent area of thin shaly rock.

Throughout eastern Tennessee, northwestern Georgia, northern Alabama, and northeastern Mississippi, interval A, as represented by the Maury Formation, consists almost entirely of claystone and siltstone. Minor amounts of sandstone present in some areas do not affect the lithofacies pattern. Phosphate and glauconite are conspicuous components of the entire unit.

SOURCES AND ENVIRONMENTS

The sandy rocks that make up the sheet of Cussewago Sandstone and *Murrysville sand* had an eastern source area (pl. 10, fig. 1; pl. 12, fig. 1). The rocks were deposited as the nearshore parts of a large delta complex (pl. 11, fig. 1) that covered parts of western and central Pennsylvania, eastern Ohio, and northern West Virginia (Pepper and others, 1954, p. 40).

The source area for the red Bedford Shale and the Berea Sandstone of Ohio lay far to the north in the eastern part of the Canadian Shield (Pepper and others, 1954, p. 99). The Knapp Formation of northwestern Pennsylvania, another delta deposit, similarly had a northern source. It is thickest and coarsest grained in the north part of its outcrop and becomes progressively thinner and finer grained to the south. Both the Knapp and the Berea accumulated as deltas which were built southward into a shallow epicontinental sea by rivers that headed in highlands in eastern Canada (Pepper and others, 1954, pl. 12).

The *Berea sand* of West Virginia, like the *Murrysville sand* of Pennsylvania, was derived from an eastern source. Sand from the eastern source was carried west into the epicontinental sea in western West Virginia and eastern Kentucky by several rivers. Sand-filled channels of the ancient Gay-Fink and Cabin Creek Rivers (fig. 5) have been known for many years (Pepper and others, 1954, p. 75–79). Recent drilling in Tazewell County, Va., has confirmed the presence of the channel of the Virginia-Carolina River (J. T. Galey oral commun., 1967) another of the large streams that drained from the eastern source area.

The main source of clastic sediment for the Maury Formation was probably surficial material from the lowland areas that were submerged by the Early Mississippian sea (Conant and Swanson, 1961, p. 69). Accumulation was slow, resulting in deposition of only 2–5 feet of sediment throughout most of the southern Appalachian basin during entire Kinderhook time. The predominantly fine-grained character of the Maury Formation suggests extensive reworking of the material by currents and waves before final deposition. Minor amounts of coarser detrital material in the formation were probably derived from the low islands of Late Devonian rocks as they were submerged by the deepening, but probably still fairly shallow, Early Mississippian sea.

Phosphate was presumably precipitated from the seawater. It occurs as nodules in a thin persistent bed in the lower part of the formation and more sparsely scattered throughout the upper part and as tabular masses which may be parallel to or cut across bedding in the upper part of the formation. Conant and Swanson (1961, p. 66) believed that the nodular phosphate formed simultaneously with the accumulation of the mud and shortly afterward but that the tabular masses that cut across bedding are a result of local migration and redeposition of the phosphate material after burial and compaction. Glauconite, presumably an alteration product, commonly occurs in the Maury Formation as abundant visible grains where it is associated with sandstone, but it is rarely visible in the upper claystone part of the formation. There it may occur as "pigmentary glauconite" (Conant and Swanson, 1961, p. 64) and possibly gives the claystone its characteristic green color. Some writers believe (for example, Conant and Swanson, 1961, p. 68) that both phosphate and glauconite are formed during periods of slow clastic sedimentation and that glauconite is formed in a reducing marine environment.

PALEOTECTONIC IMPLICATIONS

The strata of interval A record several episodes of uplift in source areas to the north and east of the Appalachian basin. The eastern source area produced two

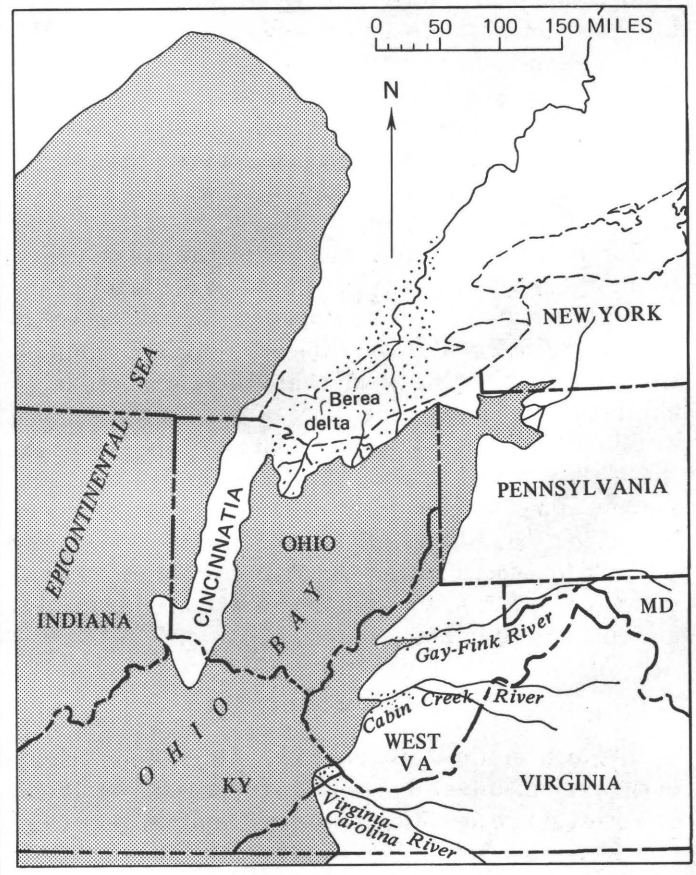


FIGURE 5. — Paleogeographic map of middle Berea time showing the location of the Gay-Fink, Cabin Creek and Virginia-Carolina Rivers. (Modified from Pepper and others, 1954.)

floods of clastic debris — the *Murrysville sand* of Pennsylvania and the *Berea sand* of West Virginia — and appears to have been more active than the northern source area which produced only one flood of coarser clastics — the Berea Sandstone of Ohio.

The presence of a large delta, the Red Bedford delta (fig. 6) with fringing offshore bars, from Cuyahoga County south across Ohio to Lawrence County at the Ohio River shows that the epicontinental sea was shallow and locally nonexistent during a part of Bedford time (Pepper and others, 1954, pl. 13B). Submergence of the delta by a late Bedford sea indicates that the sea began to deepen toward the close of Bedford time. However, the presence of a well-established shoreline in western West Virginia and eastern Kentucky and the scour of river channels to the east of the shoreline indicate that a major regression of the sea took place in the basin during early Berea time. In all probability this regression was the maximum retreat of the Early Mississippian sea. A ridge of shoals and low islands, Cincinnatia, marked the crest of the Cincinnati arch across western Ohio and in north-central Kentucky during much of the Kinderhook (Pepper and others, 1954, pls. 13A–I). Cincinnatia was sufficiently positive to prevent exchange of sediments between the Appalachian and Illinois basins during accumulation of the Bedford Shale and Berea Sandstone. If sediments were contributed to the Appalachian basin by Cincinnatia, they have not been identified.

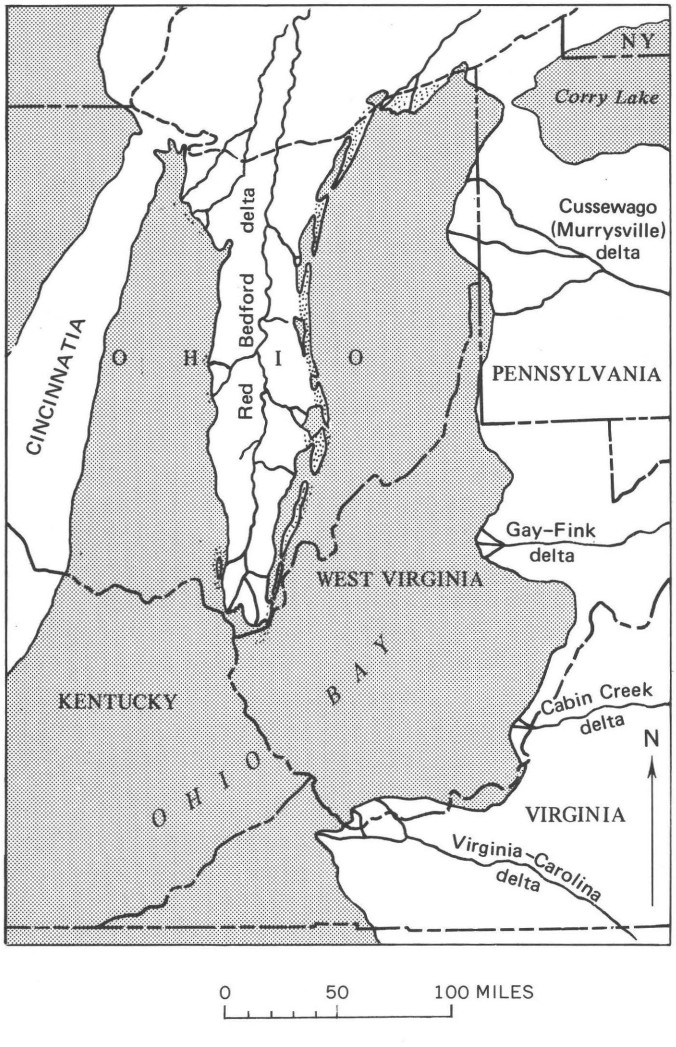


FIGURE 6. — Paleogeographic map of early Bedford time showing the Red Bedford delta with associated ofsshore bars. (Modified from Pepper and others, 1954.)

Gradual and continual submergence of the coastal plain by transgressive overlap occurred throughout the rest of Berea time. During the late Kinderhook the western part of the Appalachian basin returned to a period of quiescence and stagnation similar to that of the Late Devonian. Black mud of the Sunbury accumulated in an euxinic environment. Only small amounts of fine-grained detritus were carried into this part of the basin from land areas to the north and east. The widespread and relatively uniform sheet of black Sunbury Shale suggests strongly that Cincinnatia was submerged and did not interrupt the deposition of black mud, which covered adjacent parts of the Appalachian, Illinois, and Michigan basins in the late Kinderhook.

In the southern part of the Appalachian basin interval A was a time of gradual regional change from the poorly oxygenated waters of the Late Devonian to the well-aerated sea of later Mississippian time (Conant and Swanson, 1961, p. 68). Gradual subsidence caused the sea to spread widely, submerging the Late Devonian islands and encroaching upon the peneplaned shoreline. Conant and Swanson (1961, pl. 14) inferred that the landmass which bordered the Late Devonian and earliest Mississippian sea extended across the Southern United States westward from the Carolinas across southern Georgia and Alabama, northward up the Mississippi embayment region into part of Tennessee, Kentucky, Missouri, and Arkansas and then westward again across Louisiana and into east Texas. (See pl. 12, fig. 1.)

INTERVAL A-B

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

In contrast to the many well-marked stratigraphic units occupying intervals A and B in the western part of the basin, a thick wedge of dominantly coarse-grained clastic strata of Kinderhook and Osage age fills the eastern part of the basin. These rocks are poorly known in much of the area. The absence of widespread key beds and well-marked floral and faunal zones at many places precludes subdivision of the mass into discrete

intervals. Consequently, these Lower Mississippian rocks are herein mapped collectively as combined interval A-B and are shown combined on the accompanying maps (pls. 3-A, 3-B, 4-A, 4-B).

Interval A-B received its most intensive study in the Anthracite region of eastern Pennsylvania. Two thick units of clastic rock, the Spechty Kopf Member of the Catskill Formation (Trexler, and others, 1962, p. C36) and the overlying Pocono Formation, compose the interval (pl. 15; pl. 9–B, sec. *h–h'*, interval A). The Spechty Kopf Member was originally named the *Pocono-Catskill transitional group* by White (1883, p. 47–52) because it contains rocks that are characteristic of both the Catskill red beds and the quartzitic conglomerates of the Pocono. Red, purple, greenish-gray, and black shale are intercalated with gray and white pebbly sandstone and conglomerate. Read (1955, p. 8; Read and Mamay, 1964, p. K5) obtained an *Adiantites* flora from the middle and upper parts of the Spechty Kopf Member which he designated as lower Pocono. The *Adiantites* flora established an Early Mississippian age for most of the member. As yet a flora has not been obtained from the lower part of the member; however, the conformable relation and gradational boundary of the member with the subjacent Buddys Run Member of the Catskill Formation suggest a Late Devonian age for the basal beds of the Spechty Kopf Member.

Trexler, Wood, and Arndt (1962, p. C37) assigned a maximum thickness of 2,400 feet to the Spechty Kopf Member at the type area near Lykens, adjacent to the southwestern part of the Southern Anthracite basin; whereas D. M. Hoskins (oral commun., 1965) stated that the member was about 1,000 feet thick in the type area. Adopting a conservative view, we have here tentatively accepted the lesser thickness figure pending resolution of these conflicting data. In the type area near Lykens, the Pocono Formation locally lies with angular unconformity upon the Spechty Kopf (Trexler and others, 1961, p. B84), and locally pre-Pocono folding and erosion have combined to remove more than 300 feet of the member. The Spechty Kopf Member thins away from the type area and appears to be about 300 feet thick in the vicinity of the Northern Anthracite basin. The time span and topographic relief on the unconformity at the top of the member also appear to decrease in magnitude away from the Southern Anthracite basin. The hiatus is represented in the Northern Anthracite basin by a gentle disconformity or an obscure paraconformity.

The massive and resistant Pocono Formation makes up much of interval A-B in the Anthracite basins and in the Broad Top basin of Pennsylvania, as well as along the Allegheny Front in Pennsylvania, Maryland, and the Virginias. In the Southern Anthracite basin, the Pocono is divided into members, the Beckville below and the Mount Carbon above (Trexler and others, 1962, p. C38). Both members are coarse-grained conglomeratic sandstone that locally contains some shale and impure coal. An *Adiantites* flora in the subjacent Spechty Kopf suggest that the Beckville may be of Kinderhook age, whereas the Osage age of the Mount Carbon is indicated by a *Triphyllopteris* flora (Read, 1955, p. 8–15; Trexler and others, 1962, p. C36). The two members of the Pocono cannot be distinguished north of the Southern Anthracite basin. In the Northern Anthracite basin the Pocono contains a massive conglomerate in the basal part, the Griswold Gap Conglomerate of White (1881, p. 56–57). Recent work by M. J. Bergin and others (oral commun., 1966) corroborates the suggestion by Willard (1946, p. 787) and Leonard (1953) that at the type locality, Griswold Gap, at the northern end of the Northern Anthracite basin, the basal conglomeratic sandstone of the Pottsville Formation was incorrectly assigned to the Pocono. Elsewhere in the Northern basin, however, the Griswold Gap is a part of the Pocono Formation and is of Early Mississippian age.

The Pocono Formation makes up interval A-B in the Broad Top basin and along the Allegheny Front in central Pennsylvania (pl. 9-B, sec. *h–h'*, interval A). The lower part of the formation, which contains many beds of siltstone, shale, and some impure coal, is less resistant than the upper part of the Pocono. An *Adiantites* flora occurs locally in the lower part of the formation in Blair, Centre, and Clearfield Counties (Read, 1955, p. 8, 15; Read and Mamay, 1964, pl. 1), suggesting that this shaly part of the Pocono is temporally equivalent to the Spechty Kopf Member of the Catskill Formation in the Anthracite basins. The upper part of the Pocono, the Burgoon Sandstone Member, is a hard resistant quartzite which can be readily identified in both the surface and subsurface parts of central Pennsylvania. A *Triphyllopteris* flora that occurs in the Burgoon at several places, including the type exposures along Burgoon Run, Blair County, (Swartz, 1965, p. 18), suggests a correlation of the member with the Mount Carbon Member in the Southern Anthracite basin.

In northeasternmost West Virginia (Morgan and Berkeley Counties), the deep Meadow Branch syncline contains a thick sequence of continental strata which compose the Pocono Group (pl. 9-B, sec. *g–g'*, interval A). In ascending order these strata are divided into the Rockwell Formation, Purslane Sandstone, and Hedges Shale (Read, 1955, p. 10–11). Previously, Stose and Swartz (1912, p. 14) had assigned the overlying Myers Shale and Pinkerton Sandstone to the Pocono, but Read has indicated that the age of post-Hedges formations is open to question and suggests that lithologic charac-

teristics favor correlation of the Myers with a younger Mississippian unit and the Pinkerton with basal Pennsylvanian.

The base of the Rockwell Formation was arbitrarily placed at the base of a pebbly quartzite in a sequence of red and gray strata at the top of the Catskill (Hampshire) Formation (Stose and Swartz, 1912). Dally (1956) has demonstrated that the Rockwell and Hampshire Formations intertongue and that at many places along the Allegheny Front in West Virginia the upper few feet of red beds of the Hampshire contain a marine fauna of Early Mississippian age. The Rockwell, which contains many beds of gray and red shale, siltstone, and some impure coal, as well as some resistant beds of quartzitic sandstone, is similar in composition and correlates with the sub-Burgoon part of the Pocono Formation of southern Pennsylvania (de Witt, 1969, p. 31). The Purslane Sandstone, which overlies the Rockwell Formation in Washington and eastern Allegany Counties, Md., as well as in northeastern West Virginia, is a resistant pebbly orthoquartzite. The soft coal-bearing beds of the overlying Hedges Shale contain a *Triphyllopteris* flora in the Meadow Branch syncline which demonstrates the Osage age of these soft strata (Read, 1955, p. 1). The Hedges Shale is restricted to the Meadow Branch syncline, whereas Purslane Sandstone is of wider extent. Although recognized in the eastern part of Allegany County, Md., the Purslane is not present along the Allegheny Front in the western part of the county.

In eastern West Virginia (parts of Greenbrier, Pocahontas, Randolph, Pendleton, and Tucker Counties), Dally (1956) has shown that the upper part of the red beds of the Hampshire Formation is of Early Mississippian age. The red beds of Mississippian age are overlain by Dally's Pocono Group throughout much of these five counties except in the vicinity of Beverly, Randolph County, where the strata in interval A-B are extremely thin and the Pocono is absent. There, carbonate rocks of Chester age are in contact with the Hampshire (pl. 9-B, sec. *i–i'*, interval A). In an area immediately to the west (parts of Barbour, Braxton, Lewis, and Webster Counties), Flowers (1956) demonstrated that Catskill red beds directly underlie Chester age beds of the Greenbrier Limestone. Thus, within a small area in eastern West Virginia interval A-B is extremely thin and is represented only by red beds. Most probably, the upper part of Flowers' red Catskill is the subsurface equivalent of Dally's Mississippian age Hampshire of the Allegheny Front. The absence of paleontologic control in the subsurface prevents confirmation of this supposition.

In Montgomery County, Va., and south along the Valley and Ridge province, clastic coal-bearing rocks of the Price Formation and the overlying red beds of the Maccrady Shale make up interval A-B (pl. 9-B, sec. *j–j'*, interval A, A-B, B). Locally in the vicinity of Saltville, Smyth County, Va., the Maccrady contains salt and gypsum which have been exploited commercially for many years. In Pulaski County, Glover (1953) has subdivided the Price Formation into a lower predominantly shaly unit, which Cooper (1963, p. 22) formally designated the *Parrott Formation*, and an overlying sandy unit, the restricted *Price Formation*. Cooper's *Parrott Formation* contains an Early Mississippian fauna and overlies the *Broad Ford Sandstone* of Reger (1926, p. 505, 520), which at its type locality in Smyth County contains a Late Devonian *Cyrtospirifer* fauna (Glover, 1953). Most probably, the *Parrott Formation* correlates in a general way with the upper part of the Hampshire Formation and with the *Marlinton Formation* of Dally (1956), both of which contain similar Early Mississippian fossils in parts of eastern West Virginia (Greenbrier, Pocahontas, Randolph, and Tucker Counties). The restricted *Price Formation*, which contains several coal beds in west-central Virginia, has a *Triphyllopteris* flora that suggests a general correlation with the Purslane Sandstone and Hedges Shale of the Meadow Branch syncline in northeastern West Virginia and with the Burgoon Member of the Pocono in south-central Pennsylvania.

At some places along the Allegheny Front south of Mercer County, W. Va., and in the Greendale syncline south of the front, the Price Formation overlies the Big Stone Gap Member of the Chattanooga Shale (pl. 9-B, sec. *k–k'*, interval A). The upper part of the Big Stone Gap Member contains a Kinderhook *Siphonodella* fauna that is identical to the fauna in the Sunbury Shale, whereas the lower part of the member contains a Late Devonian fauna (Roen and others, 1964, p. B47). The age of the Price Sandstone in southwestern Virginia is not well documented; however, fossils from the Maccrady Shale in Lee County indicate an early Osage age for the red beds there. Both the Price and the Maccrady intertongue with the greenish-gray Grainger Shale in Lee County and cannot be recognized in the section at Cumberland Gap in extreme western Lee County (pl. 3-A; pl. 9-B, sec. *i–i'*, intervals A, A-B, and B). A northeast-thinning wedge of cherty Fort Payne separates the Grainger from the carbonate rocks in interval C at Cumberland and Jellico Gaps (fig. 4).

UPPER BOUNDARY OF INTERVAL A-B

From the Anthracite basins of Pennsylvania south along the Valley and Ridge province to the Blacksburg-Pulaski synclinorium of Cooper (1964, p. 103), in southwestern Virginia, the sandy rocks of interval A-B are overlain by red beds of Meramec and Chester age. These rocks are the Mauch Chunk Formation in Penn-

sylvania, the Myers Shale in the Meadow Branch syncline in West Virginia, and the *Stroubles Formation* in Virginia. Commonly, the contact between sandstone and red shale is sharp, but at places the rock types intertongue and the boundary between intervals must be arbitrarily delineated. From Altoona, Blair County, Pa., south along the Allegheny Front and in the Greendale syncline in southwestern Virginia, calcareous rocks of Meramec and Chester age overlie the clastic strata in interval A-B. Wells (1950, p. 918) has shown that from southern West Virginia (Mercer County) north along the front a hiatus of increasing magnitude separates the rocks of interval A-B from the calcareous strata above. Limestone of early Meramec age caps the Maccrady Formation in Tazewell County, Va., and adjacent Mercer County, W. Va., whereas farther north in West Virginia in parts of Randolph County, early Chester limestone rests on the clastic strata of interval A-B. The magnitude of the hiatus is less in south-central Pennsylvania, where the very calcareous sandstone or very sandy limestone of the Loyalhanna is paraconformable on the Burgoon Member of the Pocono Formation.

THICKNESS TRENDS

Three areas of thick strata dominate the isopach of interval A-B (pls. 3-A, 4-A). The largest area, in which the rocks are more than 2,900 feet thick, is centered in the Southern Anthracite basin in eastern Pennsylvania and marks the site of the Pocono delta. A second lies in the Broad Top basin in south-central Pennsylvania and is probably also a part of the Pocono delta. A third area, in which the rocks are more than 1,400 feet thick, is located in the Blacksburg-Pulaski synclinorium of southwestern Virginia and is the site of the Price delta. The rocks of interval A-B are exceptionally thin or are locally absent in parts of eastern West Virginia (parts of Barbour, Braxton, Lewis, Randolph, and Webster Counties). A smaller spot of thin strata exists in central Allegany County, Md. In general the rocks of ir terval A-B are thickest in the Valley and Ridge province and thin westward into the Appalachian Plateaus province. The rate of thinning is most marked in southern West Virginia and southwestern Virginia adjacent to the Allegheny Front.

LITHOFACIES TRENDS

Commonly areas containing the most sandstone correspond to the areas of thick interval A-B. A tongue of quartzose clastics projects west and southwest from eastern Pennsylvania across western Maryland and into adjacent northern West Virginia (pls. 3-B, 4-B). The more rapid westerly transition from sandstone to shale in southwestern Virginia and adjacent West Virginia than in Pennsylvania suggests that the Price delta was considerably smaller than the Pocono delta.

A wedge of calcareous cherty rock underlies a small part of southeastern Kentucky. These strata are the distal fringe of a large mass of Fort Payne Formation that lies west of the main part of the Appalachian basin.

SOURCES AND ENVIRONMENTS

Clearly, the source for the clastic rocks of interval A-B lay east of the Appalachian basin. The rocks thicken and become coarser grained to the east (pls. 3-A, 3-B). Pelletier (1958, p. 1041, 1061) suggested that a terrain of sedimentary and low-rank metamorphic rocks lying more than 40 miles east of the present Atlantic shoreline was the source for the detritus that made up the Pocono delta of eastern and central Pennsylvania. Much of the strata of interval A-B was deposited in nonmarine environment as demonstrated by the absence of marine fossils, the presence of continental floras, and the abundance of local coaly beds. The coaly rocks of the Pocono in Pennsylvania probably accumulated in swamps on the coastal part of a large delta complex (Pelletier, 1958, p. 1057). A similar environment seems plausible for the coaly part of the Price Formation in Virginia. The presence of marine fossils in the upper part of the Hampshire of West Virginia and in the *Parrott Formation* and the lower part of the Price Formation in parts of Virginia shows that some of these strata were deposited in a shallow neritic marine environment. Certainly, the Big Stone Gap Member of the Chattanooga Formation and the associated Grainger Shale are marine deposits. From Pennsylvania southwest along the axis of the geosyncline the rocks of interval A-B show a gradation from coarse-grained clastics of a high-energy continental environment to marine shale typical of a tranquil euxinic environment.

The paleotectonic implications of the thickness and lithofacies trends for interval A-B are discussed in conjunction with the data from interval B.

INTERVAL B

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Throughout most of the outcrop in Ohio two formations, the Cuyahoga below and the Logan above, make up interval B (pl. 9-B, secs. g–g', k–k', interval B). Both formations have been subdivided into members. Facies relations within the Cuyahoga are complex because coarse-grained pebbly sandstone intertongues abruptly with shale (Hyde, 1915; Holden, 1942; Root and others, 1961, p. 7–20). In contrast the members of the Logan are relatively extensive and are of uniform thickness.

Because two thin but characteristic beds of conglomerate, the *Berne* and *Allensville Members*, are rarely found in drill cuttings, the Logan cannot be readily subdivided in the subsurface. Commonly, the drillers' *Big Injun sand* fills most of interval B in the Ohio subsurface. The *Big Injun sand* is a thick sequence of sandstone and siltstone which extends through much of eastern Ohio, western Pennsylvania, eastern Kentucky, and West Virginia. In Ohio the coarse-grained pebbly Black Hand Member of the Cuyahoga Formation and the sandstones of the Logan Formation make up the *Big Injun sand*. Thickness variations in the *Big Injun* of Ohio are commonly the result of abrupt changes in thickness of the Black Hand. In the northern half of Ohio and adjacent Pennsylvania coarse-grained sandstones of Early Pennsylvanian age are in contact with and fill channels scoured into the Osage age sandstones of interval B. Because the drillers are unable to differentiate between the sandstones from the two systems in well cuttings, commonly, in this area the *Big Injun* is shown to contain considerably more strata than just the rocks of interval B.

Correlation of units in interval B in northwestern Pennsylvania is not well established. The Cuyahoga and part of the Shenango Sandstone in Pennsylvania appear to be equivalent to similar rocks in northeastern Ohio (pl. 15, cols. 8, 9). However, the age of the upper part of the Shenango in parts of northwestern Pennsylvania (parts of Venango and Clarion Counties) is equivocal. Generally, all the Shenango has been assigned to the Osage (Weller and others, 1948, chart 5). However, D. A. Busch (in Dickey and others, 1943, p. 149–154) presented faunal data indicating that the upper part of the Shenango may be as young as Chester. The area has not been intensively studied since Busch introduced these intriguing data, and the conflicting age assignments of the upper part of the Shenango merit a thorough reexamination. The relation of the Shenango Sandstone of northwestern Pennsylvania to the Burgoon of western Pennsylvania is not clear. Both sandstones grade into the thick *Big Injun sand* in the subsurface, but the equivalence or nonequivalence of the two sandstones remains unsettled.

The *2d gas sand* of western Pennsylvania, which lies at the base of interval B (pl. 15), can be traced east into the thick sandstones of interval A-B; west into Ohio, where several local names are applied to the discontinuous sheet of sand; and south into West Virginia, where the *2d gas sand* appears to correlate with the lower part of the *Weir sand*.

In parts of West Virginia the upper half of interval B is filled by the coarse-grained *Big Injun sand* which appears to be equivalent to some of the Pocono or Price Formations to the east (pl. 9-B, sec. *k*–*k'*, interval B). However, in a segment of central West Virginia centering in Calhoun and Gilmer Counties (Flowers, 1956, p. 11) the strata of interval B are mainly shale and siltstone. In this area the drillers' *Big Injun sand* is a calcareous sandstone of Meramec age in the basal part of interval C rather than an Osage-age sandstone in interval B.

In much of the southern half of West Virginia and in adjacent eastern Kentucky the drillers recognize one or more *Weir sands* in interval B. Correlation of these sands is somewhat of an enigma. Most probably the lower or *2d Weir sand* is roughly equivalent to the *Hamden sand* of Ohio and the *2d gas sand* of Pennsylvania. The drillers record a *Red Injun sand*, the subsurface part of the Maccrady Shale, at many places in both southern West Virginia and eastern Kentucky. The red stratum is a widespread key bed that commonly marks the upper limit of interval B in this part of the basin. Similarly, a thin bed of red shale is reported locally at or near the top of interval B in part of northern West Virginia and in Monroe and Washington Counties, Ohio.

In the Valley and Ridge province of eastern Tennessee, interval B consists of the Grainger Formation, a thick shale and sandstone unit. Elsewhere in the southern part of the basin, interval B contains the Fort Payne Chert which generally ranges from bedded chert to a cherty limestone. However, in north-central Tennessee, Klepser (1937) described the Fort Payne Formation as consisting of silty shale, siltstone, and crinoidal limestone, which he named the "Greasy Creek" facies and applied the name "Short Mountain" facies to the predominantly chert and carbonate sequence to the south. Butts (1926, p. 164) described as much as 60 feet of dark shale in the lower part of the Fort Payne Chert in northern Alabama, and Charles Cressler (written commun., 1967) stated that the Fort Payne Chert in Polk County, Ga., is very impure and contains large quantities of silt and clay.

UPPER BOUNDARY OF INTERVAL B

Throughout most of the western part of the Appalachian basin carbonate rocks of interval C overlie either the red beds of the Maccrady Shale or the sandstone of the Burgoon–Big Injun–Logan sequence in areas where the Maccrady was not deposited. The hiatus between the carbonate rocks and the underlying clastics is least in southwestern Virginia and southern West Virginia, where a thin sequence of Little Valley Limestone or Hillsdale Limestone lies on the Maccrady Shale. The hiatus is greatest in east-central Ohio, where the Logan Formation is overlain by limestone of early Chester age (Scatterday, 1963). Pre-Pennsylva-

nian erosion has removed much of the Mississippian strata in northeastern Ohio and adjacent Pennsylvania, and there basal Pennsylvanian sandstone rests on the beveled edge of the clastic rocks of interval B (pl. 9-B, sec. *k*–*k'*, interval B).

In the southern part of the Appalachian basin carbonate rocks of interval C generally conformably overlie the Fort Payne Chert and the Grainger Formation. A hiatus above interval B is present in Polk County, Ga., and in parts of northeast Alabama and northeast Mississippi, where rocks of early Chester age rest on the Fort Payne Chert (pl. 15). The upper boundary of carbonate rocks of interval B, where overlain by younger carbonate rocks, is generally selected to be where the chert content decreases markedly.

THICKNESS TRENDS

A linear body of interval B, which is more than 500 feet thick, extends from north-central Ohio (Medina and Wayne Counties) south across Ohio and West Virginia to the southern boundary of West Virginia (McDowell County), where the rocks of interval B merge into the rocks of interval A-B (pl. 4-A). Interval B is generally thinner to the east across eastern Ohio and central West Virginia. However, the thickness of the interval at the north end of this belt is not due to original deposition but has been produced largely by both pre-Chester and pre-Pennsylvanian erosion which cut deeply into interval B north of the Ohio River. Also, pre-Pleistocene erosion has removed a considerable amount of Mississippian rocks in northern Ohio; consequently, thicknesses do not indicate the original amount of interval B deposited in this area.

Western Pennsylvania is covered by an extensive sheet of rocks of Osage age, which is thickest adjacent to the boundary with combined interval A-B. These strata are the western part of the Pocono delta. A smaller patch of thick strata in southeastern West Virginia and eastern Tennessee appears to be a digitation of the Price delta. (See p. 24.) An area of thin interval A-B in north-central West Virginia (Braxton and Lewis Counties) is associated with the area of very thin Lower Mississippian (interval A-B) rocks to the east in Randolph, Upshur, and Webster Counties.

The Fort Payne Chert in east-central Tennessee, Georgia, Alabama, and Mississippi is a sheet of fairly uniform thickness, generally 100–200 feet. It thickens to more than 300 feet to the south and east in parts of the Valley and Ridge province of northeastern Tennessee, northwestern Georgia, and east-central Alabama. The thickest part of interval B in the southern part of the Appalachian basin is in the Valley and Ridge province of eastern Tennessee, where the Grainger Formation is as much as 1,200 feet thick.

LITHOFACIES TRENDS

A west-thinning relatively shale-free wedge of sandstone covers much of western Pennsylvania (pl. 4-B). It is thickest and contains the greatest amount of coarse-grained sand adjacent to the boundary with interval A-B in southern Pennsylvania (Cambria and Somerset Counties). In northwestern Pennsylvania a small tongue of conglomeratic sandstone lies in McKean County adjacent to the New York State line.

In parts of central Ohio the Black Hand Member of the Cuyahoga, a thick mass of sandy rock, is dominant, as shown on the lithofacies map. Elsewhere, a thick sequence of shale and siltstone in the Cuyahoga accentuates the shale pattern and masks the sandstone of the overlying Logan Formation (pl. 9-B, sec. *g*–*g'*, interval B). In the subsurface only the thickest sequences of *Big Injun* or *Weir sands* are indicated by sandstone patterns on the interval B lithofacies map (pl. 4-B) because in general the rocks of interval B contain sufficient amounts of siltstone intercalated with the sandstone to skew the lithofacies patterns to a shale lithofacies symbol. The small area of sandy shale in southern West Virginia (Mercer County) appears to be the fringe of a thick mass of sandstone whose center of accumulation lay to the southeast in Montgomery and Pulaski Counties, Va.

A sheet of carbonate rock of fairly uniform thickness, the Fort Payne Formation, extends from east-central Tennessee southwest to east-central Mississippi. A small lobate tongue of shale with some sandstone extends southeast into the carbonate sequence in north-central Tennessee. In eastern Tennessee, where interval B rocks of the Grainger Formation are preserved in the Valley and Ridge province, the lithologic composition ranges from shale to shaly sandstone.

SOURCES AND ENVIRONMENTS

The rocks which make up interval B in northwestern Pennsylvania and probably some of the strata in the adjacent part of interval A-B had a northern source (pl. 10, fig. 2) in eastern Canada that was roughly similar in location to the source of the Berea Sandstone and the Knapp Formation. The Shenango Sandstone is thickest and coarsest grained in the north. It thins and becomes finer grained to the south. The Burgoon Member of the Pocono Formation in central Pennsylvania (Clearfield County) shows a similar southward decrease of grain size and thinning (V. E. Sheps, oral commun., 1965). Evidence of a northern source for the Cuyahoga and Logan Formations of Ohio is less definite. Ver Steeg (1947, p. 703–727) concluded that the Black Hand Member of the Cuyahoga probably had a north-

ern source, although depositional structures indicate local transport of detritus from the east or southeast.

The Pocono Formation of eastern and central Pennsylvania and the Price Sandstone of Virginia and West Virginia were clearly derived from an eastern source. These strata compose most of interval A-B as well as large areas of both intervals A and B. The rocks are coarsest grained and thickest in the east. Beds of coal and coaly shale intercalated in the finer grained strata, together with primary sedimentary structures in these beds, indicate continental accumulation. The coarser grained rocks, conglomerate and coarse-grained pebbly sandstone, thin and grade laterally into finer grained strata containing marine fossils in the western part of the basin. Although the zone of transition from continental to marine deposition (pl. 11, fig. 2) is in the subsurface, well cuttings indicate that it lies not far west of the boundary between interval A-B and interval B (Pelletier, 1958, p. 1055). The rocks of these two intervals are part of a large delta complex whose center lay in southeastern Pennsylvania and whose distal extremities spread across parts of eastern Kentucky, southeastern Ohio, western Virginia, and northeastern Tennessee. The source area of the rocks that make up the northern part of this delta complex was a mountainous area (pl. 12, fig. 2) that lay east of a fall line across northwestern New Jersey, southeastern Pennsylvania, and eastern Maryland (Pelletier, 1958, p. 1055). By analogy, the rocks of the Price delta, which make up the south part of the complex, probably had a source area along the Virginia-North Carolina border south of Richmond. The small area of predominantly silty shale and siltstone in interval B in north-central Tennessee (pl. 4-B) may be a distal remnant of the delta complex although its source could have been a slightly positive area to the southwest in the vicinity of the Nashville dome (fig. 3).

Pelletier (1958, p. 1057) indicated that fluviatile deposition was dominant on a broad coastal plain west of the fall line on which coal swamps formed in favorable localities. In the shallow fringing sea high-energy waves and currents winnowed and sorted sand, silt, and mud. The large marine fauna in some parts of the Cuyahoga and Logan of Ohio and the equivalent strata in eastern Kentucky demonstrate that these rocks accumulated in a marine environment locally favorable to a host of invertebrates. The large volume of medium- to coarse-grained sand and the presence of lenses of pebbles in the *Big Injun sand* of western Pennsylvania and adjacent northern West Virginia attest to the efficiency of sorting and transport by the nearshore high-energy waves and currents. The presence of coal flora and abundant marine fauna suggests a moderate to warm benign climate for the basin.

In contrast to the high-energy marine environment of western Pennsylvania, the well-laminated siltstone and shale, which make up much of interval B in eastern Kentucky and north-central and eastern Tennessee, appear to have accumulated in a lower energy tranquil environment. The increasing amount of carbonate and chert in the southwestern part of the basin in east-central Tennessee, northwestern Georgia, northern Alabama, and northeastern Mississippi suggests that these strata accumulated in a deeper water environment some distance from the shore of the expanding delta complex that dominated the eastern part of the basin during the accumulation of interval B.

Locally, salt, anhydrite, gypsum, and dolomite were deposited in a small trough around Saltville in parts of Smyth and Washington Counties, Va. The hypersaline environment in which these evaporites accumulated was developed in a local trough as a part of Maccrady deposition. Cooper (1966, p. 23, 29) has indicated that folding and faulting during accumulation of the Maccrady were major factors in the localization of the Saltville deposit. Local tectonic control and evaporation in a small basin rather than a general arid climate is indicated by the presence of abundant flora and fauna in lateral equivalents of the Maccrady.

PALEOTECTONIC IMPLICATIONS

The great volume of dominantly coarse grained clastic rocks that makes up much of interval A-B and interval B in the northern part of the region indicates that source areas to the east and north of the Appalachian basin were actively uplifted and eroded during much of Early Mississippian time. Progressive offlap of terrestrial strata over marine rocks and the restriction of the sea to the western part of the basin suggest that the sources supplied detritus more rapidly than the subsiding basin could readily accommodate them. The widespread sheet of Burgoon–*Big Injun*–Black Hand Sandstone indicates the extent to which coarse-grained clastics filled the northeastern part of the basin.

The presence of a sheet of fine-grained clastic, which occurs at the top of intervals B and A-B in the central part of the basin and occupies entire interval B in the south-central part of the basin, suggests a slackening of the rate at which coarse clastics were being transported to these parts of the Price-Pocono delta complex, the southern extremities of which extend into eastern Tennessee. The small area of fine-grained clastics in north-central Tennessee may be part of the delta complex; however, it could suggest uplift to the southwest in the vicinity of the Nashville dome.

Accumulation of coal and shale in the Hedges Shale

of the Meadow Branch syncline in northeastern West Virginia also suggests that the south part of the eastern source area had been considerably degraded by late Osage time and was no longer capable of producing much coarse-grained detritus. In contrast the northern part of this eastern source area continued to supply coarse-grained clastic debris well into Meramec time.

The thin sequence of Lower Mississippian rocks in north-central West Virginia (Barbour, Braxton, Lewis, Randolph, and Webster Counties) (pl. 9-B, sec. *i–i'*, interval B) marks a shoal area of little or no deposition throughout most of the Early Mississippian. Although surrounded by gray Pocono Sandstone, this area contains only a few tens of feet of Hampshire red beds with an Early Mississippian marine fauna (Flowers, 1956, p. 8; Dally, 1956, p. 195). Uplift was apparently sufficient to prevent the accumulation of younger Early and Middle Mississippian strata. Possibly, this area was uplifted at about the same time that the Spechty Kopf rocks were locally folded in the Southern Anthracite basin in Pennsylvania and at about the same time that the Mississippian sea retreated from much of the Ohio bay in the western part of the Appalachian basin (Pepper and others, 1954, pl. 13E).

A thin sheet of Lower Mississippian rock covers part of western Allegany County, Md. Although Pocono Sandstone is present there, it is only about one-third as thick as in surrounding areas. Possibly, western Allegany County was an area of some uplift during the Early Mississippian. The available data are not definitive.

In the southern part of the basin the widespread sheet of carbonate and chert indicates the presence of a deeper, better aerated sea than had existed during interval A.

Silt and clay in the Fort Payne Chert in parts of Mississippi, Alabama, and Georgia suggest the possiblity of uplift along the southern margin of the sea during interval B.

The landmass that had extended from Georgia westward into Texas during interval A had probably become submerged along its northern and western margins and was now restricted to east Texas, Louisiana, and the southern parts of Mississippi and Alabama (pl. 12, fig. 2).

INTERVAL C

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

The Little Valley Limestone, which is the oldest rock of the basal part of interval C (pl. 15, col. 24), is present in the Greendale syncline of southwestern Virginia and along the adjacent part of the Allegheny Front in Virginia and West Virginia (pl. 9-C, secs. *f–f'*, *g–g'*, *j–j'*, interval C). The limestone, which contains a *Warsaw fauna* (Averitt, 1941, p. 17–21), is as much as 500 feet thick in Scott County, Va., where salt and gypsum are locally present in the basal part (Cooper, 1966, p. 17). The Little Valley thins to the southwest along the Allegheny Front (pl. 9-C, sec. *f–f'*, interval C). Price (Price and others, 1938, p. 220) reported a similar fauna from his *Maccrady* Series in southern West Virginia (Mercer and Monroe Counties). There Price's *Maccrady* contains a large amount of yellow calcareous mudrock similar to some beds in the Little Valley of the Greendale syncline. These data suggest that some of the *Maccrady* Series of Mercer and Monroe Counties may well be a Little Valley equivalent.

The dark-gray commonly cherty Hillsdale Limestone overlies and overlaps the Little Valley Limestone. The Hillsdale contains a *Lithostrotionella* fauna (Wells, 1950, p. 901) which suggests correlation with the St. Louis Limestone of the Mississippi Valley.

The abundantly fossiliferous, commonly oolitic *Denmar Formation* overlies the Hillsdale Limestone in the Greendale syncline and northeast along the Allegheny Front (fig. 7) from Cumberland Gap to northern West Virginia (southern Randolph County) (Butts, 1940b, p. 366–374; Wells, 1950, p. 918). Throughout its extent in eastern Kentucky, Virginia, and southern West Virginia, the *Denmar* contains the characteristic stem plates of the crinoid *Platycrinites huntsvillae* which suggests a correlation with the Ste. Genevieve Limestone of the Mississippi Valley section. In the subsurface of eastern Kentucky, southwestern Virginia, and much of West Virginia, strata equivalent to the Hillsdale and *Denmar* make up the lower part of the well drillers' Big Lime, the lower part of the Greenbrier Limestone of Virginia, or the Greenbrier Group of West Virginia. Locally, small amounts of anhydrite are intercalated in the basal part of the Greenbrier, and some oolite zones are productive of gas and oil (Youse, 1964, p. 465–486). In part of central West Virginia in and near Calhoun County, the basal part of the Greenbrier is a calcareous sandstone which yields oil and gas. Although well drillers included these petroliferous sandy rocks in their *Big Injun sand*, sample studies clearly indicate that the sandy strata are a facies of the Greenbrier Limestone (Martens, 1945; Flowers, 1956, p. 11). In this area the drillers' *Big Injun* is excessively thick, and the overlying *Big lime* is correspondingly thinner than normal.

In southwestern Pennsylvania, western Maryland, southeastern Ohio, and northern West Virginia, the Loyalhanna Limestone constitutes the strata of interval C (pl. 9-C, secs. *f–f'*, *h–h'*, interval C). At some places the Loyalhanna is a calcareous sandstone;

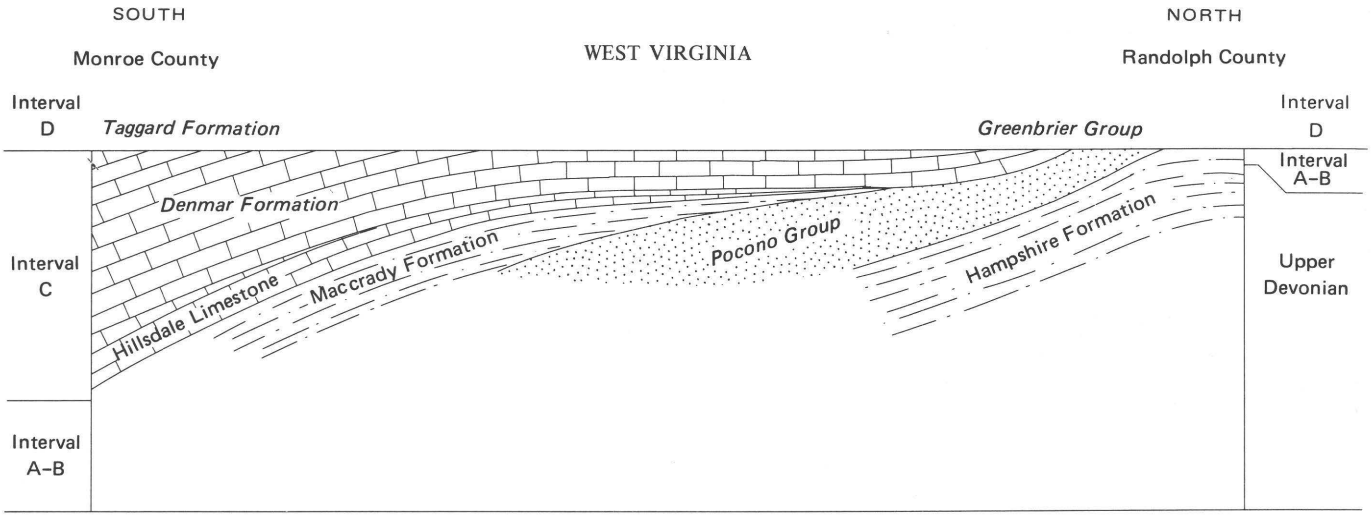


FIGURE 7. — Overlap of Meramec carbonate units (brick pattern) northward along the Allegheny Front in eastern West Virginia. No vertical scale. Adapted from data by Wells (1950) and Dally (1956).

whereas at other places it is a sandy limestone. In parts of southeastern Ohio (parts of Monroe and Washington Counties), the drillers' *Lime sand*, which is locally petroliferous, appears to be an equivalent of the Loyalhanna of Pennsylvania. The Loyalhanna Limestone is of Meramec age and is an equivalent of the Ste. Genevieve Limestone (Butts, 1945, p. 13–14; Adams, 1964).

Adams (1964) stated that the Trough Creek Limestone, which consists of as much as 25 feet of limestone and intercalated reddish-gray shale in the Broad Top basin, is an eastern nonsandy part of the Loyalhanna and is of Meramec age. Previously, the Trough Creek was considered the basal member of the Mauch Chunk Formation and of Chester age (Butts, 1945, p. 14). The absence of fossils above the Trough Creek precludes determining how much of the overlying 1,000 feet of Mauch Chunk may also be of Meramec age. To the south in the Meadow Branch syncline, Read (1955, p. 11) correlated the 800–900 feet of red Myers Shale with the Mauch Chunk. Again the absence of fossils precludes determining how much, if any, of the Myers Shale belongs in interval C. Presumably, a small thickness of the basal Myers is equivalent to the Meramec age Mauch Chunk of the Broad Top basin.

In parts of the Anthracite region the basal 300–500 feet of the Mauch Chunk Formation contains intercalations of light-gray conglomeratic sandstones similar in lithology to the underlying Pocono Formation. Because this sequence of beds appears to be transitional with the Pocono Formation and occurs closely above beds containing an Osage age *Triphyllopteris* flora (Read and Mamay, 1964, p. K5), we tentatively assign them a Meramec age and place them in interval C.

D. A. Busch (in Dickey and others, 1943, p. 149–154) suggested that the upper part of the Shenango Sandstone of northwestern Pennsylvania (Venango and Clarion Counties) is younger than Osage and may be of Meramec or Chester age. Some of the Shenango is shown on the lithofacies map of interval C (pl. 5-B) although its relationships to the Loyalhanna are unresolved.

In east-central Ohio the basal part of the Maxville Limestone is of St. Louis age and is the only part of the formation that falls in interval C (Scatterday, 1963); the Ste. Genevieve equivalent is absent owing to nondeposition. Scatterday has shown that most of the Maxville is of Chester age and belongs in interval D.

Throughout most of the southern part of the Appalachian basin in eastern Tennessee, northeastern Alabama, and northwestern Georgia the formations that make up interval C are, from oldest to youngest: Warsaw Limestone, St. Louis Limestone, and *Ste. Genevieve Limestone*. In eastern Tennessee, Subdivisions I, II, and III of the Monteagle Limestone (Stearns, 1963) are equivalent to the *Ste. Genevieve Limestone* of uppermost interval C. In parts of northwestern Georgia and northeastern Alabama interval C is represented by the lower part of the Floyd Shale, and in northern Alabama and Mississippi interval C is represented by the Tuscumbia Limestone which is of early and middle Meramec age. The Tuscumbia is separated from Chester age rocks or interval D by an unconformity (Welch, 1959).

The Warsaw and St. Louis Limestones are generally dark- to light-gray thick-bedded coarsely crystalline crinoidal cherty limestones and are lithologically very similar. Butts (1926, p. 175) reported that the St. Louis

Limestone in Alabama and Tennessee contains layers of fine-grained dark limestone unlike any rock in the underlying Warsaw. Results of this study showed the basal part of the Warsaw to be sandy and shaly in some localities.

Smith (1967, p. 40) described the Tuscumbia Limestone as a medium- to thick-bedded, crinoidal, cherty limestone which in western Alabama is slighty asphaltic and rarely contains scattered thin partings of pale-green calcareous shale. He further stated that although two divisions, Warsaw and St. Louis Limestone, have been suggested (Butts, 1926) for the Tuscumbia, megascopically there is little lithologic difference between the upper and lower parts, and most workers see little reason for the divisions. Thin sections do reveal subtle difference between the strata in the two parts of the formation.

The *Ste. Genevieve Limestone* is lighter gray and although in places is quite cherty it is generally less so than the underlying Tuscumbia or St. Louis Limestone. It is porous, fossil fragmental, and generally oolitic, and it contains some interbedded nonoolitic limestone. The lower three subdivisions of the Monteagle Limestone, which are correlated with the *Ste. Genevieve*, consist of light-gray fine- to medium-grained fossil-fragmental oolitic cherty limestone containing minor amounts of slit and shale.

In Polk County, Ga., interval C was not recognized, and the Floyd Shale of Chester age rests on the Fort Payne Chert (C. Cressler, written commun., 1967).

UPPER BOUNDARY OF INTERVAL C

In outcrop, from eastern West Virginia (Pocahontas County) south along the Allegheny Front and along Pine Mountain to Jellico Gap, Tenn., the top of interval C is the base of the *Taggard Formation* of Wells (1950, p. 916), which is also the Taggard Red Member of the Greenbrier Formation (Wilpolt and Marden, 1959, p. 594). The Taggard ranges in thickness from a few inches to 40 feet and is a sequence of red and yellow shale and intercalated silicious reddish-gray limestone. The red beds can be seen in outcrop; however, because they cannot be identified with certainty in the subsurface or may be missing in much of southwestern West Virginia and adjacent eastern Kentucky, the top of interval C is drawn more or less arbitrarily in the subsurface at the main change from light-gray or white limestone below to light-brown limestone above. Locally, in the subsurface of eastern Kentucky, a bed of limestone conglomerate is closely associated with the zone of color change. If this conglomeratic bed is the equivalent of the *Bryantsville Breccia Bed* of the Greenbrier in outcrops along the east side of the Cincinnati arch (McFarlan and Walker, 1956, p. 10), the boundary based on color change closely approximates the Meramec-Chester boundary as determined by paleontologic and other sedimentary criteria.

In northern West Virginia, western Maryland, and southwestern Pennsylvania, the upper boundary of interval C is customarily drawn at the top of the sandy Loyalhanna Limestone (pl. 9-C, secs. *f–f'*, *h–h'*, *j–j'*, interval C). The change from sandy to nonsandy limestone is abrupt, and at some places a tongue of red Mauch Chunk Shale separates the Loyalhanna of Meramec age from the nonsandy *Deer Valley Limestone Member* of the Mauch Chunk of Chester age. In southeastern Ohio, the top of interval C appears to be the top of the drillers' *Lime sand*, which is overlain by a sequence of nonsandy limestone at some places. In east-central Ohio an unconformity at the top of the St. Louis part of the Maxville Limestone marks the top of interval C. Locally, the St. Louis part has been removed by an episode of Late Meramec erosion, and the Chester part of the Maxville rests upon the upper part of the Logan Formation (Osage) (Scatterday, 1963). The Ste. Genevieve equivalent, which is missing in east-central Ohio, is present in southernmost Ohio (Lawrence County) and in the adjacent part of eastern Kentucky. There, the top of the Meramec limestone sequence is an oolitic sandy limestone that underlies the *Bryantsville Breccia Bed* (McFarlan and Walker, 1956, p. 35).

Because widespread key beds are not present near the top of interval C in much of the subsurface Appalachian basin, delimiting the top of the interval is necessarily subjective and arbitrary. In much of the subsurface of central and east-central West Virginia and parts of eastern Kentucky, deposition appears to have been continuous from Meramec to Chester. In this part of the basin, subdivision of the Meramec-Chester carbonate sequence is accomplished in part by splitting the stratigraphic sequence at the base of the main mass of brownish-gray or brown limestone, close to the boundary between the *Denmar Formation* (Ste. Genevieve) and the younger *Pickaway Limestone* (Girkin), and in part by projecting intervals from adjacent outcrops where lithologic and paleontologic criteria permit delineation of the Meramec-Chester boundary.

In Mississippi and Alabama where the Pride Mountain Formation or Floyd Shale of Chester age rests on the Tuscumbia Limestone, the top of interval C is placed at the top of the massive gray limestone which forms the Tuscumbia and provides a distinct upper boundary. However, throughout the rest of the southern part of the Appalachian basin the upper boundary is extremely difficult to select. Where the boundary falls within the Monteagle Limestone or between the *Ste. Genevieve Limestone* below and the Girkin Formation above (pl. 15), it is placed at the top of the lower-

most oolitic limestone. Where no oolite is present or where the boundary falls within the Floyd Shale, its location is arbitrary.

THICKNESS TRENDS

The isopach (pl. 5-A) shows two areas of thick strata in the Valley and Ridge province that grade west into a broad sheet of relatively thin rocks underlying the Appalachian Plateaus province. The greatest thickness of interval C, more than 2,600 feet, occupies the Greendale syncline and impinges on the Allegheny Front in Mercer County, W. Va.

From the southern end of the Greendale syncline, where interval C is nearly 900 feet thick, southwestward in the Valley and Ridge province rocks of interval C thin progressively to zero in east-central Alabama. The absence of interval C rocks in this area is probably due to pre-Chester erosion, but the area may also be very near the limit of deposition.

Interval C may be as much as 500 feet thick in the Southern Anthracite basin; however, the thickness is suppositional because the data are much less conclusive there than in the Greendale syncline. Throughout the Appalachian Plateaus province the limestones of interval C are relatively thin and rarely exceed a thickness of 150 feet except along the Allegheny Front. The digitate pattern of the zero line in western Pennsylvania and eastern Ohio is in part depositional but is more the result of pre-Pennsylvanian erosion in Pennsylvania and a combination of pre-Chester and pre-Pennsylvanian erosion in Ohio. The zero area in the vicinity of Randolph County, W. Va., appears to be the result of nondeposition during Meramec time rather than post-Meramec erosion (pl. 10, fig. 3).

In the southern part of the Appalachian Plateaus province interval C reaches a maximum thickness of about 550 feet in northeastern Tennessee. The zero line along the eastern highland rim represents recent erosion; whereas the zero line in northeast Mississippi is probably a result of pre-Chester erosion (Welch, 1959).

LITHOFACIES TRENDS

Shaly and sandy clastic strata cover much of the Anthracite basins, the Broad Top basin of Pennsylvania, and the Meadow Branch syncline of West Virginia. Scant data indicate that the greatest quantity of coarse-grained clastics is concentrated in Carbon and Schuylkill Counties, Pa., in the eastern and middle parts of the Southern Anthracite basin.

A broad tongue of calcareous sandstone and sandy limestone, which extends from central Pennsylvania west to the Ohio River in south-central Ohio, dominates the north part of the Appalachian basin (pl. 5-B; pl. 9-C, sec. h–h', interval C). Crossbedded calcareous sandstone makes up interval C in south-central Pennsylvania, whereas the equivalent strata are predominantly sandy oolitic limestone in southern Ohio and the adjacent part of northeastern Kentucky. Belts of purer carbonate strata lie northwest and south of the tongue of sandy rock. Within the blanket of limestone in eastern Kentucky, southwestern Virginia, and West Virginia are many zones of porous oolite, some of which are productive of oil or gas. The long axes of many of these oolitic zones are roughly parallel to the axis of the basin, and the zones appear to thin westward updip against less porous and less permeable strata along a shoreline rimming the eastern edge of the Cincinnati arch, which was emergent during the early Meramec (Youse, 1964, p. 484).

A linear body of argillaceous and sandy carbonate rock covers the northern part of the Greendale syncline, Mississippian outliers to the northeast in the Valley and Ridge province, and the edge of the Allegheny Front in southeastern West Virginia (Mercer and Monroe Counties). These rocks contain the largest amount of quartzose clastics in Montgomery and Pulaski Counties, Va., and grade into a thick sequence of dominantly carbonate rocks south along the axis of the Greendale syncline.

Southwest of the Greendale syncline a narrow strip of sandy carbonate rock runs northwesterly through the predominantly carbonate sequence of interval C in that part of the Appalachian Plateaus, and shale and sandstone are present in the lower part of interval C in east-central Tennessee. In southeastern Tennessee and northwestern Georgia shaly and sandy carbonate rocks are present in outliers in the southern Valley and Ridge province, whereas throughout the remainder of the southern part of the Appalachian basin interval C is represented by a sheet of relatively pure carbonate rock locally containing some chert.

SOURCES AND ENVIRONMENTS

The sources of the carbonate rocks of interval C lie within the basin of accumulation. Beds of fragmented fossils, oolite, and pellitoid limestone are indicative of high-energy depositional environments common to the shallow neritic and sublittoral zones. In contrast, beds of calcilutite, which are commonly intercalated with the calcarenite, formed from fine carbonate muds which were winnowed from the high-energy environments and which accumulated in the lower energy environments either in deeper water, shallow protected bays, and lagoons, or on mud flats. Local beds of anhydrite in the basal part of the interval, exclusive of the Saltville area of southwest Virginia, probably accumulated in a hypersaline environment in lagoons associated with

offshore bars along the edge of the shallow Meramec Sea.

The Canadian Shield north and northwest of the Appalachian basin appears to have been the source of most of the quartz sand in the Loyalhanna Limestone of southwestern Pennsylvania and contiguous parts of Maryland, Ohio, and West Virginia (Rittenhouse, 1949, p. 1722; Adams, 1964). Apparently, the pattern of sand-transporting currents in the Appalachian basin was more complex than heretofore realized. Adams showed that most of the quartz sand in the Loyalhanna was introduced into the basin of deposition from the north and spread to the east and northeast along the axis of the basin from western Pennsylvania into central Pennsylvania as well as being spread to the southwest into Ohio and West Virginia. Small amounts of quartz sand and silt were derived from an eastern source and were deposited in a relatively high-energy environment in north-central West Virginia adjacent to the zero-sediment area of Randolph County. Sandstone and shale in the lower part of the Mauch Chunk Formation in the Anthracite basins of Pennsylvania and in the *Stroubles Formation* of southwestern Virginia indicate that the eastern source areas were still undergoing erosion. The clastic material present in interval C in east-central Tennessee and northwestern Georgia is also presumably derived from the eastern source area, although the sand and shale in the lower part of interval C in east-central Tennessee could have been derived from a source area in the vicinity of the Nashville dome.

In the Greendale syncline the Little Valley Limestone contains discrete beds of sandstone intercalated in the shaly limestone in the lower part of the formation as well as local accumulations of salt and gypsum in the immediate vicinity of Saltville, Va. In general, the Little Valley is more argillaceous than the overlying Hillsdale Limestone and *Denmar Formation*, although all three units show intercalation of calcareous sandstone and shale in the thick geosynclinal accumulation preserved in the Greendale syncline (pl. 10, fig. 3). Widespread and abundant beds of oolite and fragmented fossils are characteristic of the thinner shelf deposits to the northwest of the Allegheny Front. These beds are indicative of a high-energy environment of deposition in a shallow sea populated by an abundant fauna (pl. 11, fig. 3). Youse (1964, p. 473–474) suggested that the Greenbrier sea in which these strata were deposited was shallow with sufficient wave and current energy to form oolites and to spread them as sheet deposits of considerable lateral extent. He also suggested that relatively slow subsidence of the sea floor permitted the formation of abundant oolites in the high-energy environment and that these particles were moved about into fringing bars and beaches. The beds of oolite in the upper part of interval C in the shelf segment are commonly less porous and less well sorted than are those in the lower part of the interval. Most probably, the increase of intergrain cement resulted from a less complete winnowing of the oolites in a deeper water lower energy environment.

Apparently, the shelf-sea environment extended over most of the western part of the Appalachian basin during part of interval C time. The sandy limestone of Meramec age in southeast Ohio probably represents a higher energy environment than the finer grained slightly dolomitic limestone in the basal part of the Maxville Limestone of central Ohio. The absence of terrigenous clastics in the Maxville of central Ohio indicates that the sea was open for some distance to the north of the present outcrops.

In the southern part of the Appalachian Plateaus province the Warsaw, St. Louis, and Tuscumbia Limestones of early and middle Meramec age are composed of relatively pure well-indurated cherty fine to coarsely crystalline crinoidal carbonate. The Warsaw Limestone is sandy and shaly in the lower part at some localities. Probably, these limestones were deposited in open sea, although bioclastic elements within the limestones suggest that the sediments were subjected, periodically at least, to some wave and current action.

The late Meramec—*Ste. Genevieve* and lower part of the Monteagle Limestone are oolitic fossil fragmental limestones and may be shaly or silty in part. These limestones were probably deposited in a somewhat shallower high-energy marine shelf environment.

PALEOTECTONIC IMPLICATIONS

The presence of a thick sequence of dominantly clastic carbonate rocks of Meramec age in much of the geosynclinal segment of the Appalachian basin indicates that the bordering lands, particularly those adjacent to the southern half of the basin, had been reduced to areas of low relief in which chemical erosion was dominant (pl. 12, fig. 3). As a result of the state of relative tectonic tranquility, the rate of geosynclinal subsidence (pl. 10, fig. 3) considerably exceeded the rate of supply of terrigenous detritus, as demonstrated by the sporadic occurrence of thin sheets of noncarbonate shale and sandstone in the thick sequence of limestone in the geosynclinal segment of the basin and by the virtual absence of noncarbonate rocks in much of the shelf segment of the basin. The southern borderland had retreated only slightly southward and still occupied much the same position it had during interval B (pl. 12, figs. 2, 3).

In contrast to the relative inactivity in the southern borderlands, areas adjacent to the northern half of the

Appalachian basin continued to be elevated and eroded but at a decreasing rate as indicated by the smaller volume of clastic detritus that was trapped mainly on the eastern side of the growing geosynclinal segment of the basin. The rate of sedimentation was insufficient to keep the geosynclinal segment dry in southern Virginia, and the presence of marine fossils in the upper part of the *Stroubles Formation* in the Blacksburg-Pulaski synclinorium demonstrates that the sea eventually flooded the site of the foundered Price delta. Apparently, uplift was greater to the northeast and a larger quantity of detritus was produced because much of the red beds of the Mauch Chunk Formation in the Anthracite basins of eastern Pennsylvania was deposited subaerially.

Slow subsidence of the northern part of the basin was halted during the closing half of Meramec time by uplift to the north in eastern Canada. Rejuvenation of this source area released a flood of coarse-grained quartzose detritus which was swept south along the steepened paleoslope and poured into the northern end of the shallow sea that covered much of the Appalachian basin. Adams (1964) demonstrated that the bulk of the sand in the Loyalhanna Limestone was derived from a northern source. Scatterday (1963) has shown that during the late Meramec the recently inundated coastal plain in central Ohio was upwarped and subjected to a subaerial episode of shallow channel cutting. This gentle uplift, which may have been related to the more vigorous uplift in Canada, was sufficient to prevent the accumulation of Loyalhanna detritus in the central part of the State, although a belt of sandy limestone was deposited in the shallow high-energy sea in southern Ohio.

Subsidence of the geosynclinal segment of the Appalachian basin had relatively little effect upon the shoal or island that lay in east-central West Virginia (Randolph County). This area remained sufficiently positive during Meramec time to act as a barrier against the southward migration of sandy detritus from northern sources, and it blocked transportation of terrigenous clastics into the shallow sea in southern West Virginia.

INTERVAL D

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Interval D in eastern Kentucky, western Virginia, and much of West Virginia is divisible into a lower limestone and calcareous shale sequence and an upper sandstone, siltstone, and variegated shale sequence (pl. 9-D, secs. *e–e'*, *c–c'*, *b–b'*, interval D). The name Greenbrier has long been used for the calcareous part of the sequence in West Virginia and Virginia (pl. 15), although, as generally applied, the Greenbrier includes the calcareous strata of interval C as well (Wilpolt and Marden, 1959, p. 593). Butts (1940b, p. 374) introduced the name Gasper for the Chester part of the Greenbrier Limestone in Virginia, and the name has been used widely in Virginia, part of West Virginia, eastern Tennessee, northwestern Georgia, and northeastern Alabama. However, in parts of West Virginia the Greenbrier is treated as a series composed of many local units (Reger, 1916; Price and others, 1939, p. 267; Wells, 1950, p. 917). (See table 1.) Recently, the name Gasper has been abandoned at the type area in Kentucky and has been replaced by the name Girkin (Rainey, 1963). The change in nomenclature is followed here (pl. 15).

In east-central Tennessee the name Monteagle Limestone has been applied (Vail, 1959; Stearns, 1963) to uppermost interval C, the *Ste. Genevieve Limestone*, and to lowermost interval D, the *Girkin Formation*; however, in northwestern Alabama and northern Mississippi rocks equivalent to the Girkin Formation have been named the Pride Mountain Formation, a sequence of several shale, limestone, and sandstone members (Welch, 1958). In northeastern Alabama, northwestern Georgia, and the subsurface of northeastern Mississippi, part of the Floyd Shale is equivalent to the Girkin Formation. Overlying the Girkin or its equivalents throughout Tennessee, Georgia, Alabama, and Mississippi are, in ascending order: the *Golconda Formation* locally, the Hartselle Sandstone, Bangor Limestone, and Pennington Formation. In parts of northeastern Alabama the upper part of the Floyd Shale occupies all interval D and in parts of the subsurface of northeastern Mississippi the Floyd Shale occupies all interval D below the Pennington Formation.

Throughout much of southern West Virginia along the Allegheny Front and in adjacent Virginia, the lower part of the Girkin Limestone is replaced by a sequence of red rocks — namely, the *Taggard Formation* of Wells (1950, p. 904) or the Taggard Red Member of the Greenbrier Limestone of Wilpolt and Marden (1959, p. 594). Unfortunately, this unit cannot be recognized with certainty in the subsurface where it would be most useful for delimiting the boundary between Meramec and Chester rocks.

In the Greendale syncline the Girkin Limestone is overlain by the red Fido Sandstone (Butts, 1927), and the Fido is overlain by the Cove Creek Limestone (Butts, 1927, p. 16), which is considerably more argillaceous than the Girkin Limestone. Along the Allegheny Front to the northwest of the Greendale syncline the Girkin underlies a lateral equivalent of the Cove Creek, the Bluefield Formation, which is com-

posed of argillaceous limestone and calcareous shale. A heterogeneous sequence of sandstone, siltstone, and variegated shale overlies both the Cove Creek Limestone and the Bluefield Formation. These clastic rocks, which compose the bulk of interval D in the central part of the basin, make up the Pennington Group (Wilpolt and Marden, 1959, p. 598). In ascending order the units in the group are the Hinton Formation, Princeton Sandstone, and Bluestone Formation (pl. 15, cols. 20, 23, 25). Locally, members are recognized in some of the formations. In the Greendale syncline the Princeton has not been recognized and the Pennington is treated as a formation (Averitt, 1941, p. 11–14). Recent work by Englund (1964, p. B32; Englund and DeLaney, 1966, p. D50) demonstrated that the Pennington interfingers with the orthoquartzitic Lee Formation in southwestern Virginia and that in Lee County some of the Lee Formation is of Chester age. Thus, an undetermined thickness of the Lee belongs in interval D but has been excluded from the interval in this study because of the impossibility of locating, using existing criteria, the Mississippian-Pennsylvanian contact within the formation.

TABLE 1. — *Stratigraphic units assigned to interval D in parts of West Virginia and Kentucky*

[Shaded areas indicate beds are missing]

Map intervals	WEST VIRGINIA			KENTUCKY
	Central eastern		Southern	Central
Overlying units	New River Fm (Penn)		Pocahontas Fm (Penn)	Caseyville Ss (Penn)
Interval D	MAUCH CHUNK GROUP	Bluestone Formation	PENNINGTON GROUP Bluestone Formation	Leitchfield Formation / Buffalo Wallow Formation
		"Princeton" Sandstone	Princeton Sandstone	
		"Hinton" Formation	Hinton Formation	Tar Springs Fm
		Bluefield Formation	Bluefield Formation	Glen Dean Limestone
	GREENBRIER GROUP	*Alderson Limestone*	*Alderson Limestone*	Hardinsburg Ss; Haney Ls Member; Big Clifty Ss Mbr; Golconda Fm; Beech Creek Ls Mbr
		Greenville Shale	*Greenville Shale*	Elwren Ss equiv
		Union Limestone	*Union Limestone*	Reelsville Limestone; Sample Sandstone; Beaver Bend Ls; Girkin Formation
		Pickaway Limestone	*Pickaway Limestone*	Mooretown Formation
		Taggard Formation	*Taggard Formation*	Paoli Limestone
Underlying units		*Denmar Formation*	*Denmar Formation*	Ste Genevieve Limestone

The Girkin Limestone, which may be all or just the upper part of the well drillers' *Big lime*, is present in the subsurface of eastern Kentucky, southeastern Ohio, southwestern Pennsylvania, and West Virginia. Along the Allegheny Front in northern West Virginia, Maryland, and southern Pennsylvania, the Girkin Limestone thins and interfingers with red shale of the Mauch Chunk Formation. The Mauch Chunk is equivalent to some of the limestone of Meramec age, the Girkin, and the Pennington. In general the Mauch Chunk contains more red shale and less sandstone than does its southern equivalents. In southern Pennsylvania two thin tongues of limestone are present in the Mauch Chunk. The *Deer Valley Limestone Member* (Flint, 1965, p. 41) lies at the base of the red bed sequence and is locally separated from the younger Greenbrier Limestone Member by a tongue of red and purplish-red silty shale. The Greenbrier Member has recently been renamed the *Wymps Gap Limestone* by Flint (1965, p. 46). Flint demonstrated that these two limestones are north-projecting tongues from the main body of the Greenbrier (Girkin) Limestone of West Virginia.

The Mauch Chunk (Lesley, 1876, p. 221–222) Formation is thickest and contains much coarse-grained clastic rock in the vicinity of the Anthracite basins of Pennsylvania. The formation is about 5,000 feet thick in the Southern Anthracite basin (Dauphin County) and thins to the east, north, and west. It commonly contains a larger percentage of shale in the western part of Pennsylvania than at the type locality, Jim Thorpe, Carbon County, in eastern Pennsylvania. The Myers Shale of the Meadow Branch syncline has been correlated by Read (1955, p. 10–11) with the Mauch Chunk of Pennsylvania. The Myers is also a red-bed unit in which red shale is intercalated with red and gray sandstone.

Clastic strata of Chester age are not present in Ohio. However, as much as 80 feet of Chester-age Maxville Limestone is present locally. Scatterday (1963) has indicated that most of the Maxville in central Ohio is of Chester age, whereas in southern Ohio in the vicinity of Ironton, Lawrence County, more than half of the thickness of the Maxville is of Meramec age. The drillers' *Big lime* of southeastern Ohio is probably equivalent to the Loyalhanna Limestone and Greenbrier (Girkin) Limestone of southwestern Pennsylvania and northern West Virginia. The *Lime sand* part of the *Big lime* appears to be the Loyalhanna equivalent; whereas the overlying nonsandy limestone is the Girkin equivalent.

Throughout the subsurface of eastern Kentucky, as in much of West Virginia, the rocks composing interval D can be generally divided into limestone below and heterogeneous clastics above. The limestone, which forms the upper half of the drillers' *Big lime*, is equivalent to the several Chester limestones exposed along the eastern side of the Cincinnati arch in east-central Kentucky (McFarlan and Walker, 1956). The overlying sequence of clastic rocks is commonly called Pennington, and less frequently Mauch Chunk. The Pennington consists of several zones of lenticular sandstone intercalated in variegated shale and siltstone. The well drillers have given the names *Bradley*, *Maxton*, and *Salt sands* to these lenticular sandstones. Because of intertonguing in the clastic sequence and deposition of similar-appearing sandstones in scour channels of Early Pennsylvanian age, these informally named sands can only be roughly correlated. At some places (as shown on pl. 15, cols. 17, 19, 26) the drillers' *Maxton* and *Salt sands* are of Mississippian age; at other places the same names have been applied to sands of Pennsylvanian age. Detailed examination of drill cuttings will commonly resolve the age of the drillers' *Bradley*, *Maxton*, and *Salt sands*. In general the units in the lower part of the Pennington appear to have greater lateral extent than do those in the upper part. Several relatively thin units, the *Reynolds Limestone* and the *Webster Springs Sandstone* (Flowers, 1956), for example, have been traced widely in the subsurface of southern West Virginia and adjacent Kentucky.

UPPER BOUNDARY OF INTERVAL D

In the vicinity of the Southern and Middle Anthracite basins the red beds in the upper part of the Mauch Chunk Formation intertongue with basal conglomeratic sandstones of the Pottsville Formation in a transition zone 300–600 feet thick (Wood and others, 1962, p. C39). At Pottsville on the south side of the Southern basin, the systemic boundary is at the base of the Tumbling Run Member, the basal unit of the Pottsville Formation; at Upper Lehigh in the eastern Middle Anthracite basin, however, the basal beds of the Tumbling Run Member are younger and the systemic boundary lies in the upper part of the Mauch Chunk (Wood and others, 1962, p. C39). The Mauch Chunk thins markedly to the north across the southern part of the Northern Anthracite basin from more than 1,000 feet south of Wilkes-Barre to a featheredge in the vicinity of Pittston about 10 miles to the northeast. Data are insufficient to determine if the abrupt thinning of the Mauch Chunk is due simply to deposition or is a combination of deposition and pre-Pennsylvanian erosion. Considering that Pre-Pennsylvanian erosion has cut down into the Upper Devonian rocks at the north end of the Northern Anthracite basin about 20 miles farther to the northeast, some Mauch Chunk was probably removed by pre-Pennsylvanian erosion in the area between Wilkes-Barre and Pittston. Inasmuch as all the Mauch Chunk is included in the Mississippian in this report, the Pottsville is shown (pl. 15) as the overlying unit throughout the Anthracite basins.

Similarly, Mississippian and Pennsylvanian rocks are transitional and intertongue in exposures along the Allegheny Front from Mercer County, W. Va., south to Cumberland Gap. In Tazewell County, Va., rocks of the Bluestone Formation of the Pennington Group intertongue with strata of the overlying Pocahontas Formation, the basal unit in the Pottsville Group in this area (Englund and DeLaney, 1966, p. D51). The position of the systemic boundary must be determined from the contained floras, because deposition was continuous across the boundary. To the southwest in the vicinity of Cumberland Gap, Englund (1964, p. B31; Englund and DeLaney, 1966, p. D50) demonstrated that the Pennington Formation intertongues with the overlying Lee Formation which is of Mississippian age in the basal part. The systemic boundary lies within the coarse-grained conglomeratic Lee Formation. Previously, the basal bed of the Lee had been used to delimit the Mississippian-Pennsylvanian boundary, and a hiatus equivalent to the Bluestone Formation was assumed to be represented by an unconformity at the base of the Lee Formation in extreme southwestern Virginia (Butts, 1940b, p. 407). Englund's work has shown, however, that the lower part of the Lee intertongues with the Bluestone along the Allegheny Front northeast from Cumberland Gap, and the hiatus is nonexistent. The specific placement of the systemic boundary within the coarse-grained Lee Formation is uncertain, and the top of interval D is arbitrarily placed at the only practical lithologic subdivision, the base of the Lee Formation.

In parts of northeast Alabama the Parkwood Formation, the base of which in this report is used as the arbitrary top of interval D, is of transitional Mississip-

pian-Pennsylvanian age and is apparently conformable with the Pennington Formation or the Floyd Shale, depending on which unit it overlies (Weller and others, 1948; Rothrock, 1949).

In contrast to these three areas where the sequence is conformable, the systemic boundary is an unconformity along the Allegheny Front from southern West Virginia (Mercer County) north to north-central Pennsylvania (Lycoming County), in the northwestern part of the Pennsylvania Anthracite region, and from Cumberland Gap south to northeastern Mississippi. Throughout the southern part of the Appalachian basin from northeastern Tennessee to northern Mississippi, the hiatus is approximately equivalent to the upper part of the Pennington Group or the Bluestone Formation of eastern Kentucky. In this area the upper part of the Lee Formation (Tennessee) and the Pottsville Group or Formation (Georgia, Alabama, and Mississippi) of Pennsylvanian age overlie the lower part of the Pennington Formation with erosional unconformity. In the northern half of the Appalachian basin the magnitude of the hiatus increases northward from Mercer County, W. Va., where rocks of the Pocahontas Formation (Pottsville Group) are unconformable on beds in the middle of the Bluestone Formation (Pennington Group) (Thomas, 1966, p. 489). In Somerset County, Pa., the *Sharon Sandstone* of Flint (1965, p. 54), which is equivalent to the upper part of the New River Formation (Pottsville Group) of West Virginia, overlies red beds in the Mauch Chunk. The hiatus in southern Pennsylvania is equivalent to the Pocahontas Formation and much of the overlying New River Formation of the Pottsville Group of West Virginia, and to the Tumbling Run Member of the Pottsville Formation of the Antracite basins.

Pre-Pennsylvanian erosion beveled Mississippian strata along the northern and western edges of the Appalachian basin (pl. 9-C, secs. *f–f'*, *h–h'*, interval C). The Mauch Chunk was cut away north of a southwesterly trending line through Pennsylvania (north of a line between Lackawanna, Lycoming, and Allegheny Counties). The relief on the unconformity increases to the north inasmuch as older Mississippian units were removed, and in the southern part of Allegany and Cattaraugus Counties, N. Y., the basal Pennsylvanian Olean Conglomerate rests directly upon Upper Devonian rocks.

Patches of Chester carbonate rocks are present locally below Pennsylvanian-age sandstone in eastern and southern Ohio (pl. 6-A). Relief on the extensive pre-Pennsylvanian unconformity increases in magnitude to the north, and, locally, in northern Ohio the Sharon Conglomerate rests on the top of the Sunbury Shale (pl. 9-D, sec *b–b'*, interval D). In the subsurface of Boyd and Greenup Counties in northeastern Kentucky, the Lee Sandstone is locally absent, and Osage- or Chester-age shales are overlain disconformably by similar-appearing shale of Early Pennsylvanian age. In this area the hiatus is masked by the similar lithologies, and the systemic boundary cannot be precisely located.

The stratigraphic relations in the subsurface of eastern Kentucky are similar to those in West Virginia. Near the Allegheny Front in southeastern Kentucky the massive Lee Formation is thick, and the base of the oldest "Lee" type sandstone is commonly and arbitrarily selected as the boundary between systems, although some of the massive sandstone may well be of Late Mississippian age. To the north and west, where pre-Pennsylvanian erosion has produced a marked hiatus, the basal Pennsylvanian sandstone can be identified with considerable certainty at places where it rests on variegated shale in the Pennington. However, at many places the Pennsylvanian sandstone fills channels scoured in only slightly less coarse-grained sandstone in the Pennington, and delimiting the systemic boundary becomes an arbitrary choice.

From northeastern Tennessee southward throughout most of the southern part of the Appalachian basin the top of interval D is placed at the top of the Pennington Formation and is generally marked by a change from varicolored shale and interbedded limestone below to sandy shale, sandstone, and coal above. Where the Floyd Shale underlies Pennsylvanian rocks the top of interval D has been placed at the base of the sandy, coal-bearing sequence.

In parts of northeastern Mississippi and northwestern Alabama, pre-Pennsylvanian or pre-Cretaceous erosion or both have beveled interval D, and the Tuscaloosa Formation of Cretaceous age consisting of chert gravel with lesser amounts of shale and clay rests on the eroded surface.

THICKNESS TRENDS

The overall thickness pattern of the strata in interval D (pl. 6-A) is relatively simple in the Appalachian basin. A thick mass of rock fills the eastern geosynclinal segment of the basin; whereas a relatively thin blanket of strata covers the western shelf segment and laps upon the east side of the Cincinnati arch. Centers of accumulation in the geosynclinal segment are the Anthracite basins of Pennsylvania, the Hurricane Ridge syncline of Mercer and McDowell Counties, W. Va., the Greendale syncline of southwest Virginia, the Valley and Ridge province of northwestern Georgia and east-central Alabama, and the Black Warrior basin of east-central Mississippi. Thicknesses of more than 4,000 feet of interval D were attained in several of these centers; in contrast, scant data suggest that interval D was probably not more than several hundred feet thick

in the shelf segment to the west. Local isopachs indicate that the Chester rocks are thicker along the axes of the synclines than on the adjacent anticlines in southwestern Virginia and adjacent West Virginia (Wilpolt and Marden, 1959; Cooper, 1964, p. 101; Thomas, 1966, p. 476). These data strongly suggest that local folding occurred during deposition of the Chester rocks and that the growing synclines acted as traps for detritus eroded from adjacent rising anticlines.

West and northwest thinning of the Chester rocks has been accentuated by pre-Pennsylvanian erosion. South-flowing streams incised the western and northern edges of these rocks and cut deep indentations, such as those in northwestern West Virginia (Wirt and Wood Counties), into and through the thinner sequence of interval D on the west flank of the basin. Irregular patches of Maxville Limestone in central Ohio were produced by two episodes of erosion. Maxville Limestone of Chester age filled channels cut into and through the Meramec age Maxville (Scatterday, 1963), and pre-Pennsylvanian erosion dissected and removed large amounts of the sheet of Chester-age Maxville that once covered most of southeastern Ohio.

LITHOFACIES TRENDS

Because interval D is filled by a relatively heterogeneous sequence of rocks that has been beveled by erosion at an acute angle to the depositional strike of the strata, the lithofacies trends shown on plate 6-B, which are based on the preserved rocks of interval D, are overaccentuated and may be misleading in terms of the original lithofacies distribution of the interval before erosion. In the Greendale syncline of Virginia, at a locality within the geosynclinal segment of the basin, a thick sequence of clastic rocks overlies the Girkin and Cove Creek Limestones (pl. 9-E, sec. $A-A'$, interval D). The lithofacies pattern shows a relatively large volume of shale and sandstone in the sequence, although as much as 2,200 feet of limestone and dolomite is present. In contrast, interval D in southern Ohio is all limestone, because pre-Pennsylvanian erosion has removed any Pennington or Mauch Chunk clastics that may have been deposited there. Although the Maxville in southern Ohio is only about one-thirtieth as thick as the sequence of Girkin and Cove Creek carbonate in the Greendale syncline, it is shown on the lithofacies map by the limestone pattern, whereas the carbonate rocks in southwest Virginia are masked by the suprajacent mass of clastic rocks. Largely because of truncation of the upper part of the Chester rocks in the northwest part of the basin, the lithologic pattern of the lower calcareous part of the interval is exaggerated.

Tongues of clastic strata project west and northwest from the centers of accumulation east of the Allegheny Front and lap onto the shelf segment of the basin to the west of the geosynclinal segment. Commonly, the rocks become finer grained to the west and north. The lobe of sandstone pattern in west-central Pennsylvania (Indiana County) that suggests coarsening texture to the north is again an instance of accentuation of a relatively thin sequence of beds near the eroded edge of the Mauch Chunk Formation.

In general, the lithofacies map of the southern part of the Appalachian region shows a true picture of deposition in this area during interval D time. In the shelf segment the lower part of interval D is for the most part limestone with some shale (the Girkin Formation), whereas the upper part is generally shale with some interbedded sandstone and limestone (the Golconda, Hartselle, Bangor, and Pennington Formations). In the geosynclinal segment to the southeast and in the Black Warrior basin of Mississippi, the entire interval D is made up of calcareous sandy shale (the Floyd Shale) or by a sequence of interbedded calcareous sandy shale, sandstone, and limestone (the Pride Mountain, Hartselle, Bangor, and Pennington Formations).

SOURCES AND ENVIRONMENTS

The major source of terrigenous clastic sediment during Chester time was the borderland lying east and southeast of the geosynclinal segment of the Appalachian basin (pl. 10, fig. 4). From exposed masses of metamorphic and sedimentary rocks in and east of the present Piedmont province, streams swept large amounts of detritus to the west and northwest down the paleoslope into the Appalachian basin, where the materials were winnowed and sorted by streams, waves, and currents. The source area for the Mauch Chunk red beds of Pennsylvania apparently lay east of the present coastline of New Jersey.

Polymictic conglomerates and poorly sorted coarse-grained sandstones in the Pennington Group in the Greendale and Hurricane Ridge synclines of southwest Virginia indicate a relatively local source for much of the clastics in these beds. Sandstone, siltstone, and shale in the Chester strata in the geosynclinal segment of the basin from eastern Tennessee to north-central Mississippi presumably had a southeastern source (pl. 10, fig. 4) (King, 1950, p. 663); however, some of the material may have been derived from uplift to the northwest in the vicinity of the Nashville dome and Ozark uplift (Thomas, 1967, p. 8, 9). Swann (1964, p. 653) suggested that the sediments which formed the Late Mississippian clastic rocks in the western part of the Black Warrior basin in Mississippi must have been transported by the Michigan River across the Illinois basin since "They are isolated from potential sources in the southern Appalachians by the contemporaneous

limestone sequence of Alabama and from Llanoria by deep-water deposits of the Ouachitas."

During the early part of Chester time the land east and southeast of the geosyncline was topographically low and supplied only a relatively small quantity of debris to the geosyncline (pl. 10, fig. 4). Only in the northeastern part of the basin was a sufficient amount of terrigenous clastics being deposited to fill the trough above sea level (pl 12, fig. 4*A*). Much of the geosynclinal segment of the basin, as well as the shelf to the west, was covered by a shallow sea in whose active waters carbonate clastics were sorted and winnowed into bars and banks.

The source of clastic carbonate, which is the dominant lithology in the central and southern parts of the geosynclinal segment and the western shelf segment of the basin during the first half of Chester time, was within the basin itself. Beds of oolite and whole or broken fossils formed in shallow, well-agitated water. These strata are intercalated with beds of fine-textured limestone and calcilutite. The texture of the rock depends predominantly on the energy level within the depositional environment and the distance from the source of encroaching terrigenous detritus.

Throughout much of the Appalachian basin depositional environments in the early half of Chester time contrast strongly with those in the later half (pl. 11, fig. 4 represents an intermediate stage in mid-interval D). Shallow-water high-energy deposition of clastic carbonate dominated both geosynclinal and shelf segments of the basin except in the northeastern part of the geosynclinal segment where red beds of the Mauch Chunk delta complex were filling the geosyncline. In this local area, some detritus was being deposited subaerially in a fluviatile environment.

In the Girkin Limestone and the *Monteagle Limestone* of both segments of the basin as well as in the Maxville Limestone of the shelf segment, lenticular beds and sheets of oolite and fragmented fossils indicate high-energy marine environments; whereas intercalated beds of calcilutite with some chert and shale indicate low-energy environments (pl. 12, fig. 4*A*). The coarse-textured high-energy rocks apparently accumulated as banks and bars; the finer textured lower energy strata were deposited in lagoons, on mud flats, or in the deeper waters adjacent to the growing bars and banks. The presence of scattered beds of quartzose sandstone or siltstone in the dominantly carbonate sequence marks incursions of terrigenous clastics into the sea from the growing delta complex to the east or southeast, probably during times of excessively high energy.

The presence of increasing amounts of argillaceous calcilutite in the upper part of the Girkin Limestone and in both the Cove Creek Limestone and the Bluefield Formation in the southwest Virginia segment of the geosyncline suggests a shift to a lower energy environment, probably from the shoaling of the sea and restriction of wave energy along the front of the growing Mauch Chunk–Pennington delta complex in mid-Chester time.

A greater diversity of depositional environment existed in the Appalachian basin in the later half of Chester time as the thick wedge of terrigenous detritus of the Mauch Chunk–Pennington delta complex expanded westward to fill the geosynclinal segment and to displace the sea from the shelf segment of the basin in the closing phase of the epoch (pl. 12, figs. 4*B*, 4*C*). Fluvial deposition dominated a broad coastal plain of the delta complex which lay along the eastern side of the basin and separated the rising source areas on the east from the shrinking sea on the west. Red beds accumulated abundantly in the Mauch Chunk of eastern and central Pennsylvania. Coarse-grained rocks associated with coal beds in the Pennington Group along the Allegheny Front in southwestern Virginia were deposited in a high-energy fluvial environment on a broad coastal plain that contained widespread coal swamps in protected localities between the anastomosing distributary channels of the major rivers draining from the highlands to the east.

A broad belt of sedimentary rocks consisting of intercalated marine and nonmarine beds extends along the eastern side of the shelf segment of the basin from western Pennsylvania southwest to Tennessee, and predominantly marine beds extend along the southeastern part of the shelf from eastern Tennessee into mortheastern Mississippi. In these zones, high-energy nearshore bar and beach environments are indicated by the presence of relatively well sorted and clean quartzose sandstones. These high-energy strata are intercalated with finer grained rocks characteristic of low-energy environments of lagoon, backbay, mudflat, and overbank accumulations. Some of the better sorted gas sands of the subsurface of eastern Kentucky appear to have accumulated in a high-energy nearshore environment possibly as offshore bars and shoals. The presence of impure fossiliferous calcarenites associated with some of the gas sands also suggests a marine depositional environment.

In the late Chester (pl. 12, fig. 4*B*), the influx of terrigenous detritus overwhelmed the geosynclinal segment of the basin, and in the northeast the shoreline of a large delta complex transgressed into the shelf segment west of the trough. A low coastal plain extended from northeastern Pennsylvania southwest through West Virginia, eastern Kentucky, and Tennessee into Georgia, Alabama, and Mississippi and may have reached as far west as southwestern Louisiana. The presence of coal beds in the upper part of the Chester

sequence north of Tennessee indicates that much terrestrial vegetation grew in a benign climate and that coal swamps formed in depressions on the coastal plain. The sea was displaced from the Appalachian basin except for brief transgressions which are indicated by tongues of argillaceous marine limestone in the Pennington in parts of Mississippi, Alabama, Georgia, Tennessee, eastern Kentucky, southwestern Virginia, and southern West Virginia. The greater abundance of red beds in the Mauch Chunk of eastern Pennsylvania than in the Pennington of Virginia suggests that the climate may have been more arid in the northern part of the source area and in the adjacent part of the basin of deposition.

In the southern part of the Appalachian basin evidence suggests gradual uplift in late Chester time of the eastern landmass, associated gradual filling of the geosyncline with fine-grained clastic sediments, and eventual spreading of the clastic facies northwestward over the shelf segment of the sea. If a rapid rise of the eastern land area took place, if coal swamps or deltas existed, or if any continental deposits were laid down in the southeastern part of the southern Appalachian region in Chester time, all evidence of such events has since been obliterated.

PALEOTECTONIC IMPLICATIONS

Subsidence dominated most of the Appalachian basin during the early part of the Chester. Downwarp carried the recently emergent shelf segment in central Ohio below sea level again as the Chester sea flooded widely across the basin. Source areas of terrigenous detritus were uplifted to the east of the geosynclinal segment of the basin in and contiguous to eastern Pennsylvania, and large amounts of debris, ranging from cobbles to mud, were swept from the rising hills and carried westward into the subsiding trough in eastern and central Pennsylvania (pl. 10, fig. 4). In contrast, source areas of terrigenous detritus adjacent to the central and southern part of the basin remained largely quiescent, as shown by the predominance of relatively pure carbonate strata of early Chester age in much of the Appalachian basin. Maximum flooding of the subsiding basin by Mississippian seas, which presumably coincides with maximum subsidence in the area, took place before mid-Chester time (pl. 12, fig. 4*A*).

A pulse of orogenic activity beginning about mid-Chester uplifted the source areas of terrigenous detritus to the east and southeast of the Appalachian basin. The center of activity appears to have been located near the northern part of the geosyncline where fine- to coarse-grained detritus was stripped from a terrain of complex lithology and deposited in the broad Mauch Chunk delta-alluvial plain complex to the west. Predominantly finer grained clastic sediments accumulated to the south and southwest suggesting that the source areas of these sediments were not as strongly uplifted or were farther from the Appalachian trough or a combination of both these factors. The detrital sediments produced by the uplift displaced the Mississippian sea from the geosynclinal segment of the basin and locally prograded onto the adjacent shelf segment of the basin (pl. 12, fig. 4*B*), inasmuch as the rate of accumulating sediment greatly exceeded the rate of subsidence of the basin.

Uplift of source areas contiguous to the eastern side of the basin continued throughout late Chester time, apparently at an accelerating pace. The presence of pebbles and cobbles of locally derived rocks, ranging in age from Silurian to early Chester in the Hurricane Ridge syncline of southwestern Virginia (Cooper, 1964, p. 103; Thomas, 1966, p. 481), indicates that Paleozoic rocks, including some recently deposited strata, had been upwarped and gently folded in close proximity to the syncline. Field data from several parts of southwest Virginia suggest that open folding occurred locally during the late Chester. Abundant coarse-grained pebbly sandstone in the upper part of the Mauch Chunk in eastern Pennsylvania indicates that uplift of contiguous source areas to the east continued at an increasing rate during the close of Chester time (pl. 12, fig. 4*C*).

The abundance of finer grained detritus in the southern part of the basin and the general absence of beds of cobbles and pebbles in the Pennington and its lateral equivalents indicate that sites of uplift were remote from this part of the basin and that the sediment-transporting streams may have crossed a broad alluvial plain prior to discharging their load of sediment in the Appalachian trough. Certainly, the low hills and high plains produced a large volume of detritus which filled the geosynclinal segment of the basin and ultimately displaced the sea from much of the shelf segment to the northwest.

Uplift to the north in the Canadian Shield and southward tilting of the northwest part of the basin coupled with renewed orogenic activity to the east of the basin terminated Chester deposition in most of the Appalachian basin. An episode of deep-valley cutting took place along the northern and northwestern side of the basin, where recently deposited Chester strata were brought well above sea level by differential warping. Pre-Pennsylvanian and Early Pennsylvanian erosion removed large amounts of Mississippian rocks, particularly in northeastern Ohio and northwestern Pennsylvania (pl. 9-D, secs. c–c', b–b', interval D). Basal Pennsylvanian conglomeratic sandstone and conglomerate fill south-sloping scour channels in the

Mississippian strata, indicating that in the northwest part of the basin tilting locally reversed the general west-dipping Chester paleoslope. Erosion also cut deeply into Chester-age strata in northeastern Kentucky and northern West Virginia and less deeply in Maryland, Tennessee, Georgia, Alabama, and Mississippi.

In contrast, subsidence continued along the axial part of the geosyncline, and deposition appears to have been continuous from Mississippian to Pennsylvanian near centers of accumulating clastics in eastern Pennsylvania, southwestern Virginia, and central Alabama. These data suggest that orogenic forces adjacent to the basin differentially uplifted and warped the Appalachian area at the close of Chester deposition, and so some parts of the basin were subjected to vigorous dissection and erosion, whereas other parts were unaffected.

TOTAL THICKNESS OF MISSISSIPPIAN SYSTEM

THICKNESS TRENDS

The total thickness of the Mississippian rocks (pl. 7) is similar in general outline to the thicknesses of the intervals (pls. 3-A, 4-A, 5-A, 6-A). A west-thinning wedge of strata fills the geosynclinal segment on the east side of the Appalachian basin. Two maximums of 7,000 feet or more each mark the sites of greatest accumulation. The northern site centers in the western part of the Southern Anthracite basin of Pennsylvania, where the Mississippian rocks may be more than 8,000 feet thick, and the southern falls in the Greendale syncline of southwestern Virginia. A third locality in which more than 6,000 feet of Mississippian was deposited is the Hurricane Ridge syncline of Mercer County, W. Va. Southwestward along the geosyncline in northwestern Georgia, thickness of accumulated sediment is as much as 3,000 feet. The shelf segment of the basin, northwest of the geosynclinal segment, contains a thinner sequence of Mississippian strata. The rocks in the shelf segment are about one-tenth as thick as the rocks in the geosynclinal segment in the northeast and from one-fifth to nearly one-third as thick in the south.

The thickness in the northwestern part of the basin has been profoundly affected by erosion: pre-Chester, pre-Pennsylvanian, and post-Pennsylvanian. Deep incision of pre-Pennsylvanian valleys and dissection of the Mississippian strata have overprinted a pattern of alternating thick and thin linear bodies of strata trending in directions deviating considerably from the sedimentary strike of the beds. The belt of relatively thick strata which trends south across the east-central part of Ohio is depositional. The presence of the thick sandstone of the Black Hand Member of the Cuyahoga Formation makes the sequence thicker there than in areas in eastern Ohio where the equivalent rocks consist largely of more compactible shale and mudrock. Pre-Pennsylvanian erosion has greatly thinned the rocks in northeastern Ohio.

An area of relatively thin strata centered in Braxton County, W. Va., is an anomalous indentation on the west edge of the geosynclinal segment of the basin. This thin spot results from the nondeposition of most of the strata of Kinderhook, Osage, and Meramec age in Braxton County.

Post-Pennsylvanian erosion has greatly exaggerated the northwestward thinning of the Mississippian sequence in east-central Tennessee and northern Alabama, and a 500-foot isopach showing the originally deposited thickness would probably fall considerably north and west of the present 500-foot isopach (pl. 7).

PALEOTECTONIC IMPLICATIONS

Thickness data coupled with the lithofacies of the intervals (pls. 3-B, 4-B, 5-B, 6-B) indicate that for much of the Mississippian a land area of considerable size — Appalachia — bordered the Appalachian basin on the east. Periodic uplift of the source area and accelerated erosion supplied floods of clastic debris to the geosynclinal segment of the basin. The trough subsided at a rate sufficient to trap much of the sediment, although at times the trough filled and tongues of subaerially deposited sediment spread west onto the contiguous shelf segment. The existence of two widely spaced centers of accumulation in the geosynclinal segment in the northern part of the basin, both of which contain terrestrial deposits, suggests a broad piedmont alluvial fan–delta complex that covered parts of Pennsylvania, Maryland, West Virginia, Virginia, and North Carolina.

Although sediments from a northern source area, the Canadian Shield, had a considerable effect on the lithofacies patterns in the northwestern part of the basin where the Mississippian strata are thin, the volume of detritus from this source was relatively insignificant when compared to the large quantities of terrigenous clastics shed by the eastern source area.

The southern part of the land area, Appalachia, extended southwestward from Virginia to south-central Alabama but was probably never as high or rose as rapidly as its counterpart to the northeast. The terrigenous clastics in the geosynclinal segment of this part of the basin are fine-grained marine deposits and do not indicate rapid uplift. If a marine facies consisting of coarse-grained terrigenous clastic rocks was deposited or if a swamp or fluviatile continental facies was deposited, they have since been obliterated by uplift and erosion along the southeastern margin of the Valley and Ridge province.

Plate 7 shows a relatively shallow trough en echelon to the main geosyncline extending from east-central Mississippi northeast through northern Alabama and into east-central Tennessee. This trough was bordered on the northwest by a positive area in the approximate position of the Nashville dome. Uplift of the positive area took place from time to time during the Mississippian, and it may have been nearly or slightly emergent during parts of intervals B, C, and D time.

GEOLOGIC UNITS DIRECTLY ABOVE THE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Strata of the Mississippian System are overlain by Pennsylvanian age rocks (pl. 8) throughout the Appalachian basin, except in northern Alabama and northern Mississippi where Cretaceous age Tuscaloosa gravels lie on the eroded Mississippian beds or elsewhere in the basin where post-Permian erosion has stripped away the Pennsylvanian cover and unconsolidated Quaternary detritus mantles the Mississippian rocks.

Near Pottsville, Schuylkill County, Pa., in the Southern Anthracite basin, the oldest Pottsville rocks, the Tumbling Run Member of the Pottsville Formation, overlie red beds of the Mauch Chunk Formation. The basal beds of the Pottsville become younger to the north and west of the Southern Anthracite basin, and in the Northern Anthracite basin the conglomeratic beds of the Sharp Mountain Member of the Pottsville Formation, the youngest member of the formation, disconformably overlie strata of Early Mississippian and of Late Devonian age.

To the south of the Anthracite basins along the axial part of the Appalachian basin in southwestern Virginia and contiguous parts of Kentucky and West Virginia, the oldest Pottsville rocks, which contain plants of floral zone 4 of Read and Mamay (1964, p. K6), conformably overlie the youngest Mississippian rocks, which contain plants of floral zone 3 of Read and Mamay. Deposition was virtually continuous across the systemic boundary in this segment of the Appalachian basin. In Tazewell County, Va., the Pocahontas Formation of the Pottsville Group conformably overlies the Bluestone Formation of the Pennington Group. To the southwest along the Allegheny Front toward Cumberland Gap the two formations intertongue with the Lee Formation which is of both Mississippian and Pennsylvanian age (Englund and DeLaney, 1966, p. D50). In eastern Tennessee and southward along the axial part of the basin, the Mississippian age Pennington Formation is overlain by rocks of Pennsylvanian age, except in areas of post-Pennsylvanian erosion. In Tennessee the quartzitic Lee Formation caps the Pennington. In northwest Georgia, northern Alabama, and northeastern Mississippi, the Pottsville Formation or the Pottsville Group make up the overlying unit. Locally, in east-central Alabama the Parkwood Formation, a transitional unit of both Mississippian and Pennsylvanian age, overlies the Pennington.

Because of extensive pre-Pennsylvanian erosion of the northwestern part of the Appalachian basin and filling of the basin by Pennsylvanian sediments from several source areas, strata in the Pottsville Formation are in contact with Mississippian rocks ranging in age from Chester to Kinderhook in various parts of Kentucky, Maryland, Ohio, Pennsylvania, and West Virginia. The basal beds of the Pottsville become progressively younger from Tazewell County, Va., north and northwest across West Virginia and into western Pennsylvania. In the same direction the basal Pennsylvanian strata overlap all the Mississippian beds from youngest Chester in Virginia to oldest Kinderhook in northwestern Pennsylvania and adjacent northeastern Ohio. In southern West Virginia the New River Formation of the Pottsville Group overlaps the underlying Pocahontas Formation and lies on the Pennington Formation or the Mauch Chunk in much of central West Virginia. A quartzose Lee facies, probably largely equivalent to the New River Formation, is present in the subsurface of eastern Kentucky where it fills channels scoured in the Pennington or Mauch Chunk beds. The Connoquenessing Sandstone Member of the Pottsville Formation, an equivalent of the upper part of the New River Formation or the lower part of the younger Kanawha Formation, is the basal Pennsylvanian unit in much of the Allegheny Plateau in western Pennsylvania and eastern Ohio. The Sharon Conglomerate Member of the Pottsville Formation blankets Mississippian strata in much of northeastern Ohio, and its equivalent, the Olean Conglomerate Member of the Pottsville Formation, caps the Mississippian rocks in part of northwestern Pennsylvania and adjacent southwestern New York. Both conglomerates, which appear to be slightly older than the Connoquenessing Sandstone Member, were derived from a northern source area and lie between the eroded Mississippian rocks below and the main body of the Connoquenessing and younger Pennsylvanian rocks above. They are probably equivalent to the upper part of the New River Formation of West Virginia, the upper part of the Lee Formation of east-central Kentucky, and the Schuylkill Member of the Pottsville Formation in the Anthracite basins of eastern Pennsylvania (Read and Mamay, 1964, p. K6–K7). The correlation of the coarse-grained Pennsylvanian units is commonly uncertain in the subsurface because of the similarity of the sandstones and the absence of diagnostic plant fossils in drill cuttings.

Consequently, delineation of the areas occupied by the various units of the Pottsville is not precise.

Similarly, the basal Pennsylvanian rocks decrease in age between the Southern Anthracite basin and areas in the Allegheny Plateau to the north and west; however, this relationship is not so clearly shown as it is farther south because recent erosion has left large gaps between the Pennsylvanian outliers in northern Pennsylvania.

PALEOTECTONIC IMPLICATIONS

Terminal Chester uplift rejuvenated source areas, particularly those adjacent to the Anthracite basins of Pennsylvania and to the Tennessee-Alabama geosynclinal segment of the Appalachian basin, as shown by the great influx of cobbly coarse-grained detritus in the oldest beds of the Pottsville in these areas. An episode of uplift tilted and differentially warped the northwestern edge of the basin at the close of Chester time. A source area to the north in eastern Canada was briefly uplifted, possibly concurrent with the tilting in northern Ohio, and produced the pebbly detritus, which was carried south into the basin to become the Sharon Conglomerate Member of the Pottsville Formation in northern Ohio and the Olean Conglomerate Member of the Pottsville Formation in northwestern Pennsylvania.

In contrast to positive movement along the periphery of the basin, virtually continuous deposition in the axial part of the basin from eastern Pennsylvania south to Alabama demonstrates the persistent negative character of the geosynclinal segment of the Appalachian basin during the Late Mississippian and Early Pennsylvanian.

Most of the movement in the Appalachian basin at the close of Mississippian time appears to have been vertical. However, in the vicinity of the Greendale and Hurricane Ridge synclines in southwestern Virginia, horizontally directed compressional forces that produced open folds (pl. 10, fig. 4) appear to have accompanied the uplift of adjacent source areas. Evidence of compressional forces in other parts of the basin is inconclusive.

Deposition of detritus continued in the Appalachian basin during the Pennsylvanian Period and the early part of the following Permian Period. Unwarped and folded source areas bordering the basin shed terrigenous material into the basin but in decreasing amounts after the great influx of very coarse grained debris during the Early Pennsylvanian. The Late Pennsylvanian and Early Permian rocks contain a much greater percent of fine-grained clastics than do the Early Pennsylvanian rocks. Only occasionally after Chester time was the basin flooded by marine water from the western epicontinental sea. Cyclothemic accumulation of coal-bearing rocks continued in a basin of dwindling size. Pottsville strata covered much of the same area that is underlain by Chester strata; whereas the Lower Permian Dunkard Group, the youngest Paleozoic rocks in the Appalachian basin, cover only a small area in southwestern Pennsylvania and adjacent parts of Maryland, Ohio, and West Virginia (Berryhill, in McKee and others, 1967, pl. 1). The general tectonic framework of the basin remained intact until it was destroyed by the Appalachian revolution at the close of the Paleozoic Era.

During the Appalachian revolution in the eastern part of the basin, the Valley and Ridge segment, complex folding and faulting were dominant; whereas in the western part, the Appalachian Plateaus segment, vertical uplift was dominant, although some gentle folding and faulting occurred locally. The Paleozoic rocks along the southeastern margin of the Appalachian region were gently folded and faulted and, subsequently, were beveled. After the Appalachian revolution, long periods of erosion punctuated by brief episodes of uplift have dominated the history of the Appalachian basin. Subsidence in the Mississippi embayment area in Late Cretaceous time instigated deposition of the Tuscaloosa gravel on the eroded Paleozoic rocks in northern Mississippi, northwestern Alabama, and central Tennessee. Stearns (1957, p. 1092–1093) stated that the Tuscaloosa Formation is probably a nonmarine "back beach" deposit of limited extent and that it represents the initial transgressive stage of a sedimentary cycle that lasted into early Eocene time. The sea reached its maximum transgression up the Mississippi embayment during Paleocene time. From late Eocene to Pleistocene time as the continent gradually emerged, the sea withdrew from the coastal plain, and since the last glacial period the present shoreline is believed to have developed (Hunt, 1967, p. 162).

During the Pleistocene epoch the Mississippian strata along the northern edge of the Appalachian basin in Ohio and Pennsylvania were locally scoured and eroded by several continental glaciers that moved into the Appalachian area from centers of accumulation to the north in Canada. Glacial ice and melt water from the waning ice sheets cut deep valleys into and through the exposed Mississippian rocks, particularly in northern Ohio where the softer beds in the Cuyahoga Formation were stripped back from the more resistant Berea Sandstone and from the Black Hand Sandstone Member of the Cuyahoga. Much of the work of glaciers in the northern part of the basin was constructional rather than erosive. Large amounts of detritus were deposited in till plains, kame terraces, terminal

moraines, and eskers. At many places the newly scoured Mississippian beds were buried under considerable thicknesses of glacial drift or associated lake beds.

Since the retreat of the Pleistocene glaciers, erosion has been slowly diminishing the relief of mountains and hills within the Appalachian basin.

SUMMARY

The major tectonic elements of the Appalachian basin — namely, an elongate miogeosynclinal eastern segment partly surrounded on the north, east, and southeast by bordering source areas and a western relatively stable shelf segment — were inherited from earlier Paleozoic periods. One relatively new element, a northern source area in eastern Canada, played an important role in the history of the northern part of the basin. Repeated uplift and erosion of this source area produced large amounts of terrigenous detritus which, when carried into the basin by streams and rivers, considerably modified the pattern of deposition in the northern part of the basin during several epochs of the Mississippian.

Active erosion of the uplifted bordering lands during the Late Devonian and Early Mississippian produced vast floods of coarse-grained clastic sediment which were swept into the basin by streams and rivers. The rate of geosynclinal subsidence was considerably exceeded by the rate of accumulating detritus in the northern half of the basin, and the geosynclinal segment was filled by a westward-expanding delta alluvial plain. Locally, sheets of pebbly sand and silt spread onto the relatively more stable shelf segment to the west. Concurrently, rivers transporting sediment from the northern source area built elongate deltas in the shallow epicontinental sea that covered the shelf segment in the northwestern part of the Appalachian basin. The lands bordering the southern part of the basin were of low relief and supplied only a small amount of fine-grained detritus to the basin.

An episode of uplift and local gentle folding occurred during interval A time at the close of Spechty Kopf deposition in the northeastern part of the basin, centering in the vicinity of the Anthracite basins in eastern Pennsylvania. The effects of the uplift included a brief withdrawal of the sea from a wide zone along the eastern edge of the shelf segment in West Virginia and southwestern Pennsylvania, trenching of the newly exposed coastal plain by consequent rivers and streams, and the warping of an area in east-central West Virginia which remained emergent until the end of Meramec (interval C) time.

In the late Kinderhook (interval A) the rate of subsidence overtook the dwindling rate of deposition in the northern part of the shelf segment, probably because the northern source area had been eroded to low relief, and a broad shallow stagnant sea similar to the Late Devonian Chattanooga sea covered the western part of the Appalachian basin. In contrast, in the northeastern part of the basin continued sedimentation from rising source areas promoted the growth of the broad Pocono-Price delta complex, which by late Osage (interval B) had displaced the sea from much of the northeastern part of the Appalachian basin. Renewed activity in Canada rejuvenated the northern source area during Osage time, but the volume of detritus transported into the basin from this source was relatively small when compared to the great amount of clastics shed by the eastern sources.

In contrast to the considerable activity in and surrounding the northern half of the basin during Kinderhook and Osage time, the southern part of the Appalachian basin was relatively quiescent. Contiguous source areas were of low relief and supplied only small amounts of finer grained detritus. Their major contribution was a large amount of material in solution or in colloidal suspension. Subsidence of the sea floor was slight, an extremely thin sheet of mudstone, the Maury Formation, was deposited during Kinderhook (interval A) time, and a relatively thin sheet of cherty carbonate strata, the Fort Payne Chert, spread over much of the southern half of the basin during Osage (interval B) time. By the late Osage, the source areas along the eastern side of the basin as far north as central Virginia had been reduced to low relief, and much of the detritus carried into the basin was of clay and silt sizes.

Reduction of much of the eastern borderlands to low relief and a decrease in tectonic activity in the source areas early in Meramec (interval C) time are indicated by the flooding of much of the geosynclinal segment of the basin by Meramec seas and the accumulation of a thick wedge of dominantly calcareous strata in the southern two-thirds of this segment of the basin. Only the source areas contiguous to the extreme northeastern part of the Appalachian basin stood high enough to supply detritus in volume sufficient to exceed the rate of geosynclinal subsidence. Sources adjacent to eastern Pennsylvania and northern Virginia produced a considerable amount of red mud and sand which was trapped locally in the eastern part of the geosynclinal segment and did not spread far into the shallow carbonate-depositing sea to the west.

Meramec flooding, which began in the south and spread northward along the geosynclinal segment and northwestward along the shelf segment, reached its maximum in the middle of the epoch when the St. Louis Sea inundated northeastern Kentucky and central Ohio. Rejuvenation of the Canadian source area and

uplift of the northwestern end of the shelf segment in central Ohio produced a flood of coarse-grained quartz sand which was incorporated in the carbonate sediment of a high-energy sea south of the newly emergent coastal strip to form the sandy Loyalhanna–*Ste. Genevieve Limestone* from southwestern Pennsylvania to northeastern Kentucky.

A low island in east-central West Virginia was apparently an effective barrier against the southern migration of sand and silt from the northern part of the basin. The proximity of the island to the edge of the Loyalhanna–*Ste. Genevieve* sandy limestone suggests that sand-carrying currents from the northeast impinged upon the shoals and were deflected into the western part of the basin instead of continuing south along the western edge of the geosynclinal segment into the shallow sea in southern West Virginia and southwest Virginia.

Early Chester (interval D) was also a time of widespread subsidence and flooding of the basin segments. Inundation of the Appalachian basin during the Mississippian reached a maximum during the early part of the Chester. A thick wedge of carbonate strata accumulated in the geosynclinal segment south of west-central Virginia, and a sheet of carbonate sand and mud blanketed the shelf segment from central Ohio to northeastern Mississippi. However, the tectonic quiescence implied by the sheet of carbonate strata is deceptive, because source areas bordering the basin on the east and southeast had begun to rise by late early Chester time. Locally, some gentle folding had exposed recently deposited strata to erosion. Rejuvenation of source areas became apparent first in the northeastern and eastern parts of the geosynclinal segment where large amounts of terrigenous material accumulated to form a coalesced delta complex of large size. During the early Chester these coarse-grained clastics were confined to the eastern part of the geosyncline; however, by late mid-Chester a delta alluvial plain was established along the eastern side of the Chester Sea, and an expanding wedge of detritus had displaced the sea from the axial part of the geosynclinal segment.

Throughout late Chester time the sea retreated to the west and southwest before the ever-increasing wedge of terrigenous clastics. The delta alluvial plain covered the geosynclinal segment of the basin and spread west onto the shelf segment. By late Chester time the sea had been displaced from all but the extreme western part of the shelf area, and much of the surface of the basin was above sea level. The source areas to the northeast and south of the basin were active during much of late Chester time; however, sediments from the source area on the Canadian Shield were carried into the Eastern Interior basin in Illinois and were not shed east of the Cincinnati arch.

At the close of Mississippian time the northwestern part of the shelf segment of the Appalachian basin was uplifted and differentially warped. Erosion in Early Pennsylvanian time cut valleys into and through the Mississippian strata in this part of the basin. In contrast, the axial part of the geosynclinal segment was unaffected by uplift, and deposition there was continuous across the systemic boundary. Accelerated uplift and erosion of the several source areas in early Pottsville time (Early Pennsylvanian) produced a great influx of very coarse grained detritus in the vicinity of the Anthracite basins of eastern Pennsylvania. The thick sequence of conglomerate and conglomeratic sandstone in the Southern and Middle Anthracite basins is indicative of a high-energy fluvial depositional environment. In contrast, the intercalated beds of coarse- and fine-grained strata, including coal beds in the late Chester and early Pottsville rocks in southern Virginia and in early Pottsville rocks from Tennessee southwestward into Mississippi, suggest fluvial deposition with alternating high- and low-energy environments.

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Michigan Basin Region

By GEORGE V. COHEE

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010-D

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MICHIGAN BASIN REGION

By GEORGE V. COHEE

ABSTRACT

During Mississippian time the Michigan basin was part of the Midcontinent craton, and its seaway primarily received clastic sediment largely derived from the Canadian Shield. The basin subsided sufficiently to accommodate about 2,400 feet of sediment. Chemical precipitates include some limestone, dolomite, anhydrite, and gypsum. Uplift, folding, and erosion in the basin took place in Chester time, near the close of the period.

REGION DEFINED

The Michigan basin includes the Southern Peninsula of Michigan, the eastern part of the Northern Peninsula of Michigan, eastern Wisconsin east of the Wisconsin arch, northern Indiana north of the Kankakee arch, northwestern Ohio north of the Findlay arch, and southwestern Ontario, Canada, west of the Algonquin axis (fig. 8). The deepest part of the basin is in Clare and Gladwin Counties, Mich., and the Central Basin area, as generally used, includes an area of about 100 miles in diameter surrounding those counties.

PALEOGEOLOGY

LOWER BOUNDARY OF MISSISSIPPIAN

BOUNDARY IN MICHIGAN

The boundary between the Mississippian and Devonian Systems in some parts of the Eastern United States has been in considerable debate since the establishment of the Mississippian Series of the Carboniferous System in 1891. Because of special facies relationships and the paucity of megafossils near the boundary, correlations and boundary selections have varied considerably from one worker to another. In recent years the great increase in knowledge of conodont ranges and correlations has resulted in refinement of Upper Devonian and Lower Mississippian stratigraphy in the United States. Because the rocks involved in the Devonian and Mississippian boundary of the Michigan basin are in the subsurface, it has been necessary to compare the stratigraphic sequence with the same sequence in the northwestern part of the Appalachian basin where the rocks are exposed. An erosional gap of about 50 miles across the Findlay arch in northwestern Ohio separates the rocks involved. Very little difference is found between the Bedford, Berea, and Sunbury sequence (pl. 15) in southeastern Michigan and northwestern Ohio, north of the Findlay arch, and in the northwestern part of the Appalachian basin.

In their study of the Berea Sandstone and associated rocks in the Appalachian basin, Pepper, de Witt, and Demarest (1954) placed the boundary between the Mississippian and Devonian at the top of the Cleveland Member of the Ohio Shale. The Ohio Shale correlates with the Antrim Shale in the eastern part of the Michigan basin. Recently, Wallace de Witt, Jr., stated (oral commun., 1971) that there is no reason to change his opinion set forth in the 1954 paper, namely, that the same named units — the Bedford Shale, the Berea Sandstone, and the Sunbury Shale in the Appalachian and Michigan basins — are correlatives. Paleontologic data substantiate the Mississippian age of the Sunbury Shale in the Appalachian basin. The Bedford Shale, which underlies the Berea Sandstone, has in its basal 1–5 feet a widespread fauna that has a Devonian age very close to the Famennian (Upper Devonian) — Tournaisian (Lower Mississippian) boundary. The Berea Sandstone has very few fossils; however, in the Appalachian and Michigan basins it contains (Winslow, 1962) the spore *Endosporites lacunosus*. This fossil, which is now called *Hymenozonotriletes lepidophytus* Kedo, ranges in age from Late Devonian to Early Mississippian (Streel, 1970).

Because the Antrim Shale in southeastern Michigan correlates with the Devonian Ohio Shale of the Appalachian basin and because the basal few inches of the Bedford Shale is also Devonian in age, the boundary between the Devonian and Mississippian in southeastern Michigan was arbitrarily placed at the top of the Antrim for convenience of mapping. Westward from

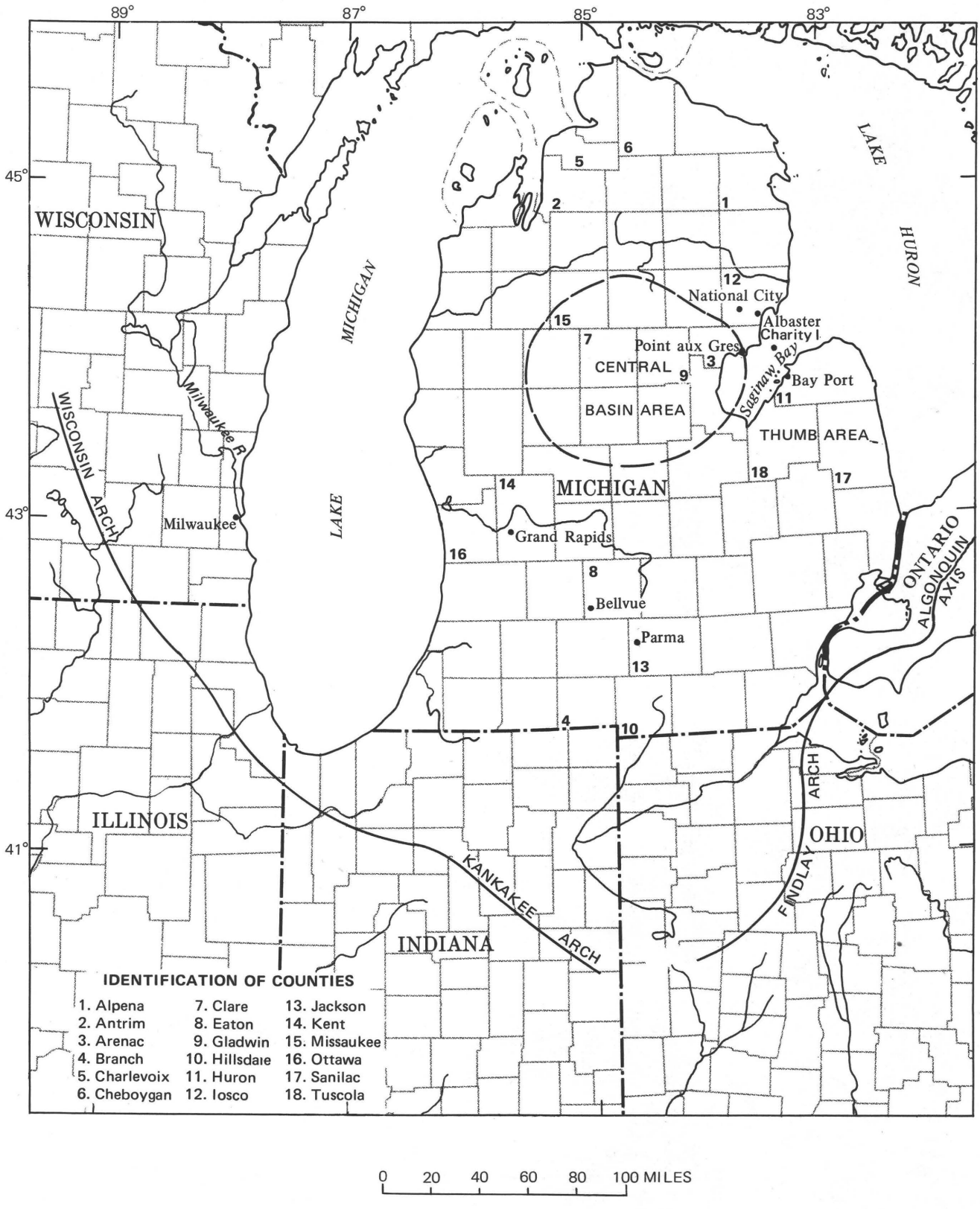

FIGURE 8. — Southern Peninsula of Michigan and surrounding area showing geologic and geographic features mentioned in text.

southeastern Michigan across the basin the black Antrim Shale thickens and grades into and intertongues with the Bedford Shale and Berea Sandstone on the east side of the basin and the Ellsworth Shale on the west side of the basin. Because of the interfingering relations, the Mississippian and Devonian boundary could not be placed closely except in southeastern Michigan, and an arbitrary thickness of 150 feet of the black shale, the thickness of the Antrim Shale in southeastern Michigan, was assumed to be Devonian and all black shale above shale of that thickness was assigned to the Mississippian.

AGE AND ASSIGNMENT OF THE *KENWOOD SHALE*, WISCONSIN

Edwards and Raasch (1921) studied the debris from a shaft sunk in the bottom of a tunnel extending more than a half mile under Lake Michigan from the lake shore at the east end of North Avenue, Milwaukee, Wis. The shaft was 132 feet deep, and it penetrated dark to black shale underlying a blue shaly limestone which is the same as the limestone in a quarry on Milwaukee River, 3 miles to the northwest. Edwards and Raasch (1921) assigned the name *Kenwood* to the black shale and determined its age to be early Late Devonian from the fossils in the black shale and in green shale interbedded with the black shale.

Pohl (1929) found in black shale in the same shaft and at the same horizon a fauna of brachiopods and other remains that indicated to him a Mississippian age. Later, on the basis of conodonts, Cooper (in Weller and others, 1948, p. 156–157) also assigned a Mississippian age to the *Kenwood Shale*. Raasch (1935) assigned to the *Kenwood* a Mississippian age, which for many years was generally accepted.

Charles Collinson of the Illinois State Geological Survey, who had also studied the black shale from the shaft under Lake Michigan, east of Milwaukee, Wis., noted (written commun., Jan. 21, 1965) that the conodonts found in the black shale indicate that it is of early Late Devonian and correlates with the Grassy Creek Shale of Missouri and Illinois, the Sweetland Creek Shale of southeastern Iowa, the Independence Shale of central Iowa, lower member of the *Blackiston Formation* of Indiana, and the lower part of the Gassaway Member of the Chattanooga of Tennessee.

Dietmar Schumacher of the Geology Department, University of Missouri (written commun., May 3, 1967), studied the *Kenwood Shale* and likewise considered it to be of Devonian age. Schumacher stated that conodont faunas obtained from the *Kenwood* contained elements of at least four of the standard Late Devonian conodont zones of Europe and that the *Kenwood* ranges in age from middle Frasnian to early Fammenian. He also stated that it correlates with the lower and middle Antrim Shale of Michigan, the upper Olentangy Shale and lower Ohio Shale of Ohio, and the Grassy Creek and lower Saverton Shales of Illinois.

The *Kenwood*, in view of these unpublished data of Collinson and Schumacher regarding its age, is excluded from this study of the Mississippian rocks of the Michigan basin.

UNITS UNDERLYING MISSISSIPPIAN

The Antrim Shale of the Michigan basin is of Devonian and Mississippian age (pls. 2, 15) and underlies the Mississippian throughout its extent in the basin. The shale in southeastern Michigan is of Late Devonian age and is equivalent to the Ohio Shale of Ohio and the New Albany Shale of Indiana. Only the basal part of the Antrim crops out in the State and that is in the northern part of the Southern Peninsula. It is 150 feet thick in the most southeastern part of the State and an arbitrary thickness of 150 feet at the base of the shale is considered to be Devonian in age and the shale above is considered to be Mississippian. The basal part of the Antrim extends beyond the Mississippian outcrop beneath the glacial drift around the basin.

INTERVAL A

FORMATIONS INCLUDED

Rocks of interval A extend throughout the Michigan basin and for the most part are the rocks of Kinderhook age. On the west side of the Southern Peninsula, interval A consists of the Mississippian part of the Antrim Shale and the Ellsworth and Sunbury Shales (pl. 15). On the east side, interval A consists of the Bedford Shale, the Berea Sandstone, and the Sunbury Shale.

The Antrim Shale is an organic-rich shale that consists of dark-gray to black hard thin-bedded brittle carbonaceous shale which, in the lower part, is interbedded with some gray shale. Dark-brown bituminous limestone concretions 2–5 feet in diameter are common near the base of the Antrim. The lowest part of the formation, which is Devonian in age, is exposed in the northwest corner of Antrim County and in Charlevoix, Cheboygan, and Alpena Counties, Mich. The Antrim is exposed at a few places in northern Indiana and northwestern Ohio (de Witt, 1960). The Antrim thickens from 150 feet in southeastern Michigan to about 650 feet in the Central Basin area. The Mississippian part of the Antrim (above the arbitrary horizon 150 ft above the base) is as much as 500 feet thick in the Central Basin area.

The Bedford Shale is a light-gray shale that is silty and sandy in the upper part and grades upward into the Berea Sandstone. Because of the gradational contact between the Bedford and Berea, the amount of shale assigned to the Bedford varies from place to place, but, generally, it is in the order of 60 feet thick. The Bedford thins westward to the Central Basin area, where it intertongues with the Antrim Shale, and in southern Michigan it merges into the Ellsworth Shale of western Michigan.

The Berea Sandstone, which is limited to the east half of the Southern Peninsula of Michigan, consists of three lithologic units. The lower unit is sandstone, which is light gray, fine grained, dolomitic, silty, and shaly, cemented with silica and dolomite, and micaceous and pyritic. The middle unit is friable, medium- to coarse-grained sandstone composed of angular quartz grains. Thin beds of shale and tightly cemented sandstone are interbedded with the friable sandstone in places. The upper unit is lithologically similar to the lower unit but is less shaly and less pyritic.

The Berea is thickest in Huron County, Mich., where there is 80 feet of the friable sandstone and 180 feet argillaceous sandstone. The sandstone thins from the Thumb area to the west, south, and north and is not present west of central Michigan.

The Berea Sandstone is overlain by the black thin-bedded Sunbury Shale which is lithologically similar to the Antrim Shale. The Sunbury is present in eastern Michigan and overlies the Ellsworth Shale in part of western Michigan, but it thins and grades into gray and greenish-gray shale of the Ellsworth in southwestern Michigan. The Sunbury is generally about 60 feet thick.

The Ellsworth Shale is a greenish-gray to gray shale that occurs in western Michigan (Bishop, 1940, p. 2150). It contains some dolomite in the upper part and is silty in places, especially in southwestern Michigan where a considerable amount of siltstone is interbedded with the greenish-gray shale. Because of the intertonguing of the greenish-gray Ellsworth Shale and the black Antrim Shale, the boundary between the two formations is difficult to establish. The Ellsworth is exposed only in a few localities in Antrim and Charlevoix Counties, Mich.

The Ellsworth Shale is 400 – 600 feet thick along the western boundary of the State, and eastward it grades entirely into black shale of the Antrim in the center of the State. Dolomite and limestone beds are common in the Ellsworth in southwestern Michigan. In Kent and Ottawa Counties, Mich., some dolomite and limestone beds are composed almost entirely of oolites. Dolomite crystals form the nuclei of some of the oolites.

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A is at the top of the Sunbury Shale and at the base of the Coldwater Shale. Where the Sunbury Shale is absent in western Michigan the top of interval A is at the top of the Ellsworth Shale.

THICKNESS TRENDS

In the area where interval A is overlain by the Coldwater Shale, the basal formation of interval B, the thickness ranges from less than 200 feet in southern Michigan to 780 feet in the northwestern part of the basin (pl. 3-A), where the interval consists entirely of the thick Ellsworth Shale. Beyond the limit of the Coldwater Shale the eroded top of interval A lies beneath the Pleistocene drift.

LITHOFACIES TRENDS

The lithofacies of interval A is dominantly shale (pl. 3-B) in the central and western part of the Michigan basin. Sandy shale forms a lesser component and is restricted to eastern Michigan. This distribution of rock types reflects the presence of relatively coarse detrital material in the Bedford Shale and Berea Sandstone which are restricted to the eastern part of the Michigan basin. On the west side of the basin, interval A is dominantly gray and greenish-gray shale. Along a narrow zone extending roughly north-south through the central part of the basin, all interval A is represented by black shale.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The relatively coarse detritus contained in the Bedford and Berea Formations was derived from the northeast (pl. 10, fig. 1) and was deposited in a large delta, the Thumb delta (Cohee, 1965), on the east side of the Michigan basin. The sediment was transported in a river system (pl. 11, fig. 1; pl. 12, fig. 1) that was a distributary to the Ontario River (Pepper and others, 1954), a river that extended across southwestern Ontario into Ohio. Gray and greenish-gray mud and silt deposited on the west side of the basin were dervied from the Wisconsin highlands to the northwest. Near the end of Kinderhook time, water in the seaway became so quiet that the black mud accumulated almost entirely across the basin.

PALEOTECTONIC IMPLICATIONS

During interval A time, the Michigan basin was a subsidiary part of the Midcontinent craton (pl. 10, fig.

1). The east side of the basin subsided more than 500 feet to accommodate the fluvial sediment transported from the northeast. These deltaic deposits did not extend westward beyond a north-south line through the central part of the Southern Peninsula.

The west side of the basin locally subsided more than 700 feet to accommodate the influx of detrital material from the Wisconsin highlands to the northwest. Southward thinning of interval A may indicate some upward movement of the Kankakee and Findlay arches.

INTERVAL B

FORMATIONS INCLUDED

The Coldwater Shale and the overlying Marshall Sandstone compose interval B (pl. 15). The Coldwater conformably overlies the Sunbury and Ellsworth Shales of interval A, and the Marshall Sandstone conformably overlies the Coldwater Shale.

In western Michigan the Coldwater Shale consists mostly of gray and bluish-gray shale containing a few thin limestone and dolomite beds. In eastern Michigan the shale is silty and sandy, and in places in the upper part of the formation some red siltstone and sandstone is present. In Tuscola, Huron, and Iosco Counties, Mich. (fig. 8), thick sandstone units occur in the shale. The Coldwater Shale, which is 500–800 feet thick in western Michigan, increases in thickness toward eastern Michigan where it is more than 1,100 feet thick in places. The Coldwater is exposed at places in Branch and Hillsdale Counties and along the shore of Lake Huron in Huron and Sanilac Counties.

A pink to red limestone and dolomite unit about 10–20 feet thick is present in the subsurface at the base of the Coldwater on the west side of the State. This basal unit is correlated with similar rocks in the subsurface in Illinois which correlate with the Rockford Limestone of Kinderhook age in Indiana. The Rockford overlies the New Albany Shale in Indiana.

The Marshall Sandstone, which overlies the Coldwater Shale, is about 300 feet thick in the outcrop area in southern Michigan and is more than 200 feet thick in western Michigan. The Marshall Sandstone consists of an upper unit, the Napoleon Sandstone Member, and a lower unit, commonly referred to as the lower member. The lower member is a fine-grained sandstone which includes siltstone in places. The thickness ranges from 155 feet in the south to 35 feet in the northern part of the Southern Peninsula. The sandstone is generally red in central and western Michigan, but in parts of eastern Michigan the red color is absent.

The Napoleon Member is a fine- to coarse-grained sandstone which is red in places, especially in western Michigan. The Napoleon, which ranges in thickness from 200 to 75 feet, is thickest in the southern part of the Southern Peninsula. The Napoleon is time transgressive northward, and in the north half of the central part of the basin, the upper part occurs in lenses and isolated bars above the main body of the Marshall. These lenses of sandstone interfinger with the limestone, dolomite, anhydrite, and shale of the lower part of the Michigan Formation.

UPPER BOUNDARY OF INTERVAL B

The upper boundary of interval B is placed at the top of the Marshall Sandstone and at the base of the Michigan Formation. At the time the uppermost part of the Marshall Sandstone was being deposited nearshore to the south, the shales and sandstones of the Michigan Formation were being deposited offshore. The lowermost part of the Michigan Formation to the south, consequently, is contemporaneous with the uppermost part of the Marshall Sandstone to the north.

THICKNESS TRENDS

The greatest thickness of interval B (pl. 4-A), more than 1,400 feet, in the area of Saginaw Bay where the Coldwater Shale is thick and contains a large amount of sandstone (fig. 9). In places in southern Michigan the interval is more than 1,300 feet thick. The unusual thickness in this area is due to thickening of the Marshall Sandstone. The thickness of interval B in the area where it is overlain by the Michigan Formation of interval C ranges from 600 feet in western Michigan to more than 1,400 feet in northeastern Michigan. Because the thickness of interval B is greater than 1,000 feet in the northern part of the Southern Peninsula, the sediment of interval B as well as that of interval A originally must have formed a fairly thick deposit over at least a part of the Canadian Shield to the north (pl. 10, figs. 1, 2).

LITHOFACIES TRENDS

The lithofacies map (pl. 4-B) shows slightly coarser detrital materials in the northeastern and northwestern parts of the basin. These materials reflect the interbedding of sandstone in the dominantly shale section. In the western and southwestern parts of the basin, some thin limestone and dolomite beds are interbedded with the gray shale of the lower part of interval B, and they affect the patterns on plate 4-B.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

During interval B (Osage) time the sea extended across the area of the Michigan basin (pl. 11, fig. 2; pl.

12, fig. 2). Much clastic material was carried into the sea primarily from the Canadian Shield to the northeast and secondarily from the Wisconsin highlands to the northwest (Stearns, 1933, p. 108). Uplift of the Canadian Shield and transport of the weathered materials from this region provided the red silt and sand of the upper part of the Coldwater Shale and lower part of the Marshall Sandstone in much of the basin.

In latest Osage time, sandstone deposition continued in the southern part of the basin while mostly limestone, dolomite, shale, and some anhydrite accumulated in the northern part of the basin. Some sandstone continued to accumulate in the northern part of the basin as lenticular bodies or sandbars in which natural gas is now found. Sandbars also formed in the southern part of the basin in latest Osage time and were of sufficient height to cut off or restrict the shallow seaway periodically so that beds of evaporite accumulated in the central and northern parts of the basin.

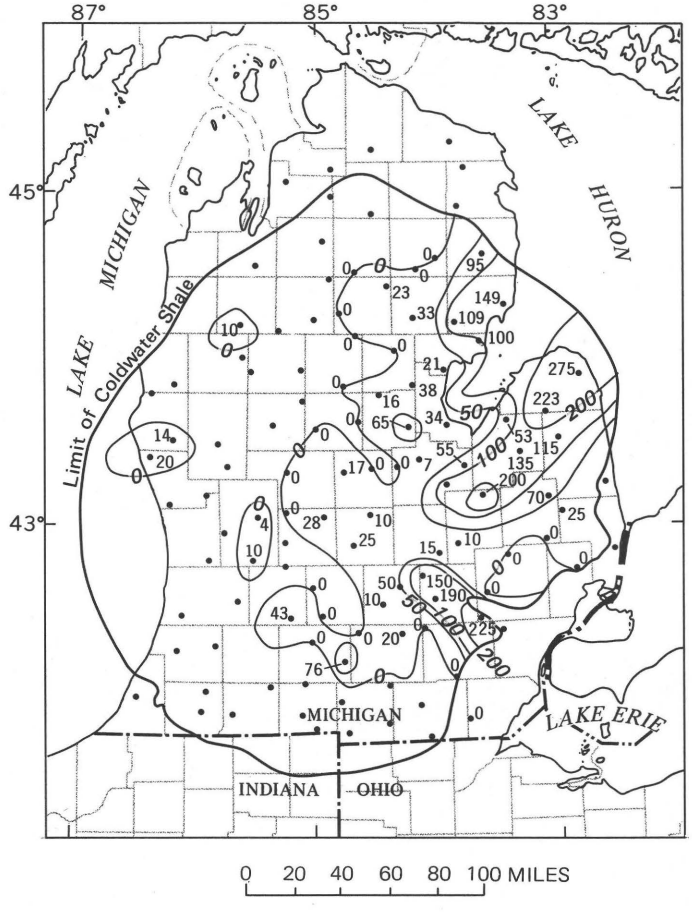

FIGURE 9. — Southern Peninsula of Michigan showing the extent of the Coldwater Shale and the cumulative thickness of sandstone in the formation. Numbers indicate thickness of sandstone in feet; unnumbered localities indicate no sandstone in the Coldwater Shale.

Limestone and dolomite beds in the Michigan Formation thicken northward, and a narrow seaway from the north may have connected the northern part of the Michigan basin with the open sea during latest Osage or earliest Meramec time, although other evidence of a northern seaway is lacking.

PALEOTECTONIC IMPLICATIONS

General subsidence of the entire Michigan basin area continued during interval B time. In at least two parts of the basin the subsidence amounted to more than 1,300 feet (pl. 10, fig. 2). Differential tectonic activity was at a minimum; the Kankakee arch was not active, and detrital material was deposited across the arch from the Michigan basin to the Illinois basin. The Findlay arch was active in the latter part of interval B time when it was elevated sufficiently to prevent the transport of detrital material from the Michigan basin across the arch into Ohio.

INTERVAL C

FORMATIONS INCLUDED

The Michigan Formation and the Bayport Limestone compose interval C, which includes rocks generally of Meramec age. The Michigan Formation overlies the Marshall Sandstone conformably. The Bayport Limestone conformably overlies the Michigan Formation but unconformably underlies the Saginaw Formation of Pennsylvanian age.

The Michigan Formation consists of gray to dark-gray and greenish-gray shale, limestone and dolomite, and thin lenses of sandstone, anhydrite, and gypsum (pl. 9-E, secs. *B–B′*, *C–C′*). The formation is exposed in Kent, Huron, and Iosco Counties, Mich. (fig. 8). Limestone is most abundant in the Michigan Formation on the north side of the basin. Gypsum beds in the Michigan are being mined at Grand Rapids and are being quarried at Alabaster and National City, Iosco County, Mich. The aggregate thickness of the gypsum in the mined areas is about 40 feet, and in the Central Basin area it is about 100 feet (chap. T, fig. 93). One 30- to 40-foot thick zone of anhydrite, which consists of three separate beds of anhydrite alternating with shale, can be traced throughout the central part of the basin by electric logs. The Michigan Formation thickens northward from less than 100 feet in Jackson County, Mich., to about 500 feet in Missaukee County, Mich.

The Bayport Limestone is light-buff and light-brown limestone that is cherty and sandy in part (pl. 9-E, secs.

B–B', *C–C'*). In many places the formation consists largely of light-gray sandstone interbedded with limestone. The sandstone is white, coarse grained, and generally poorly cemented and is most abundant in the lower part of the formation. The Bayport is generally less than 100 feet thick although it ranges in thickness from a featheredge to 120 feet; in some places, it has been removed by erosion in Late Mississippian time. The fauna of the Bayport, which is the youngest Mississippian unit in the Michigan basin, indicates correlation with the St. Louis Limestone and possibly some of the Ste. Genevieve Limestone of the Mississippi Valley (G. M. Ehlers, oral commun., 1948; Mackensie Gordon Jr., oral commun., 1964). The Bayport Limestone, which was originally called the *Point aux Gres Limestone*, is exposed in Michigan at Bay Port, Huron County, at Grand Rapids, Kent County, at Bellevue, Eaton County, near Parma, Jackson County, on Charity and other islands in Saginaw Bay, and at Point aux Gres and other localities in Arenac County.

UPPER BOUNDARY OF INTERVAL C

The top of interval C is an unconformity above which are rocks of Pennsylvanian age everywhere in the basin.

THICKNESS TRENDS

In general, interval C shows greater range in thickness (pl. 5-A) than other intervals because it was variably eroded in the latter part of Mississippian time. On some of the major folds around the basin beyond the subsurface extent of interval C, erosion has removed interval C and cut well into interval B. Interval C has its greatest thickness, a little more than 570 feet, in the central part of the basin but is almost as thick in the southwestern and northeastern parts. In the area where interval C is overlain by the Saginaw Formation at the base of the Pennsylvanian, the thickness ranges from 100 feet in southern Michigan to more than 570 feet in the northwestern part of the Central Basin area.

LITHOFACIES TRENDS

The lithofacies map of interval C (pl. 5-B) shows more diversity than the lithofacies maps of intervals A and B and reflects the more heterogenous character of rock types in interval C. In general, more fine detrital material was deposted in the eastern and southern part of the Michigan basin than in other parts, and more carbonate and evaporite was deposited in the western and northern parts of the basin.

Post-interval C erosion and irregular distribution of the numerous rock types in interval C combine to mask (pl. 5-B) some of the details of lithofacies trends of the interval. The anhydrite in the Michigan Formation thickens basinward, and the limestone and dolomite beds are most abundant in the northern part of the Central Basin area. The lithologic composition of the Bayport Limestone is variable from place to place. The sandstone, which is more abundant in the lower part of the Bayport, is interbedded upward with the limestone and dolomite, and, in some places, the formation is largely sandstone. The Bayport is the youngest unit to be affected by the Late Mississippian uplift, folding, and erosion in the Michigan basin. Consequently, it was deeply eroded, and in places it was entirely eroded from some major anticlines.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

During interval C (Meramec) time, the Michigan basin continued to receive clastic sediment which was brought into the sea from a northeastern source (pl. 10, fig. 3). Periodically, the basin was cut off from the Mississippian seaways, and anhydrite was deposited. Because of the large number of individual beds of anhydrite and gypsum in the Michigan Formation, the basin area must have had an arid climate.

In Bayport time, clean white sand was brought into the basin from the north, and limestone was deposited over the entire basin area. In places, the streams emptying into the Bayport Sea carried much silica, as indicated by the amount of chert in the limestone.

PALEOTECTONIC IMPLICATIONS

The Michigan basin was intermittently negative during most of interval C (Meramec) time and subsided locally at least 570 feet to accommodate the preserved thickness of rocks. The periodic isolation of the basin, indicated by the evaporite beds of the Michigan Formation, probably was due in part to periodic uplift of the Findlay and Kankakee arches along the southern margin of the basin.

INTERVAL D

Fossils of interval D (Chester) age have not been found in the youngest Mississippian rocks in Michigan, and it is concluded that rocks of Chester age were not deposited in Michigan.

During latest Meramec and Chester time, after deposition of the Bayport, the basin was folded and uplifted, and the Mississippian sea withdrew southward. On some large anticlinal folds, several hundred feet of Mississippian rocks were removed during this

period of erosion, and, in some places, subsequently deposited Pennsylvanian rocks rest on the Coldwater Shale of interval B. The anticlinal folds produced in the basin at this time all trend northwest to southeast with the exception of folds in southwestern Michigan where the trend is generally north and south. The major folds are 3–10 miles long and 2–4 miles wide, and they have 40–200 feet of closure. Some of the principal folds that developed at this time were undoubtedly reactivated old Precambrian structures. The uplift and folding was the most intense tectonic activity in this area since orogenic movements at the close of Early Ordovician time.

During Chester time detritus was carried from the Canadian Shield across the Michigan basin by a river system called the Michigan River (Swann, 1964). The sediment was deposited in a delta system in Indiana and Illinois. (See fig. 4, pls. 10, 11, 12.)

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

Mississippian rocks are about 1,500 feet thick near the margin of the overlying Pennsylvanian rocks, and the thickness ranges from 1,800 to 2,100 feet in the central part of the basin (pl. 7). The greatest thickness of rocks of Mississippian age was penetrated in a well in Arenac County, Mich. (fig. 8), on the north side of Saginaw Bay (pl. 7). In this well (the Basin Oil Co. Tawas 1, sec. 14, T. 20 N., R. 7 E.), there is about 450 feet of interval A, 1,480 feet of interval B, and 430 feet of interval C, for a total thickness of 2,360 feet. The zero thickness line of Mississippian rocks extends into northwestern Ohio, northern Indiana, and across the northern tip of the Southern Peninsula.

PALEOTECTONIC IMPLICATIONS

Mississippian rocks in the Michigan basin were deposited during continuous subsidence through Kinderhook, Osage, and at least a considerable part of Meramec time. The combined thicknesses of intervals A, B, and C suggest a somewhat irregular pattern of subsidence; parts of the basin subsided more than 2,300 and 2,100 feet, whereas most of the central part of the basin subsided only about 1,800 feet (pl. 7). The details of this irregular pattern of subsidence are uncertain because of pre-Pennsylvanian erosion (pl. 9-E, secs. *B–B'*, *C–C'*).

No detrital materials were derived from the immediate margins of the basin. Positive structures, such as the Kankakee and Findlay arches, served only briefly as barriers to transport of sediment; they were not sufficiently uplifted to serve as sediment sources.

In late Meramec or Chester time the region was uplifted and was eroded and was subsequently covered by Lower Pennsylvanian rocks. The erosion was accomplished by generally southward-flowing streams, and sediment was transported across the area to sites of deposition farther south.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Over an area of about 11,500 square miles in the Southern Peninsula of Michigan, the Saginaw Formation of Pennsylvanian (Morrow) age rests on the eroded top of the Mississippian sequence (pl. 8). The Saginaw Formation consists of lenticular sandstone, gray, dark-gray, and black shale, and some argillaceous limestone. In the northwestern part of the area of Pennsylvanian rocks, the lower part of the Saginaw includes red and green shale and some buff limestone. In the southern part a lenticular coarse-grained sandstone, the Parma Sandstone Member of the Saginaw, rests on the Mississippian.

Red-bed deposits of Jurassic age that overlie Pennsylvanian rocks in the Michigan basin extend beyond the boundary of the Pennsylvanian and rest on Mississippian strata in small areas on the west and north sides of the basin. The sequence of red beds consists of poorly consolidated deposits of clay, shale, sandstone, and some gypsum. The age of these deposits has been in question, but spore studies of drill samples, by Aureal Cross and his students of Michigan State University (oral commun., 1965) have indicated a Late Jurassic (Kimmeridgian) age for these rocks.

Pleistocene drift (not shown on pl. 8), which blankets Michigan generally to a depth of 200–300 feet but locally to 1,000 feet, rests on the eroded Mississippian rocks beyond limits of the Pennsylvanian rocks and the Jurassic red-bed sequence.

SUMMARY

Deposition of detrital material in the Michigan basin continued from late Devonian time into Kinderhook (Mississippian) time. Following uplift of the Canadian Shield to the northeast and deposition of the Bedford and Berea Formations in the Thumb delta in the eastern part of the Southern Peninsula and uplift of the Wisconsin highlands and deposition of the Ellsworth Shale on the west side of the State, very quiet conditions prevailed and little material was transported into the basin. Under these quiet conditions, near the close

of Kinderhook time, the black Sunbury Shale was deposited over the basin with the exception of the most western part.

Further uplift of the Canadian Shield to the northeast in Osage time provided detrital material for the thick Coldwater Shale. Coarse sandstone was deposited in the Coldwater in eastern Michigan, and thin beds of limestone were deposited in the shale in western Michigan. Near the close of Osage time, the Canadian Shield was further uplifted to provide the sand for the Marshall Sandstone that was deposited across the basin. The Wisconsin highlands were uplifted also and provided sand to the Marshall Sea in western Michigan.

In Meramec time, subsidence of the basin was sufficient for the shallow sea to periodically receive mud, silt, and some sand washed in from the adjoining landmass. Also, the sea was restricted periodically, and calcium carbonate and calcium sulphate were precipitated in an evaporating environment. Periodic uplift of the Kankakee and Findlay arches probably caused the restriction of the early Meramec Sea. In the latter part of Meramec time, following deposition of the Michigan Formation, the basin continued to subside and the Bayport Limestone was deposited. Some clean sand was carried into the sea at that time; in places the Bayport includes lenticular masses of sandstone with the limestone.

At the close of Meramec time, the sea withdrew for the remainder of Mississippian time. Folding and erosion took place during Chester time. The Kankakee and Findlay arches south of the basin were elevated. As much as several hundred feet of Mississippian rocks was removed from some of the major anticlines in the basin. The upper part of the Coldwater Shale, the Marshall Sandstone, the Michigan Formation, and the Bayport Limestone were eroded from the top of folds in the vicinity of Saginaw Bay, and in Arenac, Bay, and Tuscola Counties, Mich.

The folds throughout the basin generally trend northwestward except in southwestern Michigan where they trend almost northward. They have closure of 30–200 feet and are generally broad folds, some of which extend 3–10 miles.

During Chester time clastic sediments were carried from the Canadian Shield across the Michigan basin by a river system called the Michigan River (Swann, 1964) to a delta system in Indiana and Illinois. The Canadian Shield continued to furnish clastic sediments for the Michigan basin during Pennsylvanian time (Siever and Potter, 1956).

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Eastern Interior Basin Region

By EDWARD G. SABLE

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–E

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

EASTERN INTERIOR BASIN REGION

By EDWARD G. SABLE

ABSTRACT

Mississippian history in the Eastern Interior basin, Missouri, west-central Tennessee, and parts of eastern Kentucky is largely one of marine deposition in shallow cratonic basins and on shelves and platforms. Strata here designated as intervals A, B, C, and D broadly correspond to the Kinderhook, Osage, Meramec, and Chester Series, respectively. Maximum thicknesses of more than 3,200 feet of Mississippian strata were deposited in the Eastern Interior basin, about 700–900 feet on margins of the Appalachian basin, more than 400 feet in the Forest City basin, and more than 400 feet in southwestern Missouri; thinner deposits accumulated on adjacent arches and uplifts. Positive elements from which Mississippian strata have been largely eroded and which were alternately submergent and emergent during the Mississippian were the Cincinnati arch, the Wisconsin arch and La Salle anticlinal belt, the Ozark uplift, and the Kankakee arch. Variations in depositional thicknesses result from differential tectonic movements and also from slight topographic relief which was caused by uneven accumulation of sediment. Lithologic analysis indicates a wide variety of complexly related detrital, chemical, and biologically derived sediments. Crustal instability within parts of the region is recorded by unconformities, but widespread hiatuses induced by epeirogenic movements do not exist within rocks of the system.

Carbonate rocks, both chemically precipitated and biologically derived, dominate the Mississippian rocks in the region. Carbonate strata, overlapping from the west, covered extensive areas by the end of interval A time. During interval B, carbonates, which were initially restricted to western areas, again spread eastward as deposition of terrigenous detritus waned and reached their maximum extent during interval C. Although areally more restricted in interval D, carbonate rocks periodically accumulated over large areas in and beyond the Eastern Interior basin and eastern Kentucky. Evaporites deposited during mid-interval C time marked an episode of restricted circulation probably related to tectonic movements.

Land-derived detrital rocks constitute about one-third of the total preserved Mississippian succession. In order of importance based on the volume of sediment contributed, the source areas were highlands northeast and east of the present Appalachian Mountains, the eastern Canadian Shield, parts of the Transcontinental arch in Minnesota and Wisconsin, the Ozark uplift, and the Cincinnati arch. The presence of southern source areas is uncertain. Large volumes of detritals from northeastern sources were carried into the eastern part of the region by major river systems and accumulated as deltaic complexes during intervals B and D times; lesser amounts came from north and northwest sources during interval A and possibly intervals B, C, and D, and small volumes were shed from Ozark uplift sources during interval A and possibly intervals C and D times. The Cincinnati arch, chiefly a barrier to sediment dispersal, probably contributed small amounts of detritus during early interval B and during interval C. Siliceous and cherty rocks are abundant in interval B rocks in the southern part of the region; southern or eastern sources may have contributed clay-size detrital silica in these areas. Sources in northern Canada may have contributed to interval D detrital rocks.

Shallow-water to very shallow-water neritic marine-sedimentary environments predominated over large areas of the region during the Mississippian, but deepwater neritic environments were present in the central and southern parts of the region during interval B and probably during interval C.

Seas generally opened and deepened southward, and western connections to widespread Mississippian seas across and north of the Ozark uplift and eastern connections to Appalachian Mountains areas through the Cumberland saddle were open during much of the Mississippian. Deposits of intervals A, B, and C also indicate marine connections to a sea in the Michigan basin. Littoral and continental environments characterized large areas periodically during interval D. In general, low-energy environments are shown by the fine grain size of interval A rocks, and progressively higher energies are indicated by coarser and better sorted rocks in interval B and early interval C, except in the deeper parts of the basins. Mid-interval C deposits show low-energy conditions; alternations of high- and low-energy environments characterize interval D. Mississippian rocks attest to a mild climatic regimen during the period with aridity in mid-interval C time, and possibly a wet climate periodically during interval D.

Structural features having northeast and northwest structural trends, mostly inherited from Late Devonian time, characterized the Mississippian tectonic framework; northeast-trending negative elements responded most actively to tectonic stresses. Positive elements — the Ozark uplift, the Cincinnati arch, and the La Salle anticlinal belt — strongly influenced Mississippian depositional patterns. Crosscutting linear features, at present expressed as major fault zones, may also have been active faultlines or hinge lines during the Mississippian.

Tectonic movements affecting the region during Mississippian time included relative uplift and partial emergency of the Transcontinental arch, and corresponding subsidence of a subparallel northeast-trending trough across the central part of the region following interval A time. A regional southwest-dipping paleoslope developed during interval B and persisted during succeeding intervals. Renewed relative uplift of major positive elements restricted seas during mid-interval C time. At about the close of interval D time, a major episode of southward tilting and general emergence took place, preceding deposition of Pennsylvanian sediments. These structural movements were accompanied by widespread channel cutting and by beveling of strata near basin margins and on some gentle uplifts within the basin.

REGION DEFINED

This chapter deals with Mississippian rocks in Missouri and Illinois, western and southern Indiana, west-central Tennessee, and all of Kentucky except the eastern subsurface and southeastern counties. The chapter was initially written prior to 1970; a few pertinent additions have been added to approximately 1973.

The centrally located Eastern Interior basin (fig. 10) is the major negative element of the region and encompasses much of Illinois, Indiana, and Kentucky, and parts of Missouri and west-central Tennessee. Portions of other basins included in the region are the western margin of the Appalachian basin in eastern Kentucky, eastern part of the Forest City basin in northwestern Missouri, and shallow negative features in southwestern Missouri. Major positive cratonic features which serve to outline the Eastern Interior basin are the Cincinnati arch — including the Jessamine dome, Cumberland saddle, and Nashville dome — the Kankakee arch, the Ozark uplift, and Pascola arch. Lesser features which bound the western margin of the basin are the Mississippi River arch, the Lincoln and Dupo anticlines, and the Ste. Genevieve fault. The northern margin is bounded by east-trending anticlinal structures which appear to be related to the southern part of the Wisconsin arch.

The faults shown in figure 10 moved largely after deposition of Mississippian rocks, the basins, domes, arches, anticlines, and synclines are syndepositional, as well as postdepositional, features that influenced late Paleozoic deposition.

The north-northwesterly trending Eastern Interior basin is interrupted by two major intrabasinal structural features, the La Salle anticlinal belt and the Rough Creek fault zone. These and several lesser structural features in Illinois and Kentucky delineate two main negative elements within the major basin, the Fairfield basin in southeastern Illinois and the Moorman syncline in western Kentucky. The Eastern Interior basin contains structural features which are intermediate in trend to the northeast strike of the Appalachian Mountains belt and the largely northwest trends in the Ozark uplift and Pascola arch. Structural relief of the Eastern Interior basin is greater than 12,000 feet. Domes and cryptoexplosive structures locally interrupt major structural trends in several areas.

Many workers use the terms "Illinois basin" or "Illinois-Indiana-Kentucky basin" as synonymous with the Eastern Interior basin. Others restrict the term "Illinois basin" to the negative element between the Ozark uplift and the La Salle anticlinal belt and call the areas east and south of the La Salle belt the Indiana–West Kentucky basin. Synonyms which have also been applied to the basin in western Kentucky include the West Kentucky coal basin or Western coal basin to distinguish it from the marginal parts of the Appalachian or Eastern Coal basin in eastern Kentucky.

MISSISSIPPIAN ROCKS AND CLASSIFICATION

Mississippian rocks crop out near major positive structures or along basin margins (pl. 1). Because of a scarcity of outcrops and an extensive cover of Pennsylvanian strata, most of the data in this chapter were synthesized from subsurface information.

The region was almost continuously inundated by Mississippian epeiric seas in which many kinds of detrital and chemical sedimentary rocks were deposited. Terrigenous clastics, mostly deposited in deltas, were derived from major source areas east, northeast, and north of the region through much of Mississippian time, and lesser source areas, such as the Ozark uplift, contributed sediments intermittently. Carbonate deposits formed in areas adjoining lobes of detrital sediments or, during times when little detritus was being deposited, throughout most of the region. Arid climate and restricted circulation during middle Mississippian time produced fairly widespread evaporite beds. The Upper Mississippian sequence is characterized by rhythmic alternations of carbonate-dominated and terrigenous detrital-dominated strata.

The Eastern Interior basin region contains the standard type region for the Mississippian System in North America. Series divisions of the Mississippian are named for exposures near Kinderhook in western Illinois, the Osage River in western Missouri, the Meramec River in southeastern Missouri, and the Chester district in southwestern Illinois. The original term "Mississippi group" was proposed by Winchell (1869, p. 79), and revived as "Mississippian" by Williams (1891, p. 135) to replace the European term "Subcarboniferous." The term "Mississippian Period" was first formally used by Chamberlin and Salisbury (1906, p. 496), and the corresponding term "Mississippian System" has gained wide acceptance in the United States (Weller and others, 1948, p. 97). Several twofold to fivefold series subdivisions of the Mississippian System have been proposed in the Eastern Interior region. The standard provincial series now recognized by the U.S. Geological Survey and several State surveys in the Eastern Interior basin and adjoining Missouri areas include the terms Kinderhook (Meek and Worthen, 1861), Osage (Branner, 1888), Meramec (Ulrich, 1904), and Chester (Worthen, 1860, p. 312-313). These names were first consolidated into the Mississippian by Ulrich in 1904 and further elaborated

by him in 1905. The Illinois State Geological Survey currently uses a threefold series subdivision — the Kinderhookian, Valmeyeran (combined Osagean and Meramecian), and Chesterian Series. The U.S. Geological Survey uses a twofold designation — Lower Mississippian Series (Kinderhook and Osage equivalents) and Upper Mississippian Series (Meramec and Chester equivalents).

Classification of Mississippian and adjacent strata in the region, other than the conventional rock-stratigraphic (group, formation, member) and time-stratigraphic (system, series, stage) designations include informal classifications such as megagroup (Swann and Willman, 1961) and sequence (Sloss and others, 1949; Sloss, 1963) (fig. 11). Although the megagroup and sequence concepts are valuable in establishing a general framework within which Mississippian events can be placed, this chapter utilizes the formal and local rock and time divisions.

PALEOGEOLOGY

DEVONIAN-MISSISSIPPIAN BOUNDARY

MISSISSIPPI RIVER VALLEY AND ADJACENT AREAS

Uncertainty regarding the position of the Devonian-Mississippian boundary in North American has, in the past, stemmed from (1) lack of agreement concerning placement of the Devonian-Carboniferous boundary in Europe, (2) the fact that the historical Devonian standard section for the United States, in New York State, is distant from the Mississippi Valley type Mississippian sections, and (3) lack of agreement on stratigraphic relationships and faunal correlations of sections spanning the boundary in the Mississippi Valley itself. Criteria for the Devonian-Carboniferous boundary placements in Eurasia were reviewed by Mamet (1967) and for the Devonian-Mississippian boundary along the cratonic margin of the United States by Gutschick and Moreman (1967). Early concepts requiring diastrophism as the basis for major time subdivisions are now considered obsolete, and the faunal elements across the Devonian-Carboniferous and Devonian-Mississippian boundaries in the principal sections in Europe and the United States are generally, but not universally, considered to be transitional.

Besides the ammonoid, brachiopod, and coral macrofaunas that have been used in placement of the boundaries, conodonts and spores are currently considered to be pertinent to the problem. Conodonts seem to be the most reliable faunal elements for separating Devonian and Mississippian rocks. Gutschick and Moreman (1967) cited important contributions to conodont zonation in the United States, including the pioneering efforts in the Upper Mississippi Valley and Missouri by Branson and Mehl (1933a, b, 1941) and in Indiana by Huddle (1934). Assignments of units in the Mississippi Valley in the present report follow the classification of Scott and Collinson (1961) and Collinson (1961). Correlations with European conodont assemblages were discussed by Collinson, Scott, and Rexroad (1962). Further refinements of the Devonian-Carboniferous boundary in the United States and Europe using conodonts and spores are reported by Sandberg, Streel, and Scott (1972), who discussed conodont zonation and spore assemblages in the Upper Mississippi Valley.

In most areas in the eastern interior of the United States, units of dark shale with sparse macrofauna separate undoubted rocks of the Devonian and Mississippian Systems. These shale units, such as the Hannibal, Saverton, Grassy Creek, New Albany, Chattanooga, and Ohio Shales, have been variously placed in one system or the other, or the systemic boundary has been shown to lie within them. Lack of complete agreement on the relative ages of fauna in limestone units which separate the shale units, such as the Louisiana Limestone, the *"Glen Park" Formation* of Illinois, and the *McCraney Limestone*, has also added to the uncertainty of boundary placement.

The Devonian-Mississippian boundary in the Mississippi Valley used in this report is that of Collinson, Scott, and Rexroad (1962) and corresponds with the top of the Louisiana Limestone and Saverton Shale. These units, which probably intertongue, and the underlying Grassy Creek Shale are assigned to the Devonian. The *"Glen Park" Formation*,[1] the Hannibal Shale, and the Chouteau and *McCraney* Limestones are correspondingly placed in the Mississippian. Gutschick and Rodriguez (1967) placed the boundary stratigraphically lower and considered the Louisiana and upper beds of the Saverton to be of Mississippian age. Sandberg, Streel, and Scott (1972) indicated a boundary very similar to that used in this report.

At some places in Missouri, as reported by Branson and Mehl (1933a) and Mehl (1961), a faunal gap representing the latest Devonian occurs between the Saverton and overlying Hannibal. They suggested on this basis that a hiatus separates Devonian and Mississippian strata regionally. Complete sequences, however, were reported from the Mississippi Valley by Scott and Collinson (1961, p. 117) who described a gradual transition of faunas across the Devonian-Mississippian boundary there, although they recognized some local disconformities (Collinson, 1961, fig. 3).

[1] Used in quotation marks by the Illinois State Geological Survey to acknowledge that it is a different rock unit than the Glen Park in Missouri, which is of Devonian age.

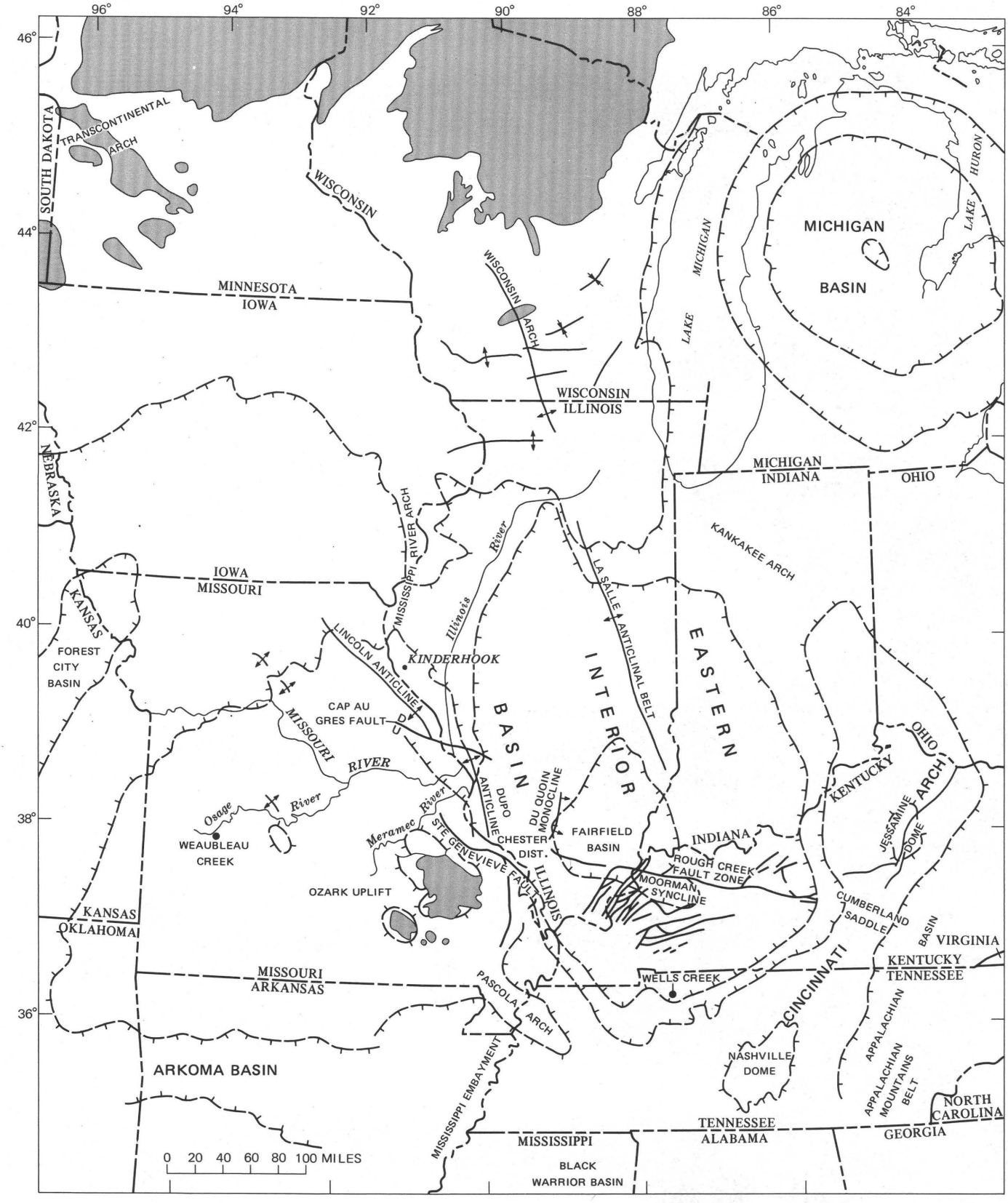
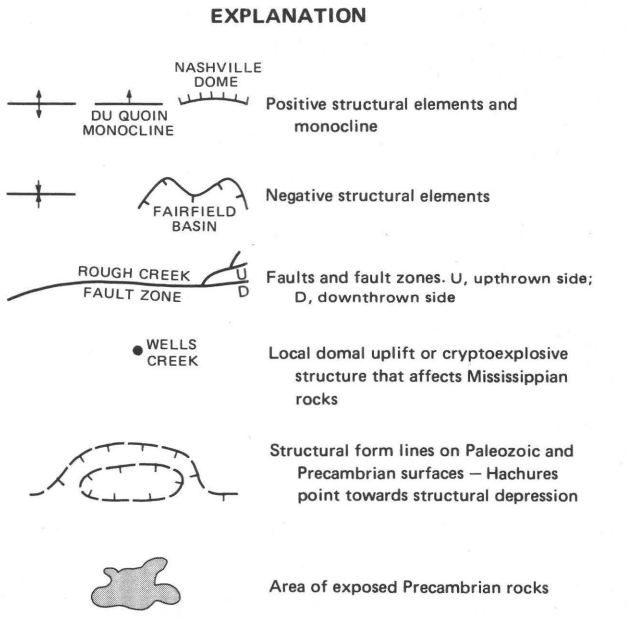


FIGURE 10.—Present generalized structural elements of the Eastern Interior basin and adjoining areas. Sources of data: Cohee and others (1961) and Willman and others (1967).



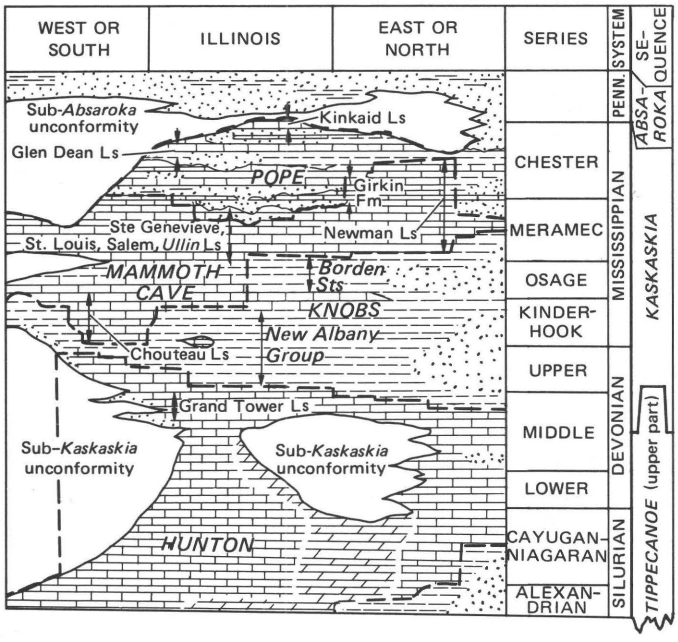


FIGURE 11. — Diagrammatic restored section showing pre-Pennsylvanian –post-Ordovician rock- and time-stratigraphic classifications in the Eastern Interior basin region. Heavy dashed line separates the *Pope, Mammoth Cave, Knobs,* and *Hunton megagroups.* Modified from Swann and Willman (1961); reprinted with the permission of the American Association of Petroleum Geologists.

Delineating the Devonian-Mississippian boundary in western Illinois, southeastern Iowa, and eastern Missouri is difficult, except in areas where conodont information is available, because of the following factors:

1. The lithologically distinctive Louisiana (pl. 2) and "*Glen Park*" Formations, which bracket the boundary in part of the Mississippi River valley, are recognizable only within a narrow belt connecting west-central Illinois with southeastern Iowa (fig. 4). In other parts of western Illinois and northeastern Missouri, however, the Devonian-Mississippian boundary is at the base of the Hannibal Shale which overlies the lithologically similar Saverton Shale. It is difficult to separate the Hannibal and Saverton in the subsurface of western Illinois and southeastern Iowa, and locations of the systemic boundary and zero limits for the Hannibal, which constitutes most of Mississippian interval A in these areas (pl. 3-A), are uncertain.
2. In northeastern Missouri correlations based on widely separated wells indicate that a unit called the Louisiana Limestone, and generally considered Devonian in age, may be in part or wholly equivalent to the Chouteau and *McCraney* Limestones or to the "*Glen Park*" *Formation,* all of Mississippian age in Illinois (J. W. Koenig, oral commun., 1963).
3. East and south of the limits of the Louisiana Limestone and the "*Glen Park*" *Formation* in Illinois, a shale underlying the Mississippian Chouteau Limestone and overlying the Devonian Grassy Creek Shale represents both Saverton and Hannibal equivalents (Lineback, 1964, fig. 4, p. 12–15; fig. 5, p. 16–18). Although the shale is relatively thin, the positioning of a systemic boundary within it strongly affects thickness and lithofacies designation for the thin Mississippian interval A rocks there.

Interval B strata, the Burlington and Fern Glen Limestones, rest unconformably on Devonian rocks in western Illinois and on rocks ranging in age from Ordovician to Late Devonian in southwestern Illinois and southeastern Missouri. Interval B rocks also overlie Devonian and older rocks on the northern and western margins of the Ozark region. In the east-central and central Missouri outcrop, Burlington and *Pierson*(?) Limestones overlie Devonian rocks or the Devonian-Mississippian Sulphur Springs Formation. Interval A strata appear to be locally absent from the outcrop of southwestern Missouri (Spreng, 1961, p. 53, 60) and the interval B formations, the *Pierson,* Reeds Spring, *Elsey,* and the Grand Falls Chert Member of the Boone Formation rest on Ordovician, Devonian(?), or unassigned Devonian-Mississippian rocks. These occurrences result from overlap by interval B sediments on areas of nondeposition or truncation of interval A sediments.

AREAS EAST OF THE MISSISSIPPI RIVER VALLEY

In Indiana, Kentucky, and Tennessee, the Devonian-Mississippian boundary can be determined fairly closely. The uppermost part of the New Albany Shale in its type locality of southern Indiana spans the boundary (Huddle, 1934; Campbell, 1946; Lineback, 1964). Members of the New Albany as recognized by Lineback (1968a) extend westward in the subsurface into Illinois and Kentucky. The uppermost member in Indiana, the *Clegg Creek Member*, spans the systemic boundary and grades laterally into the undivided Hannibal and Saverton Shales in Illinois, and is correlated with part of the Ellsworth Shale of northern Indiana and Michigan.

The position of the Devonian-Mississippian boundary in northeastern Kentucky is not precisely known but appears to lie within the basal few feet of the Bedford Shale and Berea Sandstone, a detrital wedge which separates the Mississippian Sunbury Shale and the Devonian Ohio Shale. The basal part of the Bedford Shale was reported by Hass (1947, p. 135–136) to be "very close to the Devonian-Mississippian boundary"; Scott and Collinson (1961, p. 118) suggested that conodonts in the basal Bedford are like those in the Louisiana Limestone. Southwestward, the uppermost few inches of the New Albany Shale which represent the distal edge of the Sunbury Shale are reported to be of Kinderhook age as far southwest as southeast-central Kentucky (Hass, 1947). Father south, in south-central Kentucky, beds directly overlying the Chattanooga contain Kinderhook conodonts in some localities, but late Devonian (*to*VI) forms are recorded in others (J. W. Huddle, written commun., 1964).

In western and southern Kentucky, conodont data from the New Albany Shale, and from its correlative the Chattanooga Shale, are sparse. Physical relationships of the New Albany and Chattanooga with the overlying Maury Formation and its lithic equivalents indicate apparent regional conformity, although a rather abrupt lithologic change from dark-gray shale to greenish-gray glauconitic shale occurs at the contact. Local unconformity is shown by scour channels in the uppermost New Albany. These channels are filled with siltstone of the Borden Formation at the base of interval B (R. C. Kepferle, oral commun., 1967). Cross and Hoskins (1951) cited paleobotanical data for assigning a Mississippian age to part of the New Albany in west-central Kentucky and Indiana. A hiatus between greenish-gray shale (lithologically similar to the Maury) in the basal part of the Borden and dark-gray shale in the subjacent New Albany Shale in west-central Kentucky is suggested by conodonts of probable late Burlington (Osage) age in the basal Borden (Rexroad and Scott, 1964). Age of uppermost New Albany beds in this area, however, is not known. Conodont data indicate that the entire Chattanooga is Late Devonian in age in Tennessee (Hass, 1956). The overlying thin Maury Formation is of Kinderhook age in most areas although, locally, Late Devonian forms occur in its basal bed and early Osage forms are present in its upper part. Although there is little physical evidence of unconformity between the Maury and Chattanooga, the change in conodont assemblages may indicate an obscure hiatus in deposition (Hass, 1956, p. 23). Scott and Collinson (1961, p. 119) suggested correlation of the uppermost part of the Gassaway Member of the Chattanooga Shale in Tennessee with the Louisiana Limestone in Illinois.

In summation, a widespread hiatus between the two systems (Mehl, 1960, 1961) is not generally recognized in the eastern part of the region; if present it is an obscure break that has not been clearly delineated by faunal information or physical relationships.

AREAS WEST OF THE MISSISSIPPI RIVER VALLEY

In the subsurface of northwestern Missouri fossils useful for dating are not known from the *Kinderhook shale*, and correlations with units of known age in adjoining States are the bases for defining the Devonian-Mississippian boundary. The upper part of the *Kinderhook shale*, including red mudstone and a hematitic oolite horizon, is correlated with the *Boice Shale* (Mississippian) of Nebraska (Reed, 1946; Koenig, 1961b, p. 48–49). The lower part of the *Kinderhook shale* is arbitrarily assigned to the Devonian to agree with the tentative age assignment of the laterally continuous rocks in Nebraska (Carlson, 1963, p. 37). In some Missouri well records, several beds of hematitic oolite and red shale are reported in the shale unit; the base of the uppermost oolitic bed is arbitrarily used to demark the boundary in these wells.

The Chattanooga Shale of southwestern Missouri is currently designated as Devonian or Mississippian by the Missouri Geological Survey (Koenig, 1961b, p. 43–44). Mehl (1961, p. 92) revived the name *Noel Shale* for this unit and assigned it to the Upper Devonian.

Evidence for post-Devonian–pre-Mississippian erosion north of Missouri consists of (1) the presence of the thin basal hematitic beds and red shales in the part of the *Kinderhook shale* equivalent to the *Boice Shale* (Carlson, 1963, p. 43), (2) anomalous thinning of Upper Devonian rocks, and (3) a red soil zone at the top of the Devonian in Iowa (Harris, 1947). The hiatus suggested by such evidence was short compared to the earlier, Middle Devonian hiatus.

SUMMARY

On the basis of current conodont information, all or the largest part of the black shale units are Devonian in

age (pl. 15). The systemic boundary apparently lies somewhat above the Ohio Shale in Ohio, but within the uppermost beds of the New Albany Shale in Indiana. The Chattanooga Shale of Tennessee and Kentucky and the Grassy Creek and Chattanooga (or *Noel*) Shales of southwestern Missouri are believed to be the uppermost Devonian formations, as are formations with different gross lithologic aspect — the Saverton Shale and Louisiana Limestone. In northwestern Indiana, a significant thickness of beds in the upper part of the New Albany is considered to be Mississippian in age (p. 68). The systemic boundary where it lies between the Saverton and overlying Hannibal is vague or unrecognizable in many subsurface sections in central and western Illinois and Missouri where the Louisiana Limestone is absent, and the boundary between them is picked arbitrarily in these sections. The Sulphur Springs Formation is included with pre-Mississippian rocks as are sandstone units which have been called the Sylamore Sandstone Member of the Chattanooga in Oklahoma, Bushberg Sandstone Member of the Sulphur Springs Formation, or *Bachelor Formation* in Missouri.[2] In northwestern Missouri, the boundary is placed within the *Kinderhook shale.* Unconformities between Mississippian and older rocks are common in the western part of the region; evidence of hiatus between Devonian and Mississippian rocks is known at only scattered localities in the eastern part.

UNITS UNDERLYING MISSISSIPPIAN

Upper Devonian rocks underlie the Lower Mississippian strata in the Eastern Interior basin region (pl. 2) everywhere except marginal to the Ozark uplift and in the area between the Pascola arch and Nashville dome. In these areas the rocks below Mississippian strata range in age from Early Ordovician to Middle Devonian. The pattern of distribution of rocks under the Mississippian is largely the result of Early and Middle Devonian erosion in much of the region, as well as less intense erosion prior to and during Early Mississippian time in the Ozark part of the region.

Mississippian rocks directly overlie units of Silurian and Devonian ages (Decatur Limestone and Ross Formation) in small areas between the Pascola arch and Nashville dome in west-central Tennessee (pl. 2). According to Conant and Swanson (1961, p. 53), the areas were islands during deposition of the Late Devonian Chattanooga Shale.

West of the Eastern Interior basin, Lower Mississippian strata of intervals A or B overlie Lower Ordovician to Silurian rocks in areas around the Ozark uplift. In southwestern Illinois and adjoining Missouri, Mississippian interval B rocks overlap from east to west on the truncated edges of Lower Devonian Clear Creek Formation, Silurian Edgewood, Bainbridge, and Kankakee Limestones, Upper Ordovician Fernvale Limestone and Maquoketa Shale, and Middle Ordovician Decorah Shale and Kimmswick Limestone. Within the Ozark region of Missouri, small outliers of Mississippian interval B rocks directly overlie Lower Ordovician Roubidoux and Jefferson City Dolomites (not shown on pl. 2). Ordovician rocks underlie Mississippian strata in a broad area of western and southwestern Missouri and include beds of Early Ordovician (Jefferson City and Cotter Dolomites) and Middle Ordovician (St. Peter Sandstone) age.

Middle and Upper Devonian rocks make up the upper part of the *Hunton megagroup* and the lower part of the *Knobs megagroup* (fig. 11). The Middle Devonian strata are largely carbonate rocks, but their upper beds grade eastward into detrital rocks of the *Knobs megagroup.* Upper Devonian rocks are mostly fine-grained detrital rocks of the *Knobs megagroup* but include minor carbonate-dominated units such as the Louisiana Limestone in the western part of the region.

Middle and Upper Devonian silty mudstone of the *Knobs megagroup* directly underlie Mississippian rocks with apparent conformity in most of the region northwestward from eastern Kentucky and west-central Tennessee to beyond northern Missouri. The mudstones include the widespread Devonian-Mississippian black shales and associated gray shales: the Chattanooga, most of the New Albany, the Grassy Creek, and the Saverton Shales, and the lower part of the *Kinderhook shale.*

Thin sandstone units — the Bushberg Sandstone Member of the Sulphur Springs Formation and the Sylamore and Hardin Sandstone Members of the Chattanooga Shale — underlie the black shales and unconformably overlie older Paleozoic rocks in many areas. Rocks referred to the Sylamore Sandstone Member in western Missouri are more widely distributed than shown on plage 2. Where Devonian black shales are absent, in the Ozark region, the sandstones are overlain by the Mississippian *Bachelor Formation* which, according to Mehl (1961, p. 91–92), has been mistakenly correlated with Devonian sandstones. Because the *Bachelor* is not differentiated in most described sections, it is not included in the accompanying maps. Discontinuous sandy clay, referred to as the detrital zone by Missouri geologists, however, is recorded beneath known Mississippian rocks at a few localities in west-central Missouri and may represent the *Bachelor.* The detrital zone and the other detrital units are collectively grouped with pre-Mississippian rocks on the

[2] The Sulphur Springs is considered to be a group by the Missouri Division of Geological Survey and Water Resources (Koenig, 1961a, p. 43), with the Glen Park and the Bushberg as the only proper formations of the group.

geologic map (pl. 2), but their distribution is more widespread than shown.

The Louisiana Limestone and partly equivalent Saverton Shale are the youngest Devonian rocks in western Illinois and northeastern Missouri (Collinson, 1961, p. 103–104). The Louisiana (fig. 13) lies in a narrow arcuate belt extending from Iowa across northeastern Missouri to central Illinois, and the bulk of the Saverton Shale lies within and north of this belt. This belt was later a belt of subsidence in Mississippian interval A time (pl. 3-A).

In east-central Missouri, the Sulphur Springs Formation (fig. 13; pl. 2) consists of limestone and mudstone (Glen Park Limestone Member) and overlying sandstone (Bushberg Sandstone Member). The Sulphur Springs overlies Ordovician and Devonian rocks and underlies several Mississippian units. The component units of the Sulphur Springs are considered to be mostly of Late Devonian age by Mehl (1961, p. 92). The Sulphur Springs is excluded here from the Mississippian.

The Upper Devonian Snyder Creek Shale, which is dominantly mudstone with scattered beds of limestone and sandstone, underlies Mississippian rocks of intervals A and B in central Missouri. It is of small areal distribution and is mapped with the underlying Callaway Limestone to comprise the rocks labeled as Upper and Middle Devonian on plate 2.

PALEOTECTONIC IMPLICATIONS

Tectonic elements that influenced depositional and erosional patterns of Middle and Upper Devonian sediments within the Eastern Interior basin were discussed by Workman and Gillette (1956), Meents and Swann (1965), and Collinson and others (1967); such elements in Tennessee areas were discussed by Conant and Swanson (1961).

After widespread uplift and truncation of older units, Middle Devonian carbonate units were deposited. Positive structures along which Middle Devonian unconformities are known, or which separate different rock assemblages, include the Cincinnati arch, Ozark uplift and adjoining Sparta shelf, Kankakee arch, and

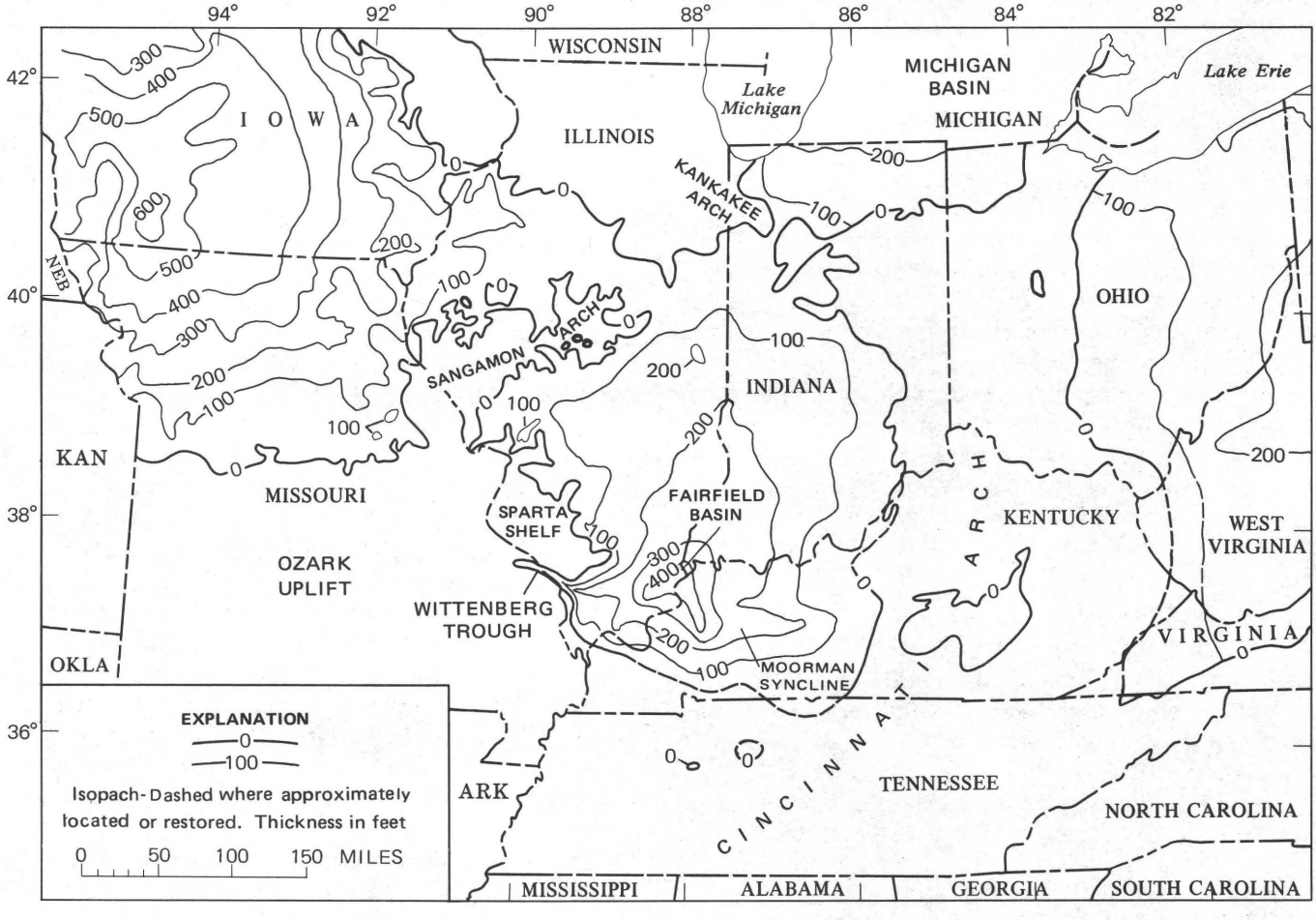


FIGURE 12.—Thickness of Middle Devonian rocks (of Erian age in and east of central Kentucky) and their relation to major structures in the Eastern Interior basin and adjoining regions. From Collinson and others (1967), Oliver and others (1967), and North (1969).

Sangamon arch (fig. 12). Depressions with relatively thick carbonate rocks include the southern part of the Fairfield basin, the Wittenberg trough, and the northern Missouri-southwestern Iowa area (Collinson and others, 1967, p. 950). The narrow Wittenberg trough was an unusually deep structural trench, now marked by faults, between the Ozark uplift and the Eastern Interior basin to the east (Meents and Swann, 1965, p. 12–16). Although modified and distorted by post-Mississippian movements, this linear feature may represent a continuing zone of crustal weakness and may have been part of the structural and depostional framework during the Mississippian. (See p. 71, 73, 88, 95.)

Acadian tectonism east of the region during late Middle and Late Devonian time heralded the onset of mud and silt deposition from the east. Although Upper Devonian black shales extended throughout most of the region east of the Ozark region, their thickness trends (fig. 13) indicate the continuing positive nature of the previously listed arches except for the Sangamon and perhaps Kankakee arches. The thin Devonian shale sequence in north-central Missouri seems to indicate the presence of a northwest-trending arch in this area.

Depressions during Late Devonian time are shown by westward thickening of black shales into the Moorman syncline and the southern part of the Fairfield basin, and by the mildly depressed Petersburg basin north of the Vandalia arch. In Missouri the Forest City basin and a weak basin in southwestern Missouri, possibly a part of the Arkoma basin, were present in the Late Devonian.

INTERVAL A

INTRODUCTION

Rocks assigned to interval A are generally those of Kinderhook age and are widespread in the Eastern In-

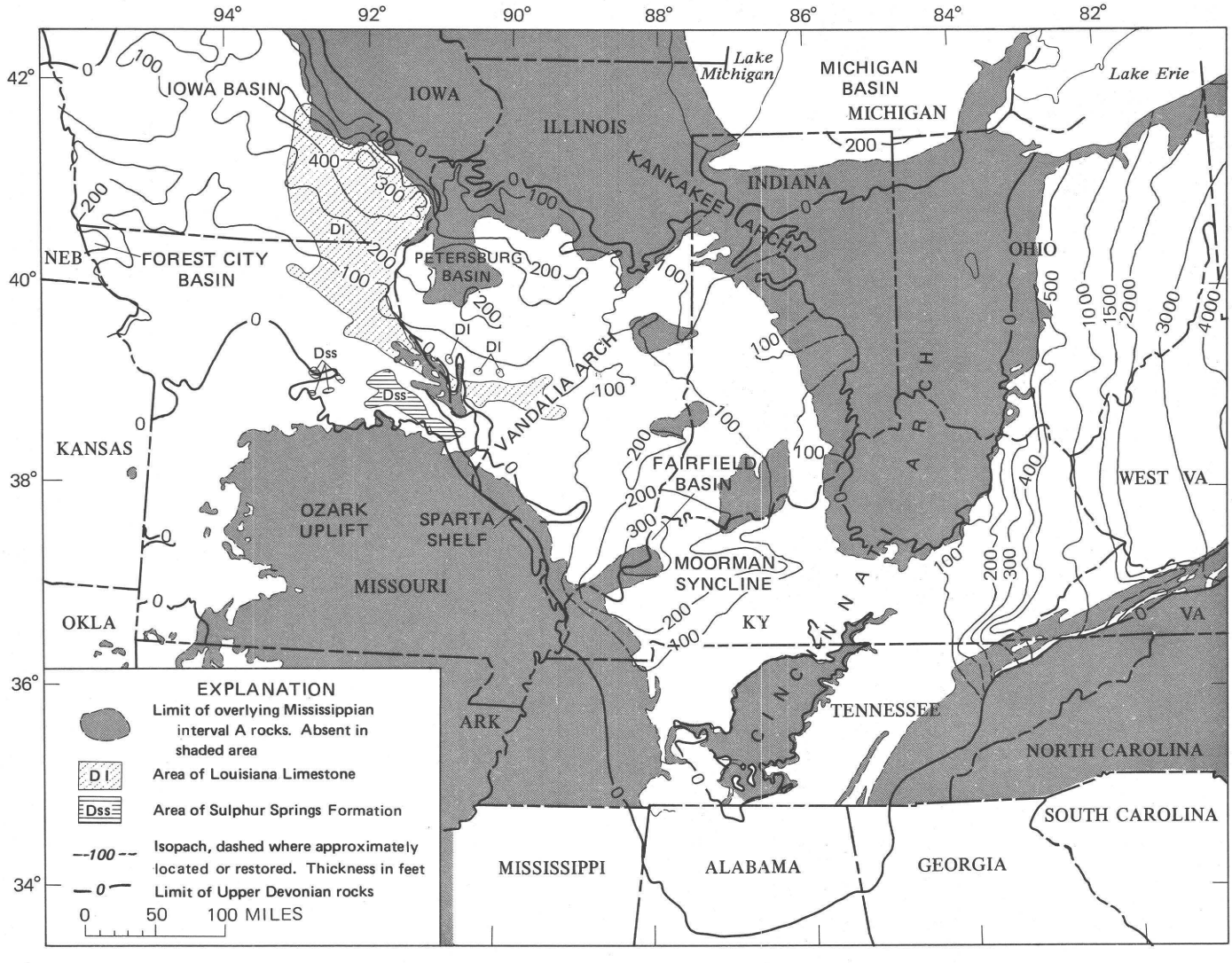


FIGURE 13. — Thickness of Upper Devonian rocks and their relation to major structures in the Eastern Interior basin and adjoining regions. Modified from Collinson and others (1967), Oliver and others (1967), Conant and Swanson (1961), and Harris (1947).

terior basin region. They are continuous with rocks of Kinderhook age that are assigned to interval A in the Appalachian basin and the midcontinent region. In the eastern part of the region, interval A rocks are terrigenous clastics, the volume and areal distribution of which are less than those of Late Devonian age but which were probably derived from the same general source region. Carbonate rocks, deposited in the western part of the region throughout interval A time, intertongued with and onlapped detrital units in the central part of the region during late interval A time.

Results of many early investigations of Lower Mississippian rocks dating from the mid-19th century have been reported. The early writings are reviewed by Cumings (1922, p. 486). Other reviews incorporating later work include: for eastern Kentucky, McFarlan (1943, p. 57–59); for Tennessee, Conant and Swanson (1961, p. 63); for Indiana, Cumings (1922, p. 486), Stockdale (1931, p. 14–15, 71–74); for Illinois, Collinson (1961); and for Missouri, Moore (1928), Branson (1944), and Spreng (1961, p. 53–58). Detailed stratigraphic studies include: eastern Kentucky, Pepper, de Witt, and Demarest (1954, p. 11–13); Tennessee and southern Kentucky, Conant and Swanson (1961, p. 62–69); Illinois, Workman and Gillette (1956) and Buschbach (1952); Missouri, Moore (1928), Branson and Mehl (1938), and Beveridge and Clark (1952). Some of these reviews also contain results of biostratigraphic studies. In more recent years, conodont faunas have been extensively used in determination of age relationships for Kinderhook rocks. Early work in Missouri (Branson and Mehl, 1933a, b, 1938, 1941; E. R. Branson, 1933) and Indiana (Huddle, 1934) has been extended by numerous studies elsewhere. Some of these are: Rexroad and Scott (1964) in Indiana; Hass (1956) in Tennessee and parts of Kentucky and Ohio; Collinson, Scott, and Rexroad (1962) and Scott and Collinson (1961) in Illinois.

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

All or parts of more than 15 formations are assigned to interval A in the Eastern Interior basin, eastern Kentucky, west-central Tennessee, and Missouri (pl. 15). The large number of formational names reflect lithologic distinctions in widely separated areas; a few of the formation names represent provincial designations for identical units. In most of the region, boundaries of interval A coincide with formation boundaries and the interval includes only rocks of Kinderhook age, but in some areas, thin beds of Devonian or Osage age are included at the base and top of the interval where such beds cannot be easily differentiated. At a few places, some beds of Kinderhook age cannot be separated from the underlying Devonian and are excluded from interval A. Interval A rocks are here discussed generally from east to west.

In northeastern Kentucky, rocks of interval A include the Bedford Shale, Berea Sandstone, and Sunbury Shale. Although the basal Bedford has been assigned to the Late Devonian (Hass, 1947) and the Sunbury is Early Mississippian, the age of the upper part of the Bedford and the Berea is uncertain. The Bedford and Berea are discrete units in some areas but intertongue in others (Morris, 1966a; Morris and Pierce, 1967). Together, they form a wedge which thins to extinction to the southwest and which is overlapped by the Sunbury Shale in east-central Kentucky. Thin Sunbury equivalents, such as the uppermost beds of the New Albany Shale, extend southwestward at least to south-central Kentucky.

In Tennessee and Kentucky, south and west of the Bedford and Berea wedge, green mudstone and phosphate nodules of the very thin Maury Formation and its lithic correlatives in the basal Borden Formation and in the Fort Payne Formation are widespread. Conodont faunas in several outcrops in southern Kentucky and Tennessee indicate that the Maury is a time equivalent of most of the Bedford, the Berea, and the Sunbury (Collinson and others, 1962, p. 13) and also that the Maury locally contains elements younger than the Sunbury (Hass, 1956, p. 23). In these areas, the Maury seems to represent continuous deposition during most or all of Kinderhook time (Conant and Swanson, 1961, p. 67). In other areas, such as along the Cincinnati arch in west-central Kentucky, Maury lithic equivalents contain a mixed conodont assemblage of Devonian, Kinderhook, and Osage age and are interpreted as a lag deposit of an erosional or nondepositional hiatus, which possibly extended through Kinderhook and early Osage times (Rexroad and Scott, 1964).

In most of Indiana and parts of western Kentucky, interval A is represented by the thin but widespread Rockford Limestone. The Rockford contains both Kinderhook and Osage conodont elements (Rexroad and Scott, 1964), but because of lack of criteria for division in the subsurface, all the Rockford is here assigned to interval A. Units in the upper few feet of the New Albany Shale underlying the Rockford in the Indiana outcrop belt are also of Kinderhook age (Huddle, 1934; Campbell, 1946), but because these thin units have not been traced in the subsurface, they are excluded from interval A. In northwestern Indiana, however, a thicker unnamed green shale in the uppermost New Albany and underlying the Rockford is considered to be Mississippian in age (Lineback, 1964, p. 56–57); it is equivalent to the Ellsworth Shale of Michigan and is included in interval A.

Rocks assigned to interval A in Illinois are continuous with those of Indiana and Kentucky are given different formational names, and in the Mississippi Valley area they also include several formations that are not represented in the region farther east. In general, interval A strata in Illinois consist of a lower unit of mudstone, the Hannibal Shale, which is continuous with the uppermost beds of the New Albany Shale of Indiana, and an upper carbonate-dominated unit, the Chouteau Limestone, which is continuous with the Rockford Limestone. The Hannibal thickens and contains increasing amounts of siltstone and sandstone to the northwest. The upper part of the Hannibal interfingers with the lower part of the Chouteau (Collinson, 1961, p. 107). This gross twofold lithologic division is recognizable to the west in north-central Missouri, although more than two formations are distinguished within the sequence in western Illinois and Missouri and Iowa. The Hannibal Shale as now used by the Illinois State Geological Survey is about the same as the *Hannibal Group* of Workman and Gillette (1956) in Illinois but excludes parts or all of the *Maple Mill Formation* of Workman and Gillette in some areas and restricts their Glen Park Formation to a thinner discontinuous basal Mississippian unit (Collinson, 1961, p. 104–106). In western Illinois the "*Glen Park*" *Formation*, locally underlying the Hannibal, contains conodonts which probably represent the oldest Mississippian assemblage in the Mississippi Valley (Collinson, 1961, p. 105). The "*Glen Park*" is included in interval A, as are beds called the *English River Siltstone* in Illinois well records. The *English River Siltstone* in Iowa, however, is in part of Late Devonian age (Collinson, 1961, p. 106).

In extreme western Illinois, adjoining southeastern Iowa and northeastern Missouri, interval A consists of the *North Hill Group* (Workman and Gillette, 1956) and includes, in ascending order: the *McCraney Limestone* (dominantly carbonate), *Prospect Hill Siltstone* (mudstone, siltstone, and sandstone), and *Starrs Cave Formation* (dominantly carbonate). Because rocks of the *North Hill Group* are not physically continuous with the Hannibal and Chouteau strata to the southeast, correlation of the *McCraney* with the Chouteau or with the Louisiana Limestone has been debated for many years. Correlation of the *McCraney* with part of the Chouteau (Scott and Collinson, 1961, p. 119–120) is accepted in this report. The underlying Hannibal (or *English River*) and the overlying *Prospect Hill* contain siltstones and sandstones which generally coarsen and become more abundant to the northwest.

The undifferentiated Chouteau Limestone (Spreng, 1961, p. 53) makes up most of interval A throughout northern Missouri, and, locally, it includes beds as young as the Gilmore City Limestone of Iowa. It is underlain locally by the Hannibal which thins to extinction towards the Ozark uplift (pl. 9-A, sec. *e*–*e'*). In northwestern Missouri the upper part of the *Kinderhook shale*, a westward-thickening wedge equivalent to the *Boice Shale* of Nebraska and Kansas (Carlson, 1963, p. 12), is also present and is assigned to interval A. In west-central and southwestern Missouri, *Chouteau* is a *group*, subdivided in ascending order into the Compton Limestone, Sedalia Limestone, and Northview Shale (Beveridge and Clark, 1952, p. 71). The Northview and Sedalia are interfingering facies (Beveridge and Clark, 1952, p. 74). These formations cannot be delineated everywhere, however, particularly where the Northview is thin or absent and the Compton is lithologically like the Sedalia. The Compton was considered by Spreng (1961, p. 53) to be equivalent to the undifferentiated Chouteau of northeastern Missouri, but the entire *Chouteau Group* of Missouri was equated to the Chouteau of Illinois by Collinson (1961, p. 106).

UPPER BOUNDARY OF INTERVAL A

Rocks of interval A are overlain by Pennsylvanian strata across north-central Illinois and western Indiana and in scattered areas marginal to the northern and western Ozark region. Cretaceous sedimentary rocks overlie interval A on the margins of the Mississippi embayment. Elsewhere, rocks of interval A are almost everywhere overlain by rocks of interval B. Intervals A and B generally appear to be conformable in most places, although evidence of erosion or nondeposition and angular relationship is present in several areas.

Fine-grained detrital rocks in the Borden Formation in interval B overlie interval A in northeastern, south-central, and west-central Kentucky, and in southwestern Indiana and eastern to southwestern Illinois. (See pl. 9-B.) Judging from the widespread persistence of thin units in interval A — including the Sunbury, Maury, Rockford, and Chouteau — little or no erosion preceded deposition of the Borden. A hiatus in the eastern part of the region along the Cincinnati arch is indicated, however, by conodont data from Indiana (Rexroad and Scott, 1964) and parts of Kentucky (J. W. Huddle, written commun., 1965), which show that basal Borden beds in northern areas are older than those to the south and probably indicate continuous deposition in the north. The Rockford Limestone is locally absent in the southern Indiana outcrop belt (Lineback, 1964), and small channels filled by interval B rocks, which cut interval A rocks and rest on older strata in Kentucky (R. C. Kepferle, oral commun., 1967), indicate at least minor scour prior to deposition of interval B sediments in those areas.

In most of the region south of the limit of rocks assigned to the Borden — in southern and western Kentucky and western Tennessee — green mudstones or siliceous carbonate rocks of the Fort Payne lie at the base of interval B and rest with apparent conformity on the Maury Formation. Abrupt contacts between Fort Payne siliceous rocks and the Maury were considered by Conant and Swanson (1961, p. 68) to be due to sudden changes in depositional environments. Conodont faunas collected at scattered localities in southern Kentucky indicate that basal beds of the Fort Payne may vary in age from place to place (J. W. Huddle, written commun., 1965; Hass, 1956), and an obscure hiatus may therefore be hidden in beds within or at the top of the Maury.

In parts of southeastern Illinois, the *Ullin Limestone* directly overlies the Chouteau Limestone, and because the *Ullin* is younger than the Borden and Fort Payne (Lineback, 1966, p. 29), a nondepositional hiatus between the Chouteau and the *Ullin* in the deepest part of the basin is suggested (Lineback, 1969). Elsewhere in southern Illinois, mudstones assigned to interval B (*Springville Shale*) conformably overlie the Chouteau, without any interruption in the conodont assemblages (Collinson and Scott, 1958).

In contrast to the generally conformable contact between interval A and interval B in the eastern part of the region, the upper boundary of interval A is marked by erosional unconformities in areas marginal to the Ozark uplift. The *Meppen*, Fern Glen, and Burlington Limestones of interval B are unconformable on interval A and older rocks in southwestern Illinois and southeastern Missouri. In western Illinois, the Burlington and *Meppen* appear to truncate interval A units (pl. 9-A, secs. a–a', c–c', d–d') and to rest on Devonian Saverton and Grassy Creek Shales. There, erosion following uplift early in Osage time (Workman and Gillette, 1956, p. 43) is interpreted to have removed part, and locally all, of interval A rocks; the eastern and western limits of the area eroded roughly coincide with the western edge of the Chouteau and eastern edge of the *McCraney*, respectively. Pre-interval B erosion also probably removed some interval A rocks along the Lincoln anticline in northeastern Missouri (pl. 9-A, secs. c–c', d–d').

In much of western Missouri, the *Pierson Limestone* overlies interval A strata abruptly, and a hiatus is indicated at the contact by conodont studies (Thompson and Fellows, 1969). Difficulty in separating the Sedalia Limestone and the *Pierson Limestone* in west-central Missouri may locally have caused inconsistencies in the placement of the interval A–interval B boundary. The two formations are lithologically similar in this area, and the intervening Northview Shale is thin or absent (Spreng, 1961, p. 60). At places where the *Pierson* may inadvertently be included with the Sedalia, the thickness of interval A might be increased by as much as 25 percent (pl. 3-A).

THICKNESS AND LITHOFACIES TRENDS

The thicknesses of rocks in interval A (pl. 3-A), which generally are less than 100 feet, are believed to be the original total rock thicknesses in areas generally east and south of central Illinois and in western and northwestern Missouri. An area of reduced thickness or local absence of interval A beneath interval B in western Illinois and and the local absence of interval A in southwestern Illinois result from erosion before or during interval B. Absence of interval A beneath interval B in three areas in eastern Illinois and southern Indiana is interpreted as due to nondeposition, rather than erosion. In most of the region, rocks of interval A probably extended far beyond their present edges, which have been eroded back either early in interval B or following deposition of the Mississippian. In areas marginal to the Ozark uplift in southwestern Illinois, southeastern and extreme southwestern Missouri, and along the La Salle anticlinal belt, the present zero limits may correspond to the depositional limit of interval A sediments.

Interval A in the Eastern Interior basin region is thin compared to the succeeding Mississippian intervals. It approaches or exceeds 100 feet in four areas: in northeastern Kentucky (Bedford, Berea, and Sunbury clastic wedge), in an arcuate belt from western Illinois through northeastern Missouri (Hannibal and "*Glen Park*" strata), in northwestern to western Missouri (Chouteau and the "*Kinderhook shale*"), and in a narrow west-northwest-trending belt in southwestern Missouri (Northview Shale).

A remarkable feature of interval A rocks is the widespread persistence of the very thin Chouteau and Rockford Limestones and the Maury Formation, which rarely exceed 10 feet in thickness throughout the large area of southeastern Illinois, Indiana, much of Kentucky, and Tennessee.

Lithofacies distribution of interval A (pl. 3-B) shows four areas in which detrital rocks are predominant. These areas are flanked by broad belts in which the sediments are mainly carbonates. The coarsest detrital rocks do not exceed sand size, and commonly the largest grains are silt size; carbonate rocks are predominantly aphanitic to fine grained, silty, and in part cherty. A discussion of areas of detrital accumulation follows, succeeded by a discussion of areas of dominant carbonate sediments.

In Tennessee and most of Kentucky, thin mudstones and siltstones of the Maury Formation and its lithic

equivalents occupy a broad northeast-trending belt. Sandstone in the Bedford and Berea sequence in northeastern Kentucky produces southwestward-convex lobate facies patterns which generally coincide with the southwestward thinning of interval A there. Chemical components in the Maury represented by phosphatic nodules (Conant and Swanson, 1961, p. 63–66), although common, are not sufficiently abundant to affect the lithofacies pattern.

Mudstone and siltstone in the Hannibal Shale and equivalents are distributed in a belt from northwestern Indiana across western Illinois and northern Missouri. Southeastwardly convex lobes extend into central Illinois. In western Illinois and northeastern Missouri, a very high mudstone content in interval A is partly the result of thickening of the Hannibal, but northward it is largely the result of the removal of Chouteau and younger carbonate rocks of the interval. Southwestward thinning of the Hannibal occurs fairly abruptly in northeastern and east-central Missouri, along a lithofacies boundary that coincides roughly with southern limits of the Louisiana Limestone (fig. 13). This boundary may have been a hinge line during Late Devonian and Early Mississippian time, which roughly coincided with the Cap au Gres fault and the axis of the Lincoln anticline (fig. 11). These structural features apparently were intermittently active in middle Paleozoic time (Rubey, 1952). Higher coarse- to fine-grained clastic ratios along the Mississippi River near southeastern Iowa result from the presence of sandy beds in the more nearly complete sections of the *North Hill Group*, *McCraney*, and Hannibal.

In northwestern Missouri, greenish-gray and gray mudstones of the *Boice* equivalent are combined with overlying thick Chouteau carbonate rocks to produce an intermediate facies.

Siltstone and green mudstone in the Northview Shale are conspicuous components of interval A in a west-trending belt in southwestern Missouri. The Northview is thickest in a narrow west-northwest-trending trough along the northeast side of the belt (Beveridge and Clark, 1952, p. 79) where siltstone–shale ratios are generally highest. A carbonate facies characterizes rocks of interval A north and south of this belt where the Northview is thin or absent. Conglomeratic rocks are locally present northeast of the main mass of Northview in the Weaubleau Creek area, west-central Missouri (Beveridge, 1951; Snyder and others, 1965).

Two ill-defined belts of carbonate rocks include a southwest-trending belt from central Indiana to southern Illinois where interval A consists largely of the Rockford and Chouteau Limestones, and a belt encompassing most of northern, central, and western Missouri where the interval consists mostly of the *Chouteau Group*. In isolated areas in southeastern and eastern Illinois, underlying mudstones of the Hannibal are sufficiently thick to change the lithofacies pattern. In this region and in adjacent parts of Illinois and Indiana, the Rockford and Chouteau are so thin that inclusion of only the top few feet of the underlying New Albany Shale where it is Mississippian in age would alter the lithofacies pattern.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The Mississippi Valley area of western Illinois and northeastern Missouri was largely emergent at the beginning of interval A time (Williams, 1957, p. 301–305). During interval A time most of the Eastern Interior basin region and extensive surrounding areas were sites of a shallow epeiric sea that opened to the south and west (pl. 11, fig. 1; pl. 12, fig. 1). Two major and one or more minor source areas supplied sediments. Streams flowed westward and southward into the eastern part of the region from highlands east of the northern and central Appalachian Mountains and from northeastern Canada. Hilly land in the Lake Superior region from which detrital sediments were shed southward was a southwestward extension of the Canadian Shield probably related to the Transcontinental arch. Low hills in the Ozark region contributed detrital sediments, most of which were dispersed westward. Low coastal plains probably adjoined the seas; a marginal swampy delta plain in Wisconsin probably represented waning deltaic deposition. Parts of the Cincinnati arch were probably emergent, but had very low relief. Small islands in central Tennessee may have been local sources for sand in the Maury Formation in west-central Tennessee. Abundant mudstone and absence of detrital grain sizes coarser than fine sand indicate either that sediment sources were distant from depositional sites or that source areas were of low relief, or both.

In northeastern Kentucky, lithofacies and thickness distribution of sandstone-dominated lobes suggest westerly and southwesterly transport directions for the Bedford and Berea terrigenous detritus from source areas east and northeast of the Appalachian Mountains (Pepper and others, 1954). The sandstones, generally orthoquartzites, were derived probably from earlier Paleozoic detrital rocks (Pepper and others, 1954, p. 91, 95). Symmetrical ripple marks in this area (Morris, 1965a, b, 1966a, b; Morris and Pierce, 1967) and in a large area of southern Ohio trend northwest, and have been ascribed to northeasterly prevailing winds (Bucher, 1919; Pepper and others, 1954, p. 91) or shoreline coastal control (Hyde, 1911). Crossbedding

and sandstone-filled channels characterize the sequence and are features compatible with deposition in a shallow-water deltaic complex. Sole markings in some sandstones suggest that they were deposited by density currents, probably at the delta front (Wilson, 1950; Rich and Wilson, 1950). The relative importance of sediment contributed by the southwest-flowing dispersal system, the Ontario River, to that contributed by the westward-flowing Gay-Fink, Cabin Creek, and Virginia-Carolina Rivers (chap. C, fig. 5; Pepper and others, 1954) is not known, although lithofacies trends suggest that the west-flowing rivers contributed most heavily in northeastern Kentucky.

Except for southwestward thinning, the Sunbury Shale in eastern Kentucky exhibits no definitive clues to the transport direction of the sediment. The Sunbury accumulated in a marine-reducing environment during transgression of the sea, which produced the other Devonian-Mississippian black shales.

The widespread thin mudstone deposit represented by the Maury Formation probably was deposited remote from source areas (Conant and Swanson, 1961, p. 68) with slow deposition in a low-energy environment favorable to the precipitation of phosphate. Alternatively, phosphate in the Maury represents a lag concentrate from which most of the fine detrital material was winnowed by submarine currents; this hypothesis is suggested by the mixed conodont assemblages. Sandstones, which appear in the Maury along the western margin of the Nashville dome, suggest local source areas (Conant and Swanson, 1961, p. 53). The normally underlying Chattanooga Shale is absent in this area indicating local uplift.

A lowland area along the Cincinnati arch effectively confined most of the sediment from Appalachia sources to the Appalachian basin, but itself was not the source for much detritus.

Sediments in the Hannibal Shale in northwest Indiana, western and central Illinois, and northeast Missouri may have been derived from a source in east-central or northern Canada, transported by a southwest-flowing stream system into a marine trough extending from western Michigan into western Illinois. It seems more probable, however, that the source of sediment was a land area along the northeastern part of the Transcontinental arch or its extension, the Wisconsin arch. Evidence for the latter interpretation includes northwestward or northward coarsening of clastics in the Hannibal and the presence of southeasterly convex lobate areas of dominantly detrital rocks adjacent to the Wisconsin arch (pl. 3-B). The Ellsworth Shale, which composes interval A in the Michigan basin, is thickest and generally coarsest grained on the west side of the Michigan basin (Cohee and others, 1951), which fact also suggests a source along the Wisconsin arch in that region.

Oolitic limestone in the discontinuous "*Glen Park*" indicates at least local agitated shallow-water environments early in interval A in the Mississippi Valley area of western Illinois and northeastern Missouri. Widespread subsidence took place in the region, except in lowland in the Ozark region, during Hannibal time. Workman and Gillette (1956, p. 43) indicated that deposition of the relatively thick sequence of Hannibal and "*Glen Park*" sediments across western and west-central Illinois may have begun in a structurally restricted embayment opening to the west, followed by regional subsidence and an extensive marine transgression. This interpretation seems to be based on the assumption that all preserved sections of Devonian-Mississippian shale in western Illinois contain Kinderhook detrital components. If the interpretation that pre-interval B erosion has removed large amounts of interval A rocks in northwestern Illinois and Iowa (p. 70) is valid, Hannibal sediments once deposited there may have been part of a widespread delta which was thicker and which contained coarser sediments in the area where erosion has taken place than in areas where the Hannibal is still preserved. In western Illinois and northeastern Missouri, volume of sediment was considerably less from northern sources than from the eastern sources in Appalachia.

The possible source area for *Boice* equivalents in northwestern Missouri probably was land of low relief along the Transcontinental arch in Minnesota or a southward extension of the arch in central Iowa. No facies trends are evident in this unit in Missouri, but the presence of hematitic oolites indicates high energy and shallow seas, and the red shales may have been derived from residual soils.

Ozark region sources for detrital rocks of the Northview Shale in southwest Missouri are strongly indicated by the lithofacies relations (Carlson, 1963, p. 41). Conglomeratic rocks and abnormal thinning of Northview beds near the Weaubleau Creek disturbance, western Missouri, were interpreted by Snyder and others (1965) as evidence of nearby Early Mississippian structural movements. The bulk of detrital sediments in the Northview, however, probably was derived from more extensive uplifts in the Ozark region.

Carbonate rocks in the Rockford and Chouteau sequence in Indiana and Illinois were mostly deposited in a low-energy environment, on a sea floor having little relief. A high iron content and a low proportion of terrigenous detritus in Indiana may indicate a very low land area to the east along the Cincinnati arch. A tongue of reddish limestone in the Chouteau extends from southwest to central Illinois (Buschbach, 1952). Ferric iron, which gives the limestone its color, may in-

dicate a local oxidizing depositional environment. Ozark region sources for detrital components in the Chouteau were suggested by Buschbach (1952).

In Missouri, fine-grained carbonate rocks in the Compton and Sedalia Limestones indicate a generally low-energy shelf environment. Dolomite or dolomitic limestone in the Sedalia may indicate restricted hypersaline or supratidal conditions. All the carbonates in interval A probably accumulated in shallow neritic environments, although the Rockford and Chouteau Limestones of southwestern Indiana and southern Illinois may have been deposited at greater depths than were those same limestones to the north and west.

PALEOTECTONIC IMPLICATIONS

Major structural elements during interval A time were of very low amplitude. There are no known intersystemic or intrainterval unconformities of much magnitude, and rock thicknesses indicate fairly even deposition over broad areas. The region was generally negative except for the Ozark uplift, including the Sparta shelf, Cincinnati arch, Wisconsin arch, parts of the La Salle anticlinal belt, and parts of the Transcontinental arch to the north and northwest. These slightly positive, northeast- and northwest-trending structural elements were present in Late Devonian time (fig. 13). The negative elements that had subsided considerably during the Devonian — the Fairfield basin, Moorman syncline, and Wittenberg trough (figs. 12, 13) — cannot be recognized as distinct structural features during interval A. It seems likely that a low-relief topography inherited from the Devonian, rather than interval A tectonism, accounts for interval A thickness variations in most of the region. Tectonic subsidence may have occurred in the area of the Lincoln anticline and Cap au Gres fault, structures which were active in the middle and late Paleozoic (Rubey, 1952; Cole, 1961); interval A strata are somewhat thicker in a northeast-trending belt there.

The widespread thin deposits of the Rockford and Chouteau Limestones in Illinois and Indiana accumulated on a broad stable shelf. Maury Formation mudstones accumulated on the southern extension of this shelf. Although the paleotectonic map (pl. 10, fig. 1) shows a stable positive area south of the shelf, an alternate interpretation would be that a negative element of large magnitude adjoined the shelf there. This interpretation is based on the assumption that paleogeographic models proposed to explain the origin of phosphate in the Phosphoria Formation of the Western United States (McKelvey and others, 1953) can be applied to phosphate in the Maury Formation. The northward progression from these phosphate-bearing rocks to carbonates to increasingly coarse terrigenous rocks might indicate landward transition from a subsiding, oceanic basin.

After interval A time, positive areas, such as the Transcontinental and Wisconsin arches, the Ozark uplift, and the Cincinnati arch are interpreted to have risen slightly. Emergent areas expanded and included parts of western Illinois and much of the region to the northwest.

INTERVAL B

INTRODUCTION

Pertinent literature pertaining to rocks assigned here to interval B includes: for Kentucky, Butts (1917, 1922), Stockdale (1939), Walker (1962), and Weir, Gualtieri, and Schlanger (1966); for Indiana, Stockdale (1931) and Smith (1965); for Tennessee, Bassler (1932) and Marcher (1962); for Illinois, Swann, Lineback, and Frund (1965) and Lineback (1966, 1968b); and for Missouri, Moore (1928), Laudon (1937), Cline (1934), Kaiser (1950), Spreng (1961, p. 59–66), Kissling (1961) and Thompson (1967).

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Interval B rocks, which are generally the rocks of Osage age, have a present distribution that closely corresponds to that of interval A rocks. However, interval B rocks, in comparison to interval A rocks, are generally thicker and lithologically more diverse, and they exhibit more complex interrelationships. In general, the bulk of interval B rocks comprise three marine rock assemblages, which are not strict equivalents but which are dominant in specific areas. These assemblages, which occur largely in the lower and middle parts of interval B, are as follows:

1. Crinoidal, in part argillaceous, carbonate assemblage (*Meppen*, Fern Glen, Burlington, and Keokuk Limestones); extends from western and southwestern Illinois across much of Missouri and in adjoining States of Iowa, Nebraska, and Kansas.
2. Siliceous and argillaceous carbonate assemblages which include part of the Fort Payne Formation, the Reed Spring and *Elsey* Formations, and the Grand Falls Chert Member of the Boone Formation; extend across southern Kentucky and Tennessee into southeastern Illinois and southwestern Indiana, and southwestern Missouri.
3. Detrital clastic assemblage (*Borden Siltstone*, *Springville* and Warsaw Shales, and parts of Fort Payne Formation in Illinois); deltaic wedges which are mostly confined to the northeastern parts of the region but whose distal equivalents extend throughout much of the area east of the Ozark region.

The oldest interval B units in the Eastern Interior basin region are the *Meppen*, Fern Glen, and *Pierson* Limestones. In western Illinois, the dolomitic limestones of the *Meppen* are the oldest interval B strata in this part of the region. They unconformably overlie interval A strata, are overlain by the Fern Glen, and are overlapped northward by the Burlington Limestone. The *Meppen* is considered to correlate with the *Pierson Limestone* of Missouri on lithologic and faunal evidence. Reddish- and greenish-gray limestones and mudstones of the Fern Glen extend in a narrow northeast-trending belt from southeastern Missouri across central Illinois. Upper beds of the Fern Glen are transitional with the Burlington Limestone (Kissling, 1961). Carbonate rocks of the *Pierson Limestone*, which is in the upper part of the *St. Joe Group* in Missouri, are best developed from southwestern to central Missouri. The *Pierson* underlies the Reeds Spring Limestone in southwestern Missouri and the Burlington in west-central Missouri. In parts of southwestern Missouri, variegated rocks in the *Pierson* resemble those in the Fern Glen. Basal *Pierson* beds are of Kinderhook age in some localities (Spreng, 1952, p. 81–86; Thompson, 1967, p. 23); beds in the lower and middle parts of the *Pierson* include the oldest Osage rocks in Missouri, the upper beds of which are correlated with the Fern Glen (Spreng, 1952). Northward onlap of the *Pierson* in southwest Missouri is indicated by Kinderhook conodonts in its basal beds in Arkansas and by Osage conodonts at its base farther north (Thompson and Fellows, 1969, p. 73).

The Reeds Spring, *Elsey* (Robertson, 1967), and Grand Falls Chert Member of the Boone Formation, which consist of more than 300 feet of cherty limestone, argillaceous limestone, and locally abundant chert, are areally restricted to southwestern Missouri and adjacent States. They conformably overlie and intertongue with the *Pierson* and underlie the Burlington or Keokuk. Difficulty in differentiating units within this sequence of cherty rock and in separating them from the overlying Burlington or Keokuk has resulted in the assignment of the equivalents of all these units to the Boone Formation, or to the Reeds Spring, *Elsey*, and Grand Falls undifferentiated. The Reeds Spring is an age equivalent of the upper part of the Fern Glen (Moore, 1935, p. 241; Kissling, 1961, p. 147). Age of the *Elsey* and Grand Falls relative to units in the Mississippi Valley sections is uncertain. Chert and the presence of small carbonate reefs (Spreng, 1961, p. 60; Troell, 1962) in the Reeds Spring and Grand Falls suggest that these units are similar to rocks of the Fort Payne Formation of western Kentucky and Tennessee. The Fort Payne is considered to be younger than the Burlington Limestone (Lineback, 1966).

The Burlington and overlying Keokuk Limestones are the most extensive and volumetrically important interval B units in the western part of the region. They occur in nearly all areas where the interval is present in western Illinois, Missouri, Iowa, Kansas, and Nebraska, and outliers of the Burlington and Keokuk are present in structurally high positions on the Ozark uplift and the Kankakee arch. The contact between the two formations is generally obscure away from their type localities in Iowa, and they are generally combined as a single map unit, the Burlington and Keokuk undifferentiated. The lower or Burlington part of the map unit, which is characterized by light-colored cherty crinoidal limestones, is present in nearly all areas in the western part of the region except for parts of southwest Missouri (pl. 15, col. 56); the upper or Keokuk part of the map unit is widespread, is generally darker, and includes thin beds of shale in northeastern Missouri and western Illinois. Thickness of both formations ranges from 50 to more than 200 feet. The basal contact of the Burlington is an unconformity in several areas. In western Illinois the Burlington rests unconformably on rocks as old as the Devonian Saverton Shale, and in Missouri on pre-Devonian strata in scattered areas within and marginal to the Ozark uplift. Elsewhere, it rests with apparent conformity on the *Meppen*, Fern Glen, *Pierson*, or the Reeds Spring, *Elsey*, and Grand Falls undivided. From crinoid studies, Laudon (1937) concluded that basal Burlington beds in eastern Missouri and Iowa became progressively younger northward and represent an extensive marine transgression. Thompson (1967) interpreted the Burlington in southwestern Missouri and underlying interval B units as progressively younger southward and westward.

In the eastern part of the Eastern Interior basin region, interval B units are largely made up of detrital rocks in the lower part overlain by carbonate-rich beds in the upper part. The Borden Formation (Weir and others, 1966), a southern continuation of the Cuyahoga Group and Logan Formation of Ohio (Hyde, 1915), makes up all or most of interval B in northeastern, east-central, south-central, and west-central Kentucky, and extends northwestward into Indiana. There, the Borden is classified as a group; its boundaries are the same as the Borden Formation boundaries in adjacent Kentucky. The Borden in these States consists of a lower unit of gray and green mudstone and siltstone, which contains minor beds of sandstone, and an upper unit of cherty or silty dolomitic limestone. Several local formation and member names are given to widespread green mudstone, minor sandstone tongues, and lenses of crinoidal limestone in the lower part of the Borden. These include the Nancy Member, New Providence Member, an unnamed shale member in Kentucky, and

the New Providence Formation in Indiana. Tongues and lenses of siltstone in the middle part of the Borden in Kentucky grade northward into more abundant coarser siltstone and sandstone units in Indiana. These include the Cowbell, Halls Gap, and Nancy in Kentucky and the Locust Point, Carwood, and Edwardsville Formations in Indiana. The uppermost division of the Borden includes silty and sandy dolomite in eastern Kentucky, the Renfro Member (Schlanger, 1964), and a thicker unit of light-colored cherty and silty limestone and dolomite (the Muldraugh Formation or Member) in south-central and west-central Kentucky and in Indiana (Weir and others, 1966; Smith, 1965). The carbonate rocks are commonly separated from underlying clastic strata by thin glauconitic siltstone beds in eastern to west-central Kentucky and by thin carbonate and glauconitic siltstone of the Floyds Knob Formation (or Floyds Knob Bed of the Muldraugh Member) in west-central Kentucky and parts of Indiana; they are overlain by the Harrodsburg Limestone. Westward thickening of the carbonate rocks and reciprocal thinning of underlying detrital units are characteristic internal relationships in the Borden. Contacts between these rocks are generally southwestward-sloping planar surfaces (pl. 9-B, sec. $c-c'$).

Megafaunas in the basal part of the Borden of northeastern and east-central Kentucky are assigned to Fern Glen and possibly Burlington ages (Butts, 1922, p. 50). In west-central Kentucky, however, fauna in the basal part of the Borden (New Providence Shale of Butts) is Burlington in age (Butts, 1917, p. 17; Conkin, 1957; Collinson and Scott, 1958; Rexroad and Scott, 1964). Thus, a hiatus representing Fern Glen and part of Burlington time may be indicated at the base of the Borden in some Kentucky areas.

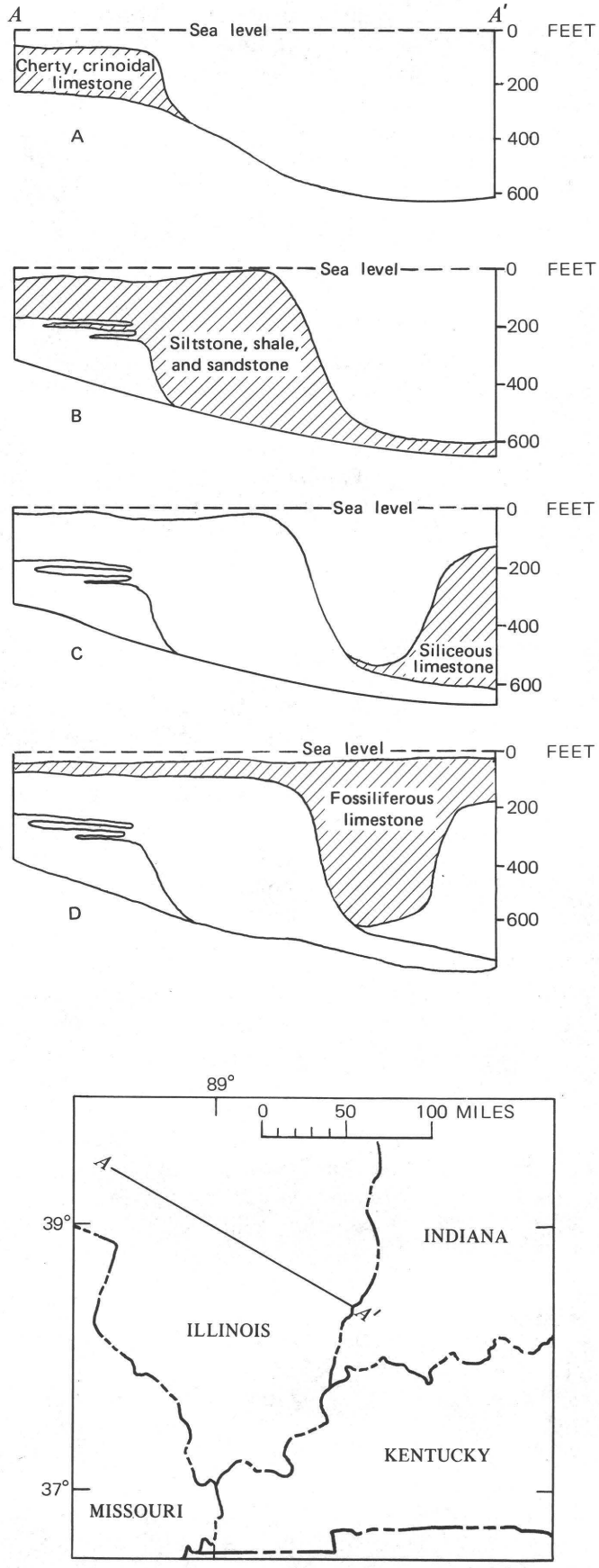
In Illinois, the name *Borden Siltstone* is restricted to detrital rocks that are continuous with the lower part of the Borden Group of Indiana and that extend southwestward across Illinois as an elongate deltaic tongue. The formation is mostly mudstone and siltstone to the south, but it contains increasing amounts of sandstone to the north. In southern Illinois, thin distal mudstone equivalents of the Borden at the base of interval B were named *Springville Shale* (Collinson and Scott, 1958; Lineback, 1966). The *Springville* is probably continuous with similar mudstones, the New Providence Shale, in the basal part of interval B in western Kentucky and Tennessee and is of early Osage age, possibly a Fern Glen equivalent (Collinson and Scott, 1958). The Warsaw Shale and associated sandstones, such as the *Sonora Formation*, are, at least in part, equivalents of the *Borden Siltstone* (Swann and others, 1965), but rocks termed Warsaw are assigned to interval B or interval C in different parts of the region as described in the following discussion.

Unnamed cherty carbonate rocks in the lower part of the Borden in northwestern Indiana and eastern Illinois (pl. 9-B, sec. $d-d'$) appear to be an easterly extension of the Burlington and Keokuk undifferentiated. Detrital rocks in the lower part of the Borden in that area also include variegated shales similar to those in the Fern Glen and to those in the Coldwater Shale of Michigan. Correlation of these rocks with the Burlington and Keokuk and the Fern Glen extends the Burlington and Keokuk carbonate sequence as far north as the western margin of the Michigan basin.

Strata in interval B younger than basal Burlington were continuously deposited in western Illinois and most of Missouri. The Keokuk overlies the Burlington with apparent gradation, and the Warsaw Shale and *Sonora Formation* (pl. 9-B, secs. $a-a'$, $c-c'$, $d-d'$) appear to represent a culmination in the deposition of terrigenous detritus whose weak beginning is shown in shaly beds of the Keokuk. Similarly, Warsaw carbonate rocks (interval C) in Missouri overlie the Burlington and Keokuk carbonate unit with apparent conformity along an obscure boundary. In southwest Missouri, the Short Creek Oolite Member in the upper part of the Keokuk is a key bed in a succession of otherwise uniform limestone. The top of the Short Creek is arbitrarily selected as the interval B–interval C boundary in that area (pl. 15, col. 56).

The *Ullin Limestone*, a light-hued fossiliferous carbonate rock that overlies the *Borden Siltstone* and Fort Payne Formation in some areas of central, eastern, and southwestern Illinois, has been subdivided there into a lower cherty carbonate unit, the *Ramp Creek Limestone Member*, and an upper crinoidal limestone containing minor chert, the *Harrodsburg Limestone Member* (Lineback, 1966). The *Ramp Creek* is correlated with the Muldraugh Formation or Member of Indiana and Kentucky, which is Osage in age (Stockdale, 1939, p. 201). The Harrodsburg Limestone, as currently used in Indiana and Kentucky, seems to be continuous with and equivalent to the *Harrodsburg Member* of the *Ullin Limestone* in Illinois.

Dark cherty siliceous silty to argillaceous carbonate rocks of the Fort Payne Formation make up most or all of interval B in a broad area encompassing southern and western Kentucky, west-central Tennessee, and parts of southern and southeastern Illinois, and lithic equivalents occur in the Borden Group in the subsurface of southwestern Indiana. In Illinois and Indiana, Fort Payne rocks thin against the margins of the *Borden Siltstone* (Lineback, 1966). The younger *Ullin Limestone* similarly thins, but it overlaps both the Fort Payne and the Borden (pl. 9-B, secs. $a-a'$, $e-e'$), indicating that local nondepositional hiatuses preceded *Ullin* deposition in some areas. Thus, the Borden is interpreted to be oldest, and the Fort Payne and *Ullin* to



be successively younger units in Illinois and Indiana (Lineback, 1966). In Kentucky, relationships of the Borden Formation to the Fort Payne indicate that both the Muldraugh Member of the Borden Formation and the Fort Payne are younger than Borden detrital rocks. The Muldraugh seems to be contemporaneous with at least the lower part of the Fort Payne, but beds in the upper part of the Fort Payne in south-central Kentucky may be younger than the Muldraugh.

Modern views on regional stratigraphic relationships of interval B rocks differ considerably from older concepts. Earlier workers broadly equated such units as the Burlington and Keokuk, the Fort Payne, and the Borden, which were considered to be facies of one another. Rock-stratigraphic units within some formations, such as the Borden, were also interpreted to be facies (Stockdale, 1931, 1939). The field and subsurface relations, however, cannot be reconciled at many places with this model. In central Illinois, Borden detrital rocks give way abruptly, rather than gradationally, to Burlington and Keokuk carbonate rocks (pl. 9-B, secs. $a-a'$, $c-c'$). East of areas where Borden detrital rocks are present, carbonate rock units in the *Ullin* overlie the Fort Payne Formation or *Springville Shale*, but elsewhere they appear to be Borden and Fort Payne age equivalents. The same is true in southern Indiana and central Kentucky, where the Muldraugh Formation at places overlies siltstones and mudstones of the Borden but at other places appears to be equivalent to the Borden detrital rocks (Stockdale, 1939, p. 202–222).

More recent interpretations of the stratigraphy indicate that the major post-Fern Glen interval B units are not contemporaneous facies. First implied by Butts (1917) in eastern Kentucky, extended in eastern regions by Walker (1962), and amplified by detailed subsurface studies in Illinois (Frund, 1953; Swann and others, 1965; Lineback, 1966, 1968b), the concept was also developed independently during surface work in central Kentucky (Weir and others, 1966; Kepferle, 1966, 1967a; Peterson, 1966). According to Lineback (1966), the Fern Glen, Burlington and Keokuk, Borden, Warsaw, Fort Payne, and *Ullin* Formations are successively younger deposits whose distribution was controlled by a depositional submarine topography (fig. 14; compare with pl. 9-B, sec. $g-g'$). The *Meppen*, Fern Glen, and most of the Burlington and Keokuk antedate the *Borden Siltstone*. The Warsaw Shale of western Il-

FIGURE 14.—Generalized reconstructed cross sections showing relations of successively deposited strata in south-central Illinois during interval B and earliest interval C times. *A*, deposits near end of deposition of Keokuk Limestone; *B*, at end of *Borden Siltstone* and Warsaw Shale deposition; *C*, at end of Fort Payne deposition; *D*, at end of *Ullin Limestone* deposition. Modified from Lineback (1969).

linois grades laterally into equivalent *Borden Siltstone*, and the *Springville Shale* of southern Illinois represents distal equivalents of the *Borden*. The Fort Payne onlaps deltaic clastics of the *Borden*; the *Ullin* overlaps the Fort Payne and the *Borden*. Similarly, in west-central and east-central Kentucky, the Muldraugh and Renfro units onlap a Borden Formation detrital wedge. In southern Kentucky and Tennessee the relationships of Borden and Fort Payne rocks are uncertain, although most of the Fort Payne cherty rocks probably are of Muldraugh and perhaps younger ages.

Although the depositional framework just outlined has removed much previous confusion, the earlier facies concept may be appropriate in local areas. In northeastern Illinois and northwestern Indiana, cherty limestones interpreted to be Burlington and Keokuk correlatives appear to be lateral equivalents of Borden detrital strata (pl. 9-B, sec. d–d'). Interbedded mudstone and limestone adjoining the main masses of Burlington and Keokuk and Borden in parts of Illinois also suggest intertonguing relationships. Mudstones in the upper part of the Keokuk of western Illinois and northeastern Missouri may be tongues of the upper part of the Borden. Finally, the chronology for Illinois may not be applicable in detail to areas such as central and southern Kentucky, where carbonates in the Muldraugh Formation may be contemporaneous with the Fort Payne.

Relatives ages of the Borden delta of Illinois and the deltaic shelf deposits of Kentucky and Indiana are uncertain, but conodont data suggest that the basal Borden deposits in southern Indiana and west-central Kentucky may be somewhat younger than those in Illinois (Rexroad and Scott, 1964). Borden sediments then, may have entered Illinois from the northeast prior to encroaching into Kentucky and southern Indiana from more easterly source areas.

UPPER BOUNDARY OF INTERVAL B

Interval B rocks are overlain by interval C rocks throughout much of the Eastern Interior basin and northern and western Missouri. Isolated outliers of interval C rocks above more continuous interval B rocks across the Cumberland saddle permit correlation of interval B strata west of the saddle with those in eastern Kentucky. There, rocks at the interval B–interval C boundary are continuous along the margins of the Appalachian basin, except for areas in northeastern Kentucky where Pennsylvanian rocks unconformably overlie interval B. Pennsylvanian units also unconformably overlie interval B rocks in parts of eastern, central, and western Missouri and in an east-trending belt across north-central Illinois into northwestern Indiana. Interval B rocks are overlain by Cretaceous beds along the margins of the Mississippi embayment in southernmost Illinois, western Kentucky, and west-central Tennessee.

The boundary between intervals B and C, which is intended to correspond with the Osage-Meramec Series boundary in the Eastern Interior basin, is the most difficult of all interval boundaries to select consistently. Uncertainties of correlation result from disagreements regarding specific ages of faunal assemblages at about the Osage-Meramec Series boundary and from imperfect understanding of complex facies relations. Although most published reports refer to the Osage and Meramec Series designations, the Illinois State Geological Survey combines these into a single series, the *Valmeyeran*. Most geologists agree that no regional hiatus marks the series boundary in the Eastern Interior basin and that deposition was generally uninterrupted, although in some areas depositional environments changed markedly.

Regional Osage-Meramec Series boundary problems mostly have revolved about the age assignments of and correlations of the Warsaw Shale near its type area in western Illinois, the rocks termed Warsaw Limestone in Tennessee and Kentucky, and the Harrodsburg Limestone in south-central Indiana. As a discrete unit the Warsaw Shale is recognizable only in western Illinois and from there a short distance westward. Faunal elements in the Warsaw Formation have been variously assigned to the Meramec (Butts, 1922; Van Tuyl, 1925; J. M. Weller and others, 1948) or to the Osage (S. Weller, 1909), or both. Although most published reports and reviews (J. M. Weller and others, 1948; Keroher and others, 1966; Sando and others, 1969) place the unit in the Meramec, a number of workers recognize it as Osage (Moore, 1928; Weller and Sutton, 1940; Laudon, 1948; Wanless, 1957). Subsurface work indicates that shales in the type Warsaw Formation (Warsaw Shale of Lineback, 1966) merge eastward into the *Borden Siltstone* (Swann and others, 1965; Lineback, 1966; see pl. 9-B, sec. d–d'), indicating that the shales are Osage.

Nomenclatural history of Harrodsburg Limestone in Indiana is complex (Cumings, 1922, p. 493–499; Smith, 1965), and ages of the faunal elements, which have been compared to those in the Warsaw and the Keokuk of Illinois, have been debated (Laudon, 1948). In the Indiana outcrop, the Harrodsburg overlies the Borden Formation and underlies Salem Limestone. Currently, siliceous and dolomitic carbonate rocks that were in the lower member, the *Ramp Creek Member* of Stockdale (1929) of the Harrodsburg Limestone are included in the *Muldraugh Formation* of the Borden Group in Indiana (Smith, 1965). The younger *Leesville* and *Guthrie*

Creek Members and an uppermost unnamed division, the upper Harrodsburg, which is predominantly crinoid- and bryozoan-bearing limestone, have been retained in the Harrodsburg by Smith. The base of the upper Harrodsburg has been considered to be the Osage-Meramec Series boundary in Indiana Geological Survey reports since 1954, although the entire unit has been assigned to the Osage by others (for example, Stockdale, 1931, 1939). The Harrodsburg in west-central Kentucky is a lithologic and time equivalent of the Harrodsburg in Indiana and likewise excludes equivalents of the *Ramp Creek Member* (Sable and others, 1966). Of the 600-foot thickness of *Ullin Limestone* in parts of southern Illinois, more than 300 feet is assigned to the Harrodsburg. No regionally traceable horizons are known in the Harrodsburg in Illinois; therefore, an Osage-Meramec boundary within the Harrodsburg, similar to the horizon used in the Indiana outcrop, cannot be located in subsurface sections.

Subsurface correlations from western Illinois to south-central Indiana (Swann and others, 1965; Lineback, 1966) indicate that the *Ramp Creek* and the *Harrodsburg Members* of the *Ullin Limestone* in Illinois are younger than the Warsaw Shale and *Borden Siltstone*. The *Harrodsburg* in Illinois is generally correlative with the Harrodsburg in Indiana. In southwestern Illinois, the *Ullin* directly overlies the Warsaw (pl. 9-B, sec. $d-d'$); but farther north in western Illinois, the *Ullin* is absent, and the *Sonora Formation*, in part a time-equivalent of upper Warsaw and lower Salem beds (Collinson, 1964), generally overlies the Warsaw (pl. 9-B, sec. $a-a'$).

Strata that are termed Warsaw in Kentucky and Tennessee may have no relationship to the Warsaw in Illinois. Stemming from Butts' (1922) use of the name in western Kentucky for rocks overlying the Fort Payne Formation or Holtsclaw Sandstone, the term Warsaw remains in current use in parts of Kentucky and all of Tennessee. In Kentucky, in addition to this usage, the name Warsaw has been used in a restricted sense for rocks corresponding to the upper Harrodsburg of Indiana (McFarlan, 1943, p. 75). Because studies in Illinois indicate that the Warsaw Shale is equivalent to part of the *Borden Siltstone*, both usages of Warsaw in Kentucky and Tennessee may be invalid. Recent mapping by U.S. Geological Survey geologists in Kentucky confirms Stockdale's (1939) opinion that Butts' Warsaw unit includes lithologic and age equivalents of Salem Limestone (Meramec), Harrodsburg Limestone (Meramec and Osage), and *Muldraugh Formation* (Osage). Rocks called Warsaw in Tennessee, even farther from the type Warsaw in Illinois, are probably also correlative with part of the Salem, the Harrodsburg, and, possibly, the Muldraugh (pl. 15, col. 31).

In most of eastern Kentucky, the Renfro Member of the Borden Formation and younger beds in the Newman Limestone, directly overlie Borden detrital rocks. In outcrop, the Renfro lies between Borden detrital rocks and St. Louis Limestone equivalents in the Newman Limestone and is relatively thin. In south-central Kentucky, thin Harrodsburg beds appear in the middle part of the Renfro and thicken westward as the lower and upper parts of the Renfro grade respectively into the Muldraugh and the Salem (Weir and others, 1966).

The interval B–interval C boundary as used in this report is defined as follows for areas from east to west:

1. The base of the Newman Limestone or subsurface *Big Light* or *Big Lime* or *Greenbriar* in eastern Kentucky.
2. The middle of the Renfro Member of the Borden Formation in east-central and eastern Kentucky, where that unit is recognizable or recorded. Where the Renfro is roughly less than 20 feet thick, its base is the interval boundary. The Renfro grades westward into the Muldraugh, Harrodsburg, and Salem (Weir and others, 1966).
3. The base of the Harrodsburg Limestone, as currently used in Indiana and west-central Kentucky where the Harrodsburg is generally less than 50 feet thick.
4. Arbitrarily, within the upper or middle part of the Harrodsburg and its lithologic equivalents in southwestern Indiana, southern Illinois, and western Kentucky, where these rocks are thick. Large footage errors in thickness may have resulted from miscorrelations in these areas.
5. The base of the Salem and Warsaw undivided or Warsaw in south-central Kentucky.
6. Within the Warsaw Limestone in western Kentucky and northern west-central Tennessee, where the Warsaw is thick and suspected to contain *Ramp Creek* equivalents; most Harrodsburg-like lithologies are excluded from interval B. At the base of the Warsaw Limestone elsewhere in west-central Tennessee. The Warsaw Limestone of western Kentucky and west-central Tennessee, judging from published descriptions (Marcher, 1962, p. 16–19) and continuity with the Illinois sections, consists of several varieties of carbonate rock including Harrodsburg and *Ramp Creek* or Muldraugh lithic equivalents; Harrodsburg-like lithologies are conspicuous parts of thick Warsaw sections in extreme western Kentucky and northern west-central Tennessee; in general, *Ramp Creek*-like carbonate rocks seem to be more

abundant in the lower part of the Warsaw, although their relationships with Harrodsburg rock types appear to differ from place to place and to include interstratification. Consistently placed in the Meramec Series in published reports, the lower part of the Warsaw of these areas may be of Osage age, and it is here arbitrarily assigned to interval B.

7. The base of *Ullin Limestone* in west-central and western Illinois.
8. The top of the Warsaw Shale and *Sonora Sandstone* or top of sandstone-dominated sections in the Salem Limestone in western Illinois westward to approximately the Illinois River. West of this area, the boundary descends to the base of the Warsaw Shale in Illinois and follows the base of the Warsaw Formation in northern and western Missouri. This arbitrary westward lowering of the interval B–interval C boundary is the result of decisions made to accommodate coworkers in adjoining Iowa, Nebraska, and Kansas. There, the interval boundary is the base of the Warsaw despite possible affinity of that unit to the Osage strata in Illinois.
9. The top of the Warsaw in southeastern Missouri.
10. The top of the Short Creek Oolite Member of the Keokuk Limestone in southwestern Missouri.

In some areas, alternate boundaries can be drawn that just as reasonably approximate the Osage-Meramec Series boundary. In intervals B and C, changes in thickness of 30 to more than 50 percent in some areas might result from such changes in selecting the boundary. These areas include parts of southern Illinois, southwestern Indiana, western Kentucky, and northern west-central Tennessee where the Harrodsburg and its lithologic equivalents are thick. The lithofacies patterns that would result from assigning these different units to interval B would be little changed. In western Illinois, however, both thickness and lithofacies patterns are strongly affected, when the Warsaw Shale is placed in intervals B or C.

THICKNESS AND LITHOFACIES TRENDS

The thickest preserved interval B rocks are in the eastern part of the region (pl. 4-A). Thicknesses of more than 600 feet in northeastern Kentucky and more than 700 feet in west-central Indiana and central Illinois indicate locations where detrital rocks in the Borden are thick; isopachs generally trend subparallel to interpreted depositional strikes. Thicknesses of 700 to more than 800 feet are found in southern Illinois and western Kentucky. There, isopachs indicate maximum thickness of Fort Payne and *Ullin* carbonate rocks and broadly correspond to the Fairfield basin, although a well-defined basin is not clearly outlined. The thickest deposits occur in the southernmost parts of this area; the interval is abruptly truncated at its southwestern edge by pre-Cretaceous erosion.

Interval B thins southward in west-central Tennessee to about 200 feet and westward in Missouri to about 100 to 300 feet along the Kansas State line. In most of Missouri, only minor variations in thickness are recorded in the Burlington and Keokuk rocks which dominate the interval. These rocks thin to less than 100 feet in extreme northwestern Missouri. In southwestern Missouri, the *Pierson*, Reeds Spring, *Elsey*, Grand Falls, and associated interval B rocks exceed 300 feet in parts of an easterly trending belt which may have extended as far east as western Tennessee (pl. 4-A).

Some marked variations in thickness occur along the La Salle anticlinal belt in eastern Illinois and Indiana and from this belt southeastward to near the Cumberland saddle in Kentucky. Interval B rocks are relatively thin in several areas along this southeast-striking belt; the thin areas also partly correspond to those in which rocks of interval A are absent (pl. 3-A).

Patterns of interval B isopach and lithofacies trends are more easily interpreted in most of the region than are those of interval A. The lithofacies pattern for interval B (pl. 4-B) shows dominantly fine grained detrital rocks in a northwest-striking belt extending from eastern Kentucky across west-central Kentucky, central Indiana, and much of Illinois. In some areas, lobate tongues project southwestward from the belt, culminating in the large tongue, the Borden delta, which extends for more than 200 miles across central Illinois. An exception to the lobate lithofacies boundary is in south-central and west-central Kentucky, where the southwest limit of predominantly detrital rocks has a remarkably straight northwest trend. Elongate bodies of reefoid crinoidal limestone and sandstone lie southwest of the main belt of detrital rocks and closely parallel this trend.

Northeastward and eastward coarsening of detrital rocks in the Borden is shown by an increase of sandstone in eastern Illinois and to a lesser extent in northeastern Kentucky. Patterns indicating detrital tongues extending northwest from the main Borden delta in west-central Illinois may not accurately depict westward directions of dispersal in that the interval includes some detrital rocks of the Warsaw and *Sonora* Formations, as well as the *Borden Siltstone*, units whose sources may have been unrelated. The arbitrary exclusion of Warsaw and *Sonora* rocks from interval B west of the Illinois River also creates an inconsistent lithofacies pattern indicating a predominance of

carbonate rocks in western Illinois and adjacent States (p. 79). Inclusion of Warsaw and *Sonora* rocks in interval B throughout the area would shift the shale lithofacies patterns west and would show a strong detrital facies component well into northeastern Missouri and southeastern Iowa.

In area, carbonate rocks far exceed terrigenously derived detrital rocks in the Eastern Interior basin. Trends of lithofacies within the carbonate rocks in Illinois south and west of the Borden delta are not clearly shown on plate 4-B because the scheme of lithofacies representation does not differentiate the different types of carbonates and because the lithologic data used for the Fort Payne Formation is not consistent. A large area extending from south-central Kentucky southwestward through west-central Tennessee and northward into eastern Illinois and southwestern Indiana is occupied by carbonate rocks and chert in the Fort Payne, *Ullin*, and Muldraugh Formations. The lithofacies pattern there is complex with enigmatic trends. Siliceous rocks in the Fort Payne dominate volumetrically from southern Illinois across western and southern Kentucky and Tennessee. North of the area of siliceous rocks, thick bodies of *Ullin Limestone* or its equivalents are shown by areas of abundant carbonate and scarce chert in southeastern Illinois, southwestern Indiana, and western Kentucky. A southeast-trending belt of the same facies across west-central Kentucky is produced by thick carbonate units, probably equivalent to the Muldraugh.

In much of Kentucky and west-central Tennessee, rock units in the Fort Payne, in both outcrop and subsurface, have been described in widely varying terms, such as limestone, silty limestone, siltstone, dolomitic siltstone, silty dolomite, shaly limestone, shale, siliceous or cherty shale, siliceous or cherty limestone, chert, and silicastone. In some records the terms reflect real variations within the fresh rock or differences brought out by weathering; in others, they represent differences of opinion or different schemes for classifying fine-grained argillaceous-siliceous-calcareous rocks. Few quantitative analyses of the characteristically very fine grained Fort Payne rocks have been published. Consequently, the lithofacies patterns shown in western Kentucky and west-central Tennessee are largely generalized. High carbonate and silica content is characteristic of these rocks, but origin of the silica content is debatable. Analyses of the Fort Payne in Illinois show that quartz of silt and clay size is an important but variable constitutent of these rocks (Lineback, 1966, p. 23). Silica content in the Fort Payne has also been ascribed to chemical precipitation or replacement during deposition and diagenesis or to secondary causes, such as weathering (Bassler, 1932, p. 154–155). I believe that the detrital component shown in parts of western Kentucky and Tennessee is probably too low.

Very high carbonate ratios in Missouri and Illinois west of the Borden delta reflect the fairly uniform lithology of the Burlington and Keokuk undivided. Chert is estimated to compose slightly greater than 20 percent of the rock except in the central part of western Missouri where chert is not as abundant. The chert occurs mostly in the lower part of the unit. A general northward increase in fine detritals in the Keokuk of northeastern Missouri and increase of detritals northeastward in the Fern Glen are masked by the large carbonate content of the interval as a whole. In southwestern Missouri, chert increases and may compose more than 50 percent of most interval B units. There, Robertson (1967) classified chert in the *Elsey*, Grand Falls, and Reeds Spring Formations as: (1) interbedded chert originating probably as an early diagenetic replacement of limestone, (2) massive sheet chert, replacing limestone after lithification and some structural activity, or representing reconstitution of carbonate-chert beds, and (3) localized replacement chert, later than the massive sheet chert. The last two varieties are restricted mainly to the Tri-State mining district of southwest Missouri, Kansas, and Oklahoma. The interbedded chert is more widespread, but the source of the silica, which was deposited along with original carbonate material and was concentrated during diagenesis, is unknown; its volume appears to be more than can be attributed to siliceous sponges and other organic sources. Chert in the Burlington and Keokuk Limestones is generally considered to be an epigenetic replacement of carbonate rocks. Fowler and others (1934) favored a hydrothermal origin; Robertson (1967) believed that much of the chert is early diagenetic. Because this chert is a detrital constituent of conglomerate of Chester age, it did form during Mississippian time.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Sand grains in basal Fern Glen beds in areas marginal to the Ozark uplift (Kissling, 1961, p. 144) indicate that Ozark areas were sources for some of the detrital components of the Fern Glen. The distribution of the Fern Glen in Illinois and the increase in its detrital content away from the Ozark region, however, suggest that its distinctive red and green detrital components were derived from northerly sources, such as the Transcontinental and Wisconsin arches, during the post-Chouteau–pre-Burlington hiatus (Sable, 1970). The sediments accumulated along a shelf between deepening water to the southeast and lowland areas to the north and northwest. Other thin red mudstones in

central Indiana, Kentucky, and Ohio probably came from Cincinnati arch areas at this time (Sable, 1970).

Sandstones in the Borden are protoquartzites or subgraywackes (Potter and Pryor, 1961), in strong contrast to the clean orthoquartzites characteristic of sandstones in intervals C and D. Major source areas for interval B sandstones were east and north of the Appalachian basin (Potter and Pryor, 1961; Walker, 1962), in the Canadian Shield (Potter and Pryor, 1961), and possibly in northernmost Canada (Swann and others, 1965, p. 15). The volume of detrital sediments transported into the Eastern Interior basin region during interval B time, as shown by thicknesses and isopach trends, indicates that source areas in the eastern or northern Canadian Shield were dominant. Sources east and northeast of the Appalachian basin contributed a greater quantity of sediment to the Eastern United States, but only the distal portions of this large volume of detritus reached the Eastern Interior basin region.

Most of the detrital sediments in the Borden were transported down a southwest-dipping regional paleoslope across the Canadian Shield by the Michigan River system (Swann, 1963, 1964). Detrital sediments in the upper part of the Keokuk, the Warsaw, and the *Sonora* may also have been derived from these source areas, as indicated for sandstone in the Warsaw Shale (Lineback, 1968b, p. 17), but their distribution and relation to the Borden rocks suggest that the Transcontinental arch or Wisconsin arch may have contributed sediments to these units. Location of source areas for the fine detrital components of the Fort Payne is uncertain; if the silica contained in the Fort Payne is largely clastic detritus, distant southern sources may be indicated. Other indications supporting this possibility are the increased thickness of mudstones in the basal Fort Payne in some places of Tennessee and western Kentucky (pl. 9-B, secs. $c-c'$, $f-f'$). Detritus in the Fort Payne rocks of south-central Kentucky and adjacent Tennessee, however, may have been derived from northern and eastern sources, like that in the Borden clastics.

Evidence that the Borden of Illinois is a deltaic deposit (Frund, 1953; Swann and others, 1965; Lineback, 1966) includes recognition and tracing in the subsurface of such classic delta structures as bottomset, foreset, and topset beds, and fairly detailed geometry and internal features of the delta have been reconstructed. According to Lineback (1966), after deposition of silty and sandy carbonates of the Fern Glen, Burlington and Keokuk, crinoidal carbonate banks developed in shallow water in western Illinois. The banks developed to heights of 200–300 feet above the surrounding sea floor in virtually sediment-free water at the margin of a deeper water basin to the east. The eastern edges of banks generally coincided with eastern limits of Fern Glen sediments. The Borden deltaic complex then advanced west into Illinois. Deflection of deltaic sediments southward by the carbonate bank on the west determined the direction of growth of the long tongue-shaped Borden delta. Some deltaic sediments in the Warsaw partly overrode the crinoid banks. Subsurface tracing of foreset beds in the deltaic sequence indicates that sediment at the foot was deposited in water depths exceeding 600 feet. Several sandstone bodies are interpreted to be turbidites which were deposited largely on the prodelta plain (Lineback, 1968b). Following cessation of active delta growth, dark siliceous carbonate rocks of the Fort Payne were deposited in deep water and on forest slopes of the delta, partly filling depressions adjacent to the delta. Although deltaic structure of the Fort Payne has not been proved, a deltaic origin is suggested by convex-upward profiles of upper surfaces of the unit where it is thick (pl. 9-B, sec. $a-a'$) and by some depositional features. After deposition of the Fort Payne, an irregular submarine topography was left in southern Illinois. Deep, narrow depressions in the sea floor were filled by the *Ullin Limestone*, which overlapped the Fort Payne and eventually onlapped the Borden delta. Cross-stratified fossil-fragmental carbonate sediments, abundant in upper part of the *Ullin* or of the Harrodsburg, formed on a shallow-water platform on the Borden delta and accumulated in and filled the adjoining depressions, resulting in shallow water throughout the area.

Most workers in Kentucky accept a concept of deltaic deposition of Borden strata similar to that previously described for Illinois. Evidence includes westerly and southwesterly depositional dips in Borden detrital rocks in south-central and west-central Kentucky (Weir and others, 1966; Kepferle, 1968), and abrupt reciprocal thickness relationships between westward-thinning Borden siltstone and shale and the overlying Muldraugh carbonates, which are in contact at a discrete depositional interface (Kepferle, 1966; Peterson, 1966). In Kentucky, however, the depth of water probably did not exceed 300 feet and restricting crinoid banks west of the encroaching deltaic sequence were absent. Discontinuous crinoidal reefoid limestones in the Fort Payne (Thaden and others, 1961) and elongate barlike sandstone bodies (Cincinnati Univ. and Indiana Univ., 1972) lie parallel to and marginal to the main mass of Borden detrital rocks in Kentucky. Carbonate skeletal material probably was washed from topset or foreset slopes of the delta into deeper water where it filled depressions or was piled up in barlike deposits by longshore currents traveling parallel to the delta slope. Attempted reconstructions of the deposi-

tional basin in western Kentucky (pl. 9-B, secs. *a*–*a'*, *f*–*f'*) suggest an interdeltaic area.

The Cincinnati arch area was an intermittently emergent source area according to Walker (1962, p. 110) who showed a coarse detrital clastic lithofacies for Osage rocks along the arch in south-central Indiana. Lithofacies patterns for interval B (pl. 4-B), however, do not indicate the same abundance of coarse detritals, although some sandstones are present in the Borden of Indiana and west-central Kentucky. These could have been transported from Canadian Shield sources. Reconstructed isopachs (pl. 10, fig. 2) and lithofacies patterns suggest that initially the Cincinnati arch may have been a drainage divide for areas east and west of it, and that it may have contributed minor amounts of fine-grained sediment to interval B (Sable, 1970), but early in the interval the arch was overwhelmed by interval B deltaic sediments. The southern extension of the Cincinnati arch, the Nashville dome, also may have contributed fine sediments during interval B time (Walker, 1962, p. 110), but clastics near the dome seem more likely a part of Borden and Fort Payne detrital wedge derived from northern sources. The Nashville dome perhaps acted much like the northern part of the Cincinnati arch in deflecting some southwest-directed detrital sediments around it to either side. During much of interval B time it was probably a depositional site.

The interval B carbonate rocks west of the Borden delta accumulated in agitated shallow water of normal salinity. Onlap of the Burlington Limestone, and hence transgression of the sea, was northward in eastern Missouri and southeastern Iowa (Laudon, 1937) and southward and westward in southwestern Missouri (Thompson, 1967). Clear, warm seas that supported abundant marine life spread from the Appalachian basin westward in a shallow trough along the Arkansas-Missouri border and from there northwestward along the east side of the Ozark uplift (pl. 11, fig. 2). The water was deeper and more turbid east of the Ozark region compared to areas of the west. During its maximum extent in late Burlington time the sea entirely covered Missouri and probably connected with a very large sea to the west. Minor amounts of fine detrital material were shed southward or southwestward into northeastern Missouri and western Illinois during late Keokuk time. Somewhat greater amounts of detritus were shed into the region during deposition of the Warsaw Shale and *Sonora Formation*, but the depositional patterns did not otherwise change greatly during the transition from interval B to interval C.

Paleogeographic reconstruction (pl. 12, fig. 2) for late-middle interval B (late Borden) time indicates that shallow marine waters covered most of the areas. A southwest-trending, deepwater trough occupied southern Illinois and parts of adjoining States. A very shallow sea is inferred for most of western Illinois and Missouri. The Ozark region may have stood as an island of low relief in this sea. Subaerially exposed lands included low hills and plains in the Transcontinental arch extension of Wisconsin and Minnesota, and in the Canadian Shield to the northeast. Positions of river systems in the area of the Transcontinental arch are unknown inasmuch as their presumed deposits are now mostly eroded away. The Michigan River system fed the Borden delta and drained the northern or northeastern land area in Canada. Plains areas adjoining hilly land are inferred from occurrence of minor amounts of fine-grained detritus in upper- and post-Borden interval B rocks.

PALEOTECTONIC IMPLICATIONS

Parts of the Ozark region, northeastern Missouri, and western Illinois were differentially uplifted during early interval B. Farther east, in Illinois and Indiana, a differentially subsiding basin existed. Still farther east, a stable, submerged shelf was present. The Cincinnati and Kankakee arches, however, were probably relatively positive and were undergoing minor erosion.

The Eastern Interior basin region was generally a mildly negative cratonic region during interval B time. There are striking differences between the tectonic framework during interval B time and that which preceded it during the Late Devonian and interval A. In Illinois, a marked positive axis (Vandalia arch) in Late Devonian (fig. 13) and earliest Mississippian times became a trough during interval B. Conversely, areas in western Illinois were uplifted and partially eroded prior to deposition of the Burlington Limestone (p. 70), after which a stable shelf persisted there through most of interval B time. The areally restricted Fern Glen Limestone accumulated during at least part of the time when these changes took place.

The Cincinnati arch, including the Nashville dome, was a stable to mildly negative structure that separated slightly more rapidly sinking basins. The La Salle anticlinal belt and its southeastward extensions in Kentucky were stable to slightly uplifted. After uplift early in interval B time and mild erosion, the Ozark region and its marginal areas in western Illinois and Missouri became a slowly subsiding platform. North of the region, the Transcontinental and Wisconsin arches, emergent in early interval B time, were also stable or structurally negative later in the interval.

A major negative element (pl. 10, fig. 2) was a southward-deepening trough which extended from the Michigan basin southwestward across central and southern Illinois and southward to beyond present interval B erosion limits.

In southwest Missouri and northern Arkansas, thickness and relationships of interval B rocks suggest

that an elongate east-trending basin formed during early interval B time. The apparent westward thickening of the Fort Payne in western Tennessee (pl. 10, fig. 2) suggest that the basin may have extended eastward onto the flank of the Nashville dome. However, data concerning the basin are absent in the southern Ozark and Pascola arch areas.

INTERVAL C

INTRODUCTION

Interval C rocks in the Eastern Interior basin region are predominantly carbonate strata which accumulated on shelves and in shallow basins. Relatively pure limestones are dominant, sandy limestones occur in the eastern and northwestern parts of the region, and some units of sandstone and mudstone are present. Restricted seas and aridity are indicated by evaporite beds in parts of the region in mid-interval C time. A regional hiatus is not recognized within the sequence, although disconformities between some units are present in areas marginal to major positive elements. Present distribution of interval C rocks is considerably less than those of intervals A and B, owing to pre-Pennsylvanian, pre-Cretaceous, and Holocene erosion.

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

The succession characteristic of interval C is (1) fossil-fragmental carbonate rocks in the lower part deposited under high-energy conditions, (2) very fine grained chemically or organically precipitated evaporites and carbonates in the middle part deposited in quiet water, and (3) oolitic and sandy limestones and minor sandstone in the upper part deposited in very shallow water under high- and low-energy conditions.

Interval C units in the Eastern Interior basin region (pl. 15) include the St. Louis and Ste. Genevieve Members of the Newman Limestone in east-central and northeastern Kentucky, the Ste. Genevieve Member of the Monteagle Limestone in southeastern Kentucky, all or upper parts of the Harrodsburg Limestone and of the Warsaw Limestone of Kentucky and Tennessee and the Renfro Member of the Borden Formation of south and southeast Kentucky, the Warsaw Shale of extreme western Illinois and Missouri except for southeast Missouri, the Salem, St. Louis, and Ste. Genevieve (including its Rosiclare Sandstone Member) Limestones of western Kentucky, and the Aux Vases Sandstone and the Levias Member of the Renault Formation in Illinois. In Indiana, the Harrodsburg and Salem Limestones constitute the *Sanders Group*, and the St. Louis and Ste. Genevieve Limestones are included in the *Blue River Group*.

Erosion has removed most interval C rocks from the Cincinnati arch area; as a result all except the lower units in the eastern part of Kentucky are separated by at least 80 miles from their western counterparts. Lithologic and faunal evidence, however, indicates that all major interval C units were once continuous across the Cumberland saddle in southern Kentucky. Yellow silty dolomite at the base of interval C, the Renfro Member of the Borden Formation, occurs also at the base of the Newman Limestone in many eastern Kentucky outcrop areas. Correlation of the Renfro with subsurface units in eastern Kentucky is uncertain; it may be correlative with the Little Valley Limestone of the Appalachian basin. The Renfro grades westward into the Salem, Harrodsburg, and Muldraugh in south-central Kentucky (Weir and others, 1966). The crinoidal Harrodsburg Limestone consists of relatively pure crinoid-bryozoan limestone and extends northwestward across west-central Kentucky and Indiana to the Borden delta in central and southwestern Illinois where the Harrodsburg has been termed the *Kidd Member* of the Salem (Baxter, 1960). The Harrodsburg has not been recognized separately west of the Borden delta, but time equivalents may be present in the undivided *Ullin Limestone*. The Harrodsburg apparently pinches out eastward and southward in south-central Kentucky; equivalents of the Harrodsburg in western Kentucky and west-central Tennessee are called Warsaw.

In Missouri, fossiliferous limestone of the Warsaw Formation included in interval C is widespread but discontinuous. The Warsaw in northeastern Missouri and westernmost Illinois is also included in interval C, but the Warsaw in southeastern Missouri and in west-central and southwestern Illinois is placed in interval B as explained earlier (p. 78).

The Salem Limestone consists of locally dolomitic medium- to coarse-grained conspicuously cross-bedded fossil-fragmental oolitic and pelletal limestone and dolomitic limestone, admixed with mudstone or containing mudstone beds in some areas. The Salem thickens irregularly from generally less than 100 feet in its Indiana and Missouri outcrop belts to more than 400 feet in southern Illinois. In south-central Kentucky, most or all of the unit termed the Warsaw or the Warsaw and Salem undivided (Thaden and Lewis, 1962; Hamilton, 1963) is considered an approximate time equivalent of the Salem of Indiana. Mudstone, argillaceous limestone, and lesser amounts of sandstone constitute appreciable parts of these units in that area, and argillaceous rocks in the Salem are common to the north in Indiana as well as being present in western Kentucky and southern Illinois. In Tennessee, the Salem is not recognized by name, but beds in the lower part of the St. Louis and upper part of the Warsaw are lithologic and faunal Salem equivalents. In most of the Eastern Interior basin region, the St. Louis

Limestone conformably succeeds the Salem, but in parts of western Illinois, St. Louis overlies disconformably the *Sonora Sandstone* or sandy Salem beds. Regionally, the Salem probably represents only an approximately equivalent unit which is in part a lateral time equivalent of the early St. Louis (Lineback, 1972). In southern Illinois, the Salem thickens and is darker and more argillaceous than it is in the Indiana outcrop belt. In the area where the *Sonora Sandstone* is present on the northwest margins of the basin, sandstone and sandy dolomitic limestone are included in a thin unit referred to the Salem. Westward, in northeast Missouri, rocks assigned to the Salem include interbedded green mudstone and fossil-fragmental limestone; mudstone content decreases southward. The Salem in western and southwestern Missouri is tentatively identified by the contained fauna (Spreng, 1961, p. 69), but the Salem is missing in extreme southwestern Missouri where rocks of interval D overlie the Warsaw and Keokuk Formations (pl. 15, col. 56).

Early workers suggested that hiatus or disconformity separated the St. Louis from older rocks in the Mississippi Valley (Weller, 1909; Van Tuyl, 1925); more recent studies indicate that conodont faunas are transitional upward into basal St. Louis beds. However, a gap in the conodont succession within the lower part of the St. Louis in western Illinois (Collinson, 1964, p. 9) may mark a fairly widespread hiatus.

The St. Louis Limestone consists mostly of very fine grained to sublithographic carbonate rock, containing little detrital material. The St. Louis is more than 400 feet thick in southern Illinois and western Kentucky, but it thins markedly along the margins of the Cincinnati arch and Ozark uplift. Much of the basinward thickening of interval C rocks is due to thickening of the lower part of the St. Louis and upper part of the Salem. Gypsum and anhydrite occur in the lower part of the St. Louis in several large areas along a southeast-trending belt from west-central Illinois into west-central Kentucky (Saxby and Lamar, 1957; McGregor, 1954; McGrain and Helton, 1964). Terrigenous clastic material includes thin carbonaceous mudstone and sandstone beds in the lower part of the formation in west-central and south-central Kentucky, respectively, and thin green to gray mudstone at various horizons in many areas. Intraformational limestone breccias occur in the lower part of the formation in Mississippi Valley exposures and in adjacent areas of Illinois, Missouri, and Iowa.

The contact of the St. Louis with the overlying Ste. Genevieve Limestone is based on lithologic differences and is regionally inconsistent (Swann, 1963, p. 27). Some abrupt vertical offsets of this contact (pl. 9-C, secs. $a-a'$, $c-c'$; Ulrich and Klemic, 1966) result from differences in contact criteria, lack of widespread datums, lack of detailed faunal studies, and characteristically poor exposures of St. Louis.

The Ste. Genevieve Limestone is a tabular predominantly carbonate unit made up largely of oolitic and fossil-fragmental limestones. Mixtures of carbonate and quartz sand are important constituents of rocks named Ste. Genevieve in eastern Kentucky and to a lesser extent in northern Missouri and western Illinois. Some quartz sand and sandstone are also present in the vicinity of the Ste. Genevieve type locality in southeastern Missouri. Thin mudstones occur at several stratigraphic horizons in Illinois, Indiana, and Kentucky, and one discrete widespread detrital unit, the *Spar Mountain Sandstone Member* (Tippie, 1945), interrupts the carbonate sequence in most of Illinois (pl. 9-C, secs. $a-a'$, $c-c'$, $d-d'$) and parts of adjoining States. In Illinois and western Kentucky several members include, in ascending order: the Fredonia, *Spar Mountain*, *Karnak*, and *Joppa*, which underlies the Aux Vases Sandstone. In western Kentucky, the uppermost members of the Ste. Genevieve are the Fredonia, Rosiclare, and Levias. In Illinois, the Rosiclare is equated with and generally superseded by the name Aux Vases Sandstone, and the Levias is the basal member of the Renault Formation. The Ste. Genevieve and overlying units were fully summarized by Swann (1963).

The Ste. Genevieve is overlain disconformably by the Aux Vases Sandstone in southeastern Missouri, Illinois, and southwestern Indiana. Where the Aux Vases is absent or not recognizable, as in much of Kentucky and Indiana, the Ste. Genevieve is overlain by interval D beds of the Renault, Paoli, Girkin, and Monteagle Limestones. Swann (1963, p. 32) indicated that the upper surface of the Ste. Genevieve is younger in western Kentucky than in southern Illinois because of miscorrelations of detrital units in the Ste. Genevieve (upper part) and Aux Vases. In northeastern Kentucky, diagnostic fossils are scarce in the Newman Limestone, and sandy carbonate rocks ascribed to the Ste. Genevieve are placed at the top of interval C.

The Aux Vases Sandstone includes sandy equivalents of *Spar Mountain*, *Karnak*, and *Joppa Members* of the Ste. Genevieve and is overlain by interval D Renault Limestone or by interval C Levias Limestone Member of the Ste. Genevieve. The Levias is considered partly equivalent to the *Bryantsville Breccia Bed* of Indiana (Swann, 1963).

UPPER BOUNDARY OF INTERVAL C

Interval C rocks are overlain by Mississippian rocks of interval D and by Pennsylvanian, Cretaceous, or Quaternary units. In several areas, Pennsylvanian rocks overlie rocks of interval C with a low degree of discordance or with disconformity. In the northern part of the Eastern Interior basin, Pennsylvanian rocks rest

on successively older interval C beds northward, and in north-central Illinois they truncate the interval B–interval C and older boundaries. In northern, western, and central Missouri, Pennsylvanian units lap onto the unconformity at the top of interal C and rest on progressively older rocks toward the Ozark region. Cretaceous rocks rest unconformably on interval C strata in westernmost and southernmost Illinois, in western Kentucky, and in west-central Tennessee. Interval D rocks overlie interval C strata in the Eastern Interior basin in southeastern Missouri, western and most of eastern Kentucky, southwestern Indiana, and central and southern Illinois.

West of the Cincinnati arch, the boundary is located on the basis of crinoids between occurrences of *Platycrinites penicillus* (Meramec) and *Talarocrinus* spp. (Chester). Other criteria for recognition of the series boundary west of the Cincinnati arch were given by Swann (1963, p. 32–33). The Meramec-Chester boundary (table 2) corresponds closely to the top of the Aux Vases Sandstone in southeastern Missouri and in southwestern to southern Illinois, to the top of the Levias Limestone Member of the Ste. Genevieve Limestone in parts of southern and eastern Illinois and western Kentucky, and to the top of the Ste. Genevieve in Indiana and west-central and southern Kentucky, which locally corresponds to the top of the *Bryantsville Breccia Bed* of the Levias Member. Locally discordant relationships at the contact, presence of sedimentary breccias, and the gap in the crinoid succession indicate hiatus at the interval C–interval D boundary (Weller and Sutton, 1940, p. 819–822); however, most modern workers consider depositional gaps at or near the boundary to represent only a short time (Swann, 1963).

In the eastern Kentucky outcrop belt adjacent to the Cincinnati arch, the upper boundary of interval C is drawn at the top of the Ste. Genevieve Member of the Monteagle or Newman Limestone. Criteria for the boundary within the Monteagle in the eastern part of southern Kentucky are both paleontologic and lithologic (Lewis and Thaden, 1965b); the boundary is placed at the top of one of several limestone breccia beds referred to the *Bryantsville Breccia Bed* by McFarlan and Walker (1956). Farther north, in east-central and northeastern Kentucky, where the Ste. Genevieve is the top of interval C, the upper boundary is not well defined lithologically (McFarlan and Walker, 1956, p. 10). The boundary is even more difficult to determine in the subsurface of eastern Kentucky. There, it is largely delineated arbitrarily on the basis of projected regional trends, and, in a few wells, by the presence in interval D of reddish beds which are thought to correlate with the Taggard Red Member of the Greenbrier Limestone (interval D) of the Appalachian basin. Evidence for disconformity at the interval C–interval D boundary in northeastern Kentucky consists of Chester equivalents in contact with St. Louis or older beds in Menifee and Rowan Counties. The boundary there is interpreted to represent overlap on a topographic and (or) structural high (McFarlan and Walker, 1956, p. 11).

THICKNESS AND LITHOFACIES TRENDS

Original thicknesses of interval C rocks in the Eastern Interior basin region are limited to parts of southeastern Missouri, Illinois, Indiana, and Kentucky. Pre- or intra-Pennsylvanian erosion has removed much of these strata from former areas of deposition in west-central Indiana, central and western Illinois, most of Missouri, and parts of northeastern Kentucky. Pre-Cretaceous erosion has removed very thick sections of interval C rocks in southernmost Illinois and westernmost Kentucky. Thicknesses of interval C (pl. 5-A) range from 400 to more than 1,100 feet. Thickest accumulation of sediment was in southern Illinois, western Kentucky, and southwestern Indiana. Marginal to the Cincinnati arch in eastern Kentucky, interval C is generally less than 200 feet thick.

Lithofacies patterns of interval C rocks are relatively simple and show carbonate rocks (pl. 5-B) throughout the area of uneroded interval C. The bordering areas in which detrital components are in sufficient abundance to affect lithofacies patterns are those in which the upper predominantly limestone units, such as the St. Louis and Ste. Genevieve, have been eroded; fine detritals in the lower units, such as the Salem and in Kentucky-Tennessee the Warsaw, make up a corresponding larger proportion of the rock that is left. The effect of mudstone in the Salem and Warsaw shows most strongly on the lithofacies map of interval C in south-central and west-central Kentucky; however, mudstone is present in about the same amount from southern Indiana to the Nashville dome.

Varying proportions of sandstone, mudstone, and carbonate in central and western Illinois reflect variations in the Salem Limestone and *Sonora Sandstone.* Sandstone extends a short distance south of this area (pl. 9-C, sec. a–a'), but its presence is masked by overlying carbonate rocks. Mudstone makes up an especially large component of the interval in western Illinois and adjacent Missouri where the boundary between intervals B and C was lowered to include the Warsaw Shale in interval C. (See p. 79.) Sandy limestone and sandstone beds in the Ste. Genevieve do not appreciably influence the dominantly carbonate lithofacies in northeastern Missouri, but an increase in the sand fraction is shown to the north in Iowa (pl. 5-B). Similarly, quartz sand in Ste. Genevieve rocks of northeastern Kentucky and in their equivalents in southern

Ohio and West Virginia is shown by the mixed sandstone-carbonate lithofacies pattern of those areas and of areas farther south in eastern Kentucky.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Although all preserved interval C rocks were deposited in marine environments, the distribution of terrigenous detrital material indicates transport from source areas mostly east and northeast of the Appalachian basin, and also along the Transcontinental arch in Wisconsin and Minnesota. Lesser sources of sediment may have been east of the present-day southern Appalachian Mountains, and small amounts of sediment may have been shed from areas of low relief along the Cincinnati arch and Ozark uplift. Williams (1957, p. 315 −316) suggested three source areas for detritals in Warsaw and Salem Formations of this region — Wisconsin, Ozarkia, and Appalachia. Rubey (1952, p. 50) reported sand grains derived from igneous and metamorphic rocks in the St. Louis Limestone of eastern Missouri and suggested the presence of exposed land areas in that region.

A northeastward-increasing sand content in the Ste. Genevieve Limestone of northeastern Kentucky, southern Ohio, northern West Virginia, and western Pennsylvania indicates a northeastern source area. Sandy debris is rare in the Ste. Genevieve immediately west of the Cincinnati arch, which suggests that the arch may have been a barrier to the dispersal of detritus from northeastern sources.

Evaporites in the St. Louis Limestone (pl. 9-C, secs. a–a', c–c', d–d', e–e') in Illinois, Indiana, and western Kentucky are indicators of restricted circulation in a local basin and high rates of evporation (Whiting, 1959, p. 9). Limestone breccias are widespread in the St. Louis Limestone west of the evaporites and may indicate dissolved evaporite deposits (Collinson, 1964, p. 7).

Source areas and transport directions are not known for the conspicuous mudstone and sandstone in the Salem and Warsaw in the Cumberland saddle area of south-central Kentucky and adjacent Tennessee and for mudstone along the west side of the Cincinnati arch in Kentucky. Land of low relief in the Cincinnati arch and Nashville dome areas may have contributed these sediments, or they may have been swept westward from sources east of the southern Appalachian Mountains through the Cumberland saddle to the west flank of the Cincinnati arch. The latter possibility seems more likely because eroded pre-Mississippian units on the Jesamine or Nashville domes probaby did not contain sufficient sand to be the source. Possibly, the mudstone and sandstone represent reworked interval B sediments of the Appalachian area.

The distribution of sandy units in the Salem and Ste. Genevieve of northern Missouri, Iowa, and western Illinois and the southward intertonguing of the detrital beds with dominantly carbonate rocks are evidence that the sand and silt were derived from northern source areas west of the Michigan basin, probably along the Transcontinental arch. The evidence is not as definitive in the Salem as in the Ste. Genevieve. The distribution of sand in the Salem in western Illinois can also be explained by a northeasterly source from which sediment was dispersed southwestward across the southeast margin of the Michigan basin. Across the central part of the region, however, sandy beds are not present in interval C in northwestern Indiana as might be expected with a northeasterly source. Source areas of small size in the Ozark region probably contributed to the Ste. Genevieve in southeastern Missouri.

The *Spar Mountain Sandstone Member* of the Ste. Genevieve Limestone and the Aux Vases Sandstone are the two major detrital tongues in the uppermost part of the otherwise carbonate-dominated interval C sequence. Sand in both units was considered by Swann (1963) to have been transported from northeastern source areas by a southwest-flowing river system, the Michigan River. Crossbeds in the Aux Vases of southwestern Illinois (Potter, 1963, p. 69) indicate that transport directions there were to the southeast, suggesting that if the Michigan River was the source, its sediment moved south along the west side of the Eastern Interior basin during Aux Vases time. Contributions of sediment from Transcontinental arch highlands to Illinois and Missouri by a separate southeast-flowing river system is also a possibility.

Depositional environments for interval C sediments were largely shallow marine except for the darker and finer grained rocks of the Salem and St. Louis in the southernmost part of the Eastern Interior basin where the water may have been moderately deep (pl. 11, fig. 3). The textures of the carbonate rocks reflect a wide variety of depositional conditions. Waters ranged from clear to turbid and from agitated to quiet. In general, agitated water is indicated by fossil fragmental, crossbedded, and interlensing beds in the Harrodsburg, Salem, Warsaw of Kentucky-Tennessee, Ste. Genevieve, and Aux Vases strata, whereas quiet water is inferred for St. Louis evaporites and lutitic and micritic, generally tabular-bedded carbonate rocks.

The Harrodsburg Limestone has not been studied in detail, but its relatively pure carbonate composition, abundant disarticulated fossil remains, and crossbedding indicate a striking change from turbid water, in which the underlying Borden Formation was deposited, to widespread clear water. Fossil debris, probably the disarticulated hard parts of fauna indigenous to the

delta platforms, were swept by currents into adjacent depressions on the sea floor of the platforms and delta slopes.

Pinsak (1957) concluded that nearshore, shallow-water offshore (shelf), and deepwater offshore (basin) environments are represented in limestone of the Salem Limestone in Indiana. The shallow-water offshore environment of Pinsak is compared to the environment on the present-day Bahama Banks (Indiana Univ., 1966). Clear-water conditions are shown by the Salem in Indiana, but markedly more turbid water is indicated by the argillaceous content of the Salem and Warsaw of Kentucky and Tennessee and southwestern and southern Illinois. Clear, agitated, freely circulating water is suggested by Salem and Warsaw rocks in most of Missouri.

A distinct change to low energy conditions and shallow water is reflected by rocks of the St. Louis Limestone (pl. 11, fig. 3). Carbonaceous mudstones and limestones in lower beds of the St. Louis of west-central Kentucky, possible time equivalents of the adjoining evaporites, may indicate that land along the Cincinnati arch acted as a barrier to free circulation, a factor favorable to evaporite deposition. Open seas probably lay to the south. A seaway may have intermittently connected the Michigan basin sea across the Kankakee arch.

Rocks in the upper part of the St. Louis largely represent a general transition to clear, freely circulating but quiet water in which both chemically and organically precipitated carbonates could accumulate. A major transgression during this time probably inundated all or large areas of the Ozark region, as well as the Cincinnati and Kankakee arches. Fine grain size, tabular beds, and general scarcity of terrigenous clastic debris indicate that wave and current activity were weak and that source areas were low or remote, perhaps like the present environment of Florida Bay. Dolomitic beds, the relative scarcity of fossil remains, and scattered occurences of gypsum suggest that a hypersaline environment existed in some areas throughout much of St. Louis time.

During deposition of the Ste. Genevieve, clear very shallow seas with normal circulation prevailed. The waters were alternately quiet and agitated. A general vertical transition from very shallow water bay to open water bank environment occurred from St. Louis to Ste. Genevieve time in much of the basin, as indicated by an upward increase of oolitic limestones, the current structures, and prolific shallow-water faunas and sedimentary limestone breccias. Comparable rocks are forming in the present Bahama Banks (Whiting, 1959, p. 9). Terrigenous clastics of the *Spar Mountain*, Aux Vases, and less widespread units in the central and southern parts of the basin are sheetlike deposits that periodically interrupted the carbonate regime. These clastics represent the onset of cyclical deposition which characterizes marine, marginal, and continental environments displayed by interval D deposits.

Paleogeographic interpretation indicates that during most of interval C time the region was covered by a shallow sea. Seaway connections existed mainly to the south, between the Ozark uplift and the Nashville dome, and intermittent connections with the Appalachian basin may have been through the Cumberland saddle.

Paleogeographic reconstruction (pl. 12, fig. 3) for interval C time shows a topographically low peninsula along the Cincinnati arch southward into Kentucky. During early St. Louis regression, low-lying land may have extended northwestward along the Kankakee arch. A low peninsula also extended southeastward from hilly land in Minnesota and Wisconsin during early St. Louis time. Small river systems flowed southward and southwestward from the Minnesota-Wisconsin uplands. In late interval C (Ste. Genevieve) time, an island in the Ozark region probably shed a small amount of sediment eastward.

Evidence for the Michigan River system, prominent in interval B and interval D times, is lacking in all except the youngest units in the Eastern Interior basin. During interval C the river may have had little sediment-carrying capacity, because of widespread aridity or because it headed in lowlands. Most likely, the point where the river debouched into the sea was too far north for much sediment to reach the Eastern Interior basin region.

PALEOTECTONIC IMPLICATIONS

Trends and positions of major tectonic elements inherited from interval B time changed little during interval C except that the basins and uplifts were somewhat accentuated. The Eastern Interior basin was a major negative feature during interval C as shown by subsidence of more than 1,100 feet in the southern part of the basin and thinning of interval C toward such positive features as the Cincinnati, Kankakee, and Mississippi River arches (pl. 5-A). Maximum subsidence of the basin occurred during Salem and early St. Louis time. This major differential downwarping ceased in late St. Louis time, and relatively even subsidence took place over wide areas. This subsidence is shown both on the isopach map of interval C (pl. 5-A) and, by the restored isopachs, on the paleotectonic map (pl. 10, fig. 3). The restored isopachs show that interval C sediments thinned markedly onto the Ozark uplift, perhaps more abruptly than indicated on the reconstruction. Thickness changes may have been partly controlled by

a southeast-trending hingeline across southeastern Missouri along the trend of the Wittenberg trough of Devonian age and the post-Mississippian Ste. Genevieve fault. The Cincinnati arch and Ozark uplift restricted seaways during St. Louis evaporite deposition, although they contributed little detrital sediment. The areas in Missouri north and west of the Ozark uplift are interpreted to have been mildly subsiding shelves on which a fairly uniform thickness of sediment accumulated. In western Missouri, southward thickening of the Warsaw Formation (pl. 9-C, sec. *b*–*b'*) indicate minor subsidence producing a shallow basin in this area. Local disconformable relationships between rocks of intervals D and B in southwestern Missouri, not shown on plate 9-C, suggest that at least parts of that area were emergent during interval C time.

The gap in the conodont succession in the lower part of the St. Louis Limestone may mark a tectonically activated hiatus. If so, movements were even, and the resulting unconformity has not been widely recognized.

In eastern Kentucky, thin shallow-water sediments imply that a platform existed between the Cincinnati arch and the Appalachian basin trough. The platform was modified in southeastern Kentucky by a shallow basin and in northeastern Kentucky by small arches which were islands during part of interval C time (pl. 10, fig. 3).

INTERVAL D

INTRODUCTION

Interval D rocks are characterized by cyclic alternations of detrital and carbonate strata. The interval is preserved in three separate areas: from south-central Kentucky to central Illinois and southeastern Missouri, in eastern Kentucky, and in southwestern Missouri (pls. 6-A, 15). Specific correlations between the somewhat different sequences in the three areas are uncertain. The interval D sequence is the Chester Series in the Eastern Interior basin and its age equivalents in the other areas.

A voluminous literature has developed dealing with many aspects of Chester rocks in the Eastern Interior basin. Swann (1963) reviewed the nomenclatural history of units; discussed time and rock stratigraphy, biostratigraphy, and depositional framework, and he included a pertinent bibliography. Chester sandstones were studied by Potter and others (1958) and by Potter (1962, 1963). Conodont studies include those from the type Chester area in southwestern Illinois (Rexroad, 1957). In Indiana rocks of Chester age were described by Malott (1952), and in western Kentucky, by Ulrich (1917), Butts (1917), and McFarlan and others (1955). Chester equivalents in eastern Kentucky were studied by Butts (1922), McFarlan and Walker (1956), and Vail (1959). Southwestern Missouri interval D rocks have been reviewed by Spreng (1961, p. 77–78), and their paleontology has been discussed by Wright (1952).

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Rhythmically deposited units of limestone and clay- to sand-sized terrigenous clastics characterize the interval D sequence in the Eastern Interior basin. Subdivision into more than 20 formations is shown in the correlation chart (table 2). Formations have been defined in outcrops in the Chester district in southwestern Illinois, in southern Illinois, in western, west-central, and south-central Kentucky, and in south-central Indiana. Correlations of nomenclatural units are generally consistent across State boundaries, although some units that are not divisible in some areas are readily divided in others. Physical criteria for some formational boundaries are different in adjoining States, and variations in terminology and groupings of units occur from State to State.

Interval D strata in the Eastern Interior basin comprise five formal groups in Illinois and three in Indiana. Groupings are based on lithologic similarity. Facies changes were produced by shifting loci of terrigenous clastic accumulations (Swann, 1963, p. 21–22). Time-stratigraphic divisions include the successively younger *Gasperian*, *Hombergian*, and *Elviran Stages* in Illinois (Swann, 1963, p. 21–23); the stage boundaries are considered to closely correspond to rock stratigraphic boundaries.

Dominantly detrital units in the Eastern Interior basin, in general ascending order, consist of the Renault Formation, Yankeetown Chert, Bethel Sandstone or Mooretown Formation, Ridenhower Shale or Sample Sandstone, Cypress Sandstone or *Elwren Formation*, *Fraileys Shale* and its equivalent, the Big Clifty Sandstone (locally a member of the Golconda Formation), the Hardinsburg, Tar Springs, Waltersburg, Palestine, and Degonia Sandstones, and *Grove Church Shale*. The Leitchfield Formation of west-central Kentucky consists largely of mudstone equivalents of the Tar Springs and younger clastic units; the Buffalo Wallow Formation is similar lithologically and overlies the Tar Springs beds. The preceding units contain minor limestone, but are composed mostly of varying proportions of mudstone and orthoquartzite sandstone which regionally constitute roughly 50 and 25 percent respectively of interval D rocks. The proportions and thicknesses of mudstone and sandstone show great lateral variation. In many places thick sandstones compose most or all of a formation and their lower boundaries are erosional. Truncation of underlying beds ranges from intraformational to removal of one or more

underlying formations. The morphology and internal features of the sandstones establish many of them as bar-finger and channel-fill deposits; some appear to be sheet sands, and upward-convex profiles of others suggest bar or beach origin. Although rarely recognized, channel fills of mudstone or of admixtures of mudstone and sandstone are also present. Marine fossils occur in the lower parts of many sandstones. Coaly beds and unfossiliferous red shales, considered to be continental, are present in the upper parts of many detrital-dominated units in some areas.

A striking example of a channel-fill deposit is exhibited by Bethel Sandstone or Mooretown Formation in western Kentucky. This deposit extends for more than 150 miles and locally cuts through more than 250 feet of pre-Bethel strata (Reynolds and Vincent, 1967). It is interpreted to have been a unique submarine channel (Indiana Univ. and Cincinnati Univ., 1969). Other elongate sandstone bodies have integrated distribution patterns suggesting a stream network (Potter, 1962, 1963; Sullivan, 1972).

Widespread units that are mostly limestone in the Eastern Interior basin include parts of the Renault Formation, and the Paoli, Girkin, Beech Creek, *Haney*, Golconda, Glen Dean, Vienna, Menard, Clore, and Kinkaid Limestones. They consist of relatively pure limestones and include micrites and sparites, and fossil-fragmental, oolitic, pelletal, argillaceous, and sandy limestone varieties. Chert and silicified limestone are relatively rare, although they characterize some units. Small amounts of mudstone and sandstone are interbedded with limestone in some areas. Mudstone is generally more abundant in the middle and upper Chester units, such as the Glen Dean, Menard, and Clore, than in the lower ones. Thin and discontinuous shale beds in the Girkin Formation of south-central Kentucky are distal facies of thicker dominantly sandstone and shale formations to the north and west. The lower surfaces of limestone units are commonly more planar than those of detrital units, although gradational and interfingering relationships between limestone and underlying mudstone and sandstone and local thickening of limestone units at the expense of underlying clastic units have been recognized. Several limestone units in the Leitchfield and Buffalo Wallow Formations in central Kentucky are continuous with the Vienna, Menard, and Kinkaid of western areas (Stouder, 1938; A. E. Smith, written commun., 1965).

In the eastern Kentucky outcrop belt, interval D rocks consist of a lower unit that is predominantly limestone and an upper unit that is mostly mudstone. In southeastern south-central Kentucky (pl. 15, col. 41), the lower unit consists, in upward succession, of the upper limestone member of the Monteagle Limestone (Lewis and Thaden, 1965b), the Hartselle Sandstone, and the Bangor Limestone. Northeastward, the Hartselle becomes unrecognizable, and the upper limestone member of the Newman Limestone, including Monteagle and Bangor equivalents, constitutes the lower carbonate part of interval D. The carbonate rocks are relatively pure; their gross features resemble those of Chester limestones in western Kentucky, and minor discontinuous beds of mudstone reflect roughly rhythmic depositional cycles similar to those in the Girkin Limestone. The Hartselle Sandstone, the principal noncarbonate unit in the carbonate-dominated section, consists of sandstone and mudstone lithologically similar to those in western Kentucky. The overlying Bangor Limestone is somewhat more argillaceous than the limestones underlying the Hartselle and in part bears a fauna similar to that of the Glen Dean Limestone of western Kentucky. Detailed correlation between eastern and western Kentucky has been attempted by McFarlan and Walker (1956) who considered the Hartselle equivalent to the Hardinsburg Sandstone. The Hartselle has been correlated with the Big Clifty by others (Peterson, 1956; Vail, 1959) on the basis of petrologic studies and correlation of overlying and underlying limestones.

The Pennington Formation is the uppermost interval D unit in eastern Kentucky. The Pennington is largely composed of mudstone in outcrop, but eastward, in the subsurface, sandstones become major constituents and include correlatives of the Stony Gap Member of the Hinton Formation of West Virginia. The Pennington has been raised to group rank where it is divisible into several formations (Vail, 1959). The Pennington is approximately equated with post-Glen Dean rocks of western Kentucky (McFarlan, 1943, p. 95), but detailed correlations of intra-Pennington units with the upper Chester rocks of western Kentucky have not been established.

Incompletely preserved outliers of interval D rocks occupy small areas in southwestern Missouri (pl. 15). They include in ascending order, three conformable formations: the Hindsville, Batesville, and Fayetteville. Another unit, the Carterville Formation, locally fills sinks in rocks underlying interval D. The Carterville consists of clay and conglomerate, sandstone, and dark mudstone. The Hindsville is dominantly crossbedded crystalline to oolitic limestone with locally interbedded mudstone and siltstone; it overlies interval B rocks. The fauna of the Hindsville has a lower Chester aspect but is also similar to Meramec forms (Spreng, 1961, p. 77). The Batesville is a calcareous sandstone interbedded with thin oolitic limestone, and the Fayetteville is dominantly black carbonaceous shale with small amounts of dark ferruginous limestone interbedded. Correlations with Eastern In-

TABLE 2.—*Stratigraphic units assigned to interval D in the Eastern Interior*

Map intervals	KENTUCKY				INDIANA	ILLINOIS
	East-central and northeastern	Southeastern and south-central	Central	Western	Southern	Southeastern
Overlying units	Lee Fm. Pennsylvanian	Lee Fm. Pennsylvanian	Caseyville Formation Pennsylvanian	Caseyville Formation Pennsylvanian	Mansfield Formation Pennsylvanian	Caseyville Formation Pennsylvanian
Interval D	Pennington Formation	Pennington Formation	Leitchfield Formation; Buffalo Wallow Formation; Tar Springs Fm.	*Grove Church Shale* equivalent; Kinkaid Ls.; Degonia Ss.; Clore Ls.; Palestine Ss.; Menard Ls.; Waltersburg Ss.; Vienna Ls.; Tar Springs Ss.	*Kinkaid Ls.*; *Degonia Ss.*; *Clore Ls.*; *Palestine Ss.*; *Menard Ls.*; *Waltersburg Ss.*; *Vienna Ls.*; Tar Springs Ss.	Kinkaid Ls.; Degonia Ss.; Clore Ls.; Palestine Ss.; Menard Ls.; Waltersburg Ss.; Vienna Ls.; *Tar Springs Ss.*
	Newman Limestone: Upper limestone member	Bangor Limestone	Glen Dean Limestone; Hardinsburg Ss.; Haney Ls. Member; Big Clifty Ss. Member; Beech Creek Ls. Mbr. (Golconda Fm.); Girkin Formation: Paoli Limestone	Glen Dean Ls.; Hardinsburg Ss.; Golconda Fm.; Cypress Ss.; Paint Creek Shale; Bethel Ss.; Renault Fm.	*Stephensport Group*: *Glen Dean Ls.*; *Hardinsburg Ss.*; *Golconda Limestone*; *Big Clifty Ss.*; *Beech Creek Ls.*; *West Baden Group*: *Elwren Fm.*; Reelsville Ls.; Sample Ss.; Beaver Bend Ls.; Bethel (or Mooretown Fm.) Ss.; *Blue River Group*: Paoli Limestone	Glen Dean Ls.; Hardinsburg Ss.; *Golconda Group*: *Haney Ls.*; *Fraileys Shale*; *Beech Creek Ls.*; *West Baden Group*: Cypress Ss.; Ridenhower Shale; Bethel Ss.; *Cedar Bluff Group*: *Downeys Bluff Ls.*; Yankeetown Chert; Renault Formation: *Shetlerville Ls. Mbr.*; Levias Ls. Mbr.
	Newman Limestone: Ste. Genevieve Ls. Mbr.	Monteagle Limestone: Upper limestone member: Ste. Genevieve Ls. Mbr.	Ste. Genevieve Limestone	Ste. Genevieve Ls.: Levias Ls. Mbr.[1]; Rosiclare Ss. Mbr.; Fredonia Ls. Mbr.	*Blue River Group*: *Bryantsville Breccia Bed*; Ste. Genevieve Limestone	Aux Vases Ss. and Rosiclare Ss. Mbr. of Ste. Genevieve Ls.; Ste. Genevieve Ls.: *Joppa Ls. Mbr.*; *Karnak Ls. Mbr.*; *Spar Mtn. Ss. Mbr.*; Fredonia Ls. Mbr.
Interval C (part)						

[1] Includes *Bryantsville Breccia Bed*.

basin region

ILLINOIS		MISSOURI	Series	Stage
Southwestern and Southern		Southeastern		
Caseyville, Abbott, and Spoon Fms.-Pennsylvanian Tuscaloosa Fm.-Cret.		Quaternary undivided		
Grove Church Shale			Chester	*Elviran*
Kinkaid Ls.				
Degonia Ss.				
Clore Ls.				
Palestine Ss.				
Menard Ls.				
Waltersburg Ss.				
Vienna Ls.		*Vienna Ls.*		
Tar Springs Ss.		*Tar Springs Fm.*		
Okaw Formation	Glen Dean Ls.	*Glen Dean Ls.*		*Hombergian*
	Hardinsburg Ss.	*Hardinsburg Ss.*		
	Haney Ls.	*Golconda Formation*		
	Fraileys Shale			
	Beech Creek Ls.			*Gasperian*
Cypress Ss.		Cypress Fm.		
Paint Creek Fm.	Ridenhower Shale	*Paint Creek Formation*		
	Bethel Ss.			
	Downeys Bluff Ls.			
Yankeetown Chert		*Yankeetown Chert*		
Shetlerville Ls. Mbr.		Renault Fm.		
Levias Ls. Mbr. / Aux Vases Ss.		Aux Vases Ss.	Meramec (part)	
Ste. Genevieve Ls.	*Joppa Ls. Mbr.*	Ste. Genevieve Limestone		
	Karnak Ls. Mbr.			
	Spar Mtn. Ss. Mbr.			
	Fredonia Ls. Mbr.			

terior basin rocks are uncertain; the Batesville is believed to be an age equivalent of lower Chester rocks, and the Fayetteville, an equivalent of the *Haney*, Hardinsburg, and Glen Dean (Spreng, 1961, p. 78).

UPPER BOUNDARY OF INTERVAL D

Coarse- to fine-grained detrital strata of Pennsylvanian or Cretaceous ages unconformably overlie rocks of interval D in the Eastern Interior basin. Pre-Cretaceous erosion has removed more than 1,300 feet of interval D strata as well as older rocks in southernmost Illinois. Basal Pennsylvanian units overlying rocks of interval D at different places represent Morrow, Atoka, or Des Moines strata, and they are dominantly sandstone and conglomerate. Despite the widespread major erosional hiatus that preceded deposition of these Pennsylvanian sediments, the systemic boundary is difficult to recognize in many places. The pre-Pennsylvanian surface has considerable local relief, commonly 100–300 feet and in some areas as much as 450 feet (Siever, 1951, p. 561–562); thus, basal Pennsylvanian strata locally rest on a number of different interval D units within short distances. Pennsylvanian basal sandstones are compositionally and texturally similar to those of interval D, and in many places unfossiliferous mudstone and siltstone of uncertain age constitute the strata between known Mississippian and Pennsylvanian beds; therefore, recognition of the systemic boundary in some areas is difficult. In areas of high control-point density and detailed data, such as many parts of Illinois, the boundary is well defined; in other areas, such as parts of eastern Kentucky, fewer control points and less accurate data may produce errors as much as 100 feet or more in interval D thicknesses given here (pl. 6-A).

Pre- or intra-Pennsylvanian erosion manifested itself in three ways: by irregular but progressively deeper beveling of pre-Pennsylvanian strata toward the northern margins of the basin, by beveling of smaller pre-Pennsylvanian structures within the basin, and by the imposition of an integrated southwest-trending channel system which is well expressed in the northeastern and southern parts of the basin. More than 600 feet of interval D strata was beveled in central and eastern Illinois and western Indiana. Channel cutting produced local relief commonly as much as 300 feet.

In much of eastern Kentucky, sandstone, conglomerate, and shale of the Lee and Breathitt Formations of

Morrow and Atoka ages unconformably overlie rocks of interval D, and, locally, in northeastern Kentucky they truncate rocks of intervals C and B. In southeastern Kentucky, however, the Pennington and Lee Formations are conformable according to Englund (1964), who considered the basal part of the Lee to be of Late Mississippian age.

In southwestern Missouri, Pennsylvanian rocks of the Hale Formation (Morrow) overlie the Fayetteville Formation at a few scattered localities. The absence of interval D in parts of southeastern Missouri where clays of the *Cheltenham Formation* (Atoka) overlie interval C rocks is probably the result of pre-*Cheltenham* erosion.

THICKNESS AND LITHOFACIES TRENDS

All originally deposited thicknesses of interval D rocks in the Eastern Interior basin have been modified by Pennsylvanian, Cretaceous, or Holocene erosion. The youngest Mississippian rocks, which include the Kinkaid Limestone and *Grove Church Shale*, occur in the southern part of the basin. There, more than 1,350 feet of interval D rocks (pl. 6-A) are recorded in extreme southern Illinois, and nearly 1,200 feet, in western Kentucky. The original upper surface of the interval cannot be confidently restored in most areas; however, within units that are widespread, the Glen Dean Limestone and older rocks thin depositionally northward and eastward toward basin margins (pl. 9-D, secs. $a-a'$, $d-d'$). Because the proportion of sandstone in interval D increases northeastward, detrital units now eroded away may once have been appreciably thicker in that direction.

In the eastern Kentucky outcrop belt marginal to the Cincinnati arch, preserved thicknesses of interval D do not exceed 500 feet. Westward thinning toward the arch and the thinness or absence of interval D in northeastern Kentucky largely result from deeper pre-Pennsylvanian erosion in those areas. The lower, dominantly carbonate, part of interval D is also thinner here than in either western Kentucky or the Appalachian basin.

The lithofacies patterns of interval D (pl. 6-B) result from combining many rhythmic alternations of detrital and carbonate units, modified by irregular amounts of erosion in the upper part of the interval. These areas are distributed roughly symmetrically along a line from west-central Indiana to southern Illinois. In the more nearly complete sections of interval D in the southern part of the basin, sandstone areas are most abundant in southern and southeastern Illinois and are common in western Kentucky. In areas of incomplete sections in the northeast part of the basin, lobate sandstone tongues converge toward the center of the basin, a pattern similar to that shown by Swann and Bell (1958). Increase of carbonate rocks along the southernmost erosional limits in Kentucky results from the southward thinning of pre-Big Clifty detrital units (pl. 9-D, sec. $a-a'$). A westward increase in mudstone in Chester rocks reported by Swann (1963, p. 15; 1964, p. 649) is not evident on the lithofacies map but is evident for some detrital- and carbonate-dominated formations in cross section $a-a'$ (pl. 9-D). A known southeastward increase in the mudstone–sandstone ratio in post-Glen Dean rocks of west-central Kentucky is not indicated by the lithofacies pattern.

In eastern Kentucky, lithofacies patterns marginal to the Cincinnati arch largely reflect the degree of pre-Pennsylvanian erosion. Areas in which fine detritals are shown are areas in which the Pennington is thick. Carbonate patterns in east-central and northeastern Kentucky indicate areas in which Pennington is thin or absent, perhaps because it was thin originally, but mostly because it was partly or entirely removed by pre-Pennsylvanian erosion.

SOURCE AREAS AND ENVIRONMENT OF DEPOSITION

An ancient southward-flowing river system, first envisioned by Stuart Weller (1927, p. 26) and later named the Michigan River (Swann, 1963, p. 12), transported detrital sediments into and across the basin (pl. 11, fig. 4; pl. 12, fig. 4). Evidence for the Michigan River consists of regional thickness and facies distribution of interval D units (Swann and Bell, 1958), southward-dipping crossbeds in generally southward-trending sandstone bodies (Potter and others, 1958; Potter, 1963), and channel-fill morphology of sandstone bodies (Potter, 1962, p. 28–29; 1963; Reynolds and Vincent, 1967). The main sediment source was in eastern Canada, either in the Shield or in northward extensions of the Appalachian Mountain belt. The possibility that the uplifted Franklinian geosyncline in northern Canada was a major source has also been suggested (Swann, 1964). According to Swann, the Michigan River flowed southwestward across Michigan and northern Indiana into a shallow sea, where detrital sediments accumulated as a birds-foot delta projecting southward beyond a N. 65° W.-trending shoreline. Lateral shifts to the northwest or southeast in the course of the Michigan River of as much as 200 miles produced belts of sand and mud in different parts of the region at different times. Major northeast and southwest oscillations of the shoreline, perhaps as much as 600–1,000 miles, resulted in numerous marine transgressions during which carbonates were deposited in the basin and surrounding areas. The resulting highly variable complex of sediments includes shallow-marine detrital and carbonate facies, littoral detrital facies, and continental detrital and coaly facies. Superimposed on the system of shifting shorelines and

positions of the river system, Swann (1963, p. 14–15) postulated a northwest-flowing sea current or drift (pl. 11, fig. 4) which carried muds northwestward, leaving relatively clear water in the southeastern part of the basin where limestone accumulated. Swann (1964, p. 652–653) estimated that sea depths in which interval D sediments were deposited were 50 to 75 feet, having a maximum and minimum range of 30 to 100 feet.

A low dip of the paleoslope, low relief of the sea floor, and shallow-marine environment are indicated by the widespread distribution of tabular carbonate units (which individually indicate similar depositional environments over very large areas), by the presence of oolitic limestone and current-bedding features, and by numerous shallow-water fossils. The spreading of some sheetlike bodies of terrigenous clastics also shows widespread and uniform shallow-marine and shoreline environments, in contrast to environments of the interval B Borden deltaic sequence which encroached into deep water and was onlapped by deepwater carbonates deposited along its front.

It is not clear whether a northwestern source area (pl. 11, fig. 4; pl. 12, fig. 4) also contributed detrital sediments to the basin, but some sandstone (Yankeetown Chert and Degonia Sandstone) is prominent in western parts of the basin, and southeastward-dipping crossbeds in limited areas there (Potter and others, 1958, p. 1038) suggest southeastward transport of the sand. Different mineral associations and crossbedding directions of basal Pennsylvanian sandstones in western Illinois and those farther east strongly suggest northwestern source areas in the Transcontinental arch region (Potter and Siever, 1956, p. 329), but petrologic studies of Mississippian sandstones are too limited to identify different source areas. Restorations of interval D paleogeography by Swann (1963, p. 14, fig. 4) and Potter (1962, p. 34, fig. 14) stress the dominance of the Michigan River system in delivering sediment to the basin.

Ozark region source areas may have contributed small amounts of interval D detrital sediments to the Eastern Interior basin. Shoal areas or lowlands in the Ozark region and the Cincinnati arch area controlled the orientation of the river system (Potter and others, 1958, p. 1038; Swann, 1963, p. 15) and limited the dispersal of detrital materials from the Michigan River. The central Ozark region may have been the source for sandstones of the Batesville Formation of southwestern Missouri and northern Arkansas.

Eastward transportation of terrigenous clastics from Michigan River sources across the Cumberland saddle during middle interval D time might be indicated by the distribution of the Big Clifty west of the saddle and its presumed equivalent, the Hartselle Sandstone (or Big Clifty of Vail, 1959, fig. 13) east of the saddle (pl. 9-D, sec. *a*–*a*′; pl. 11, fig. 4). Sandstone in this unit, presumably transported southward by the Michigan River system, is predominant along the eastern margins of the Eastern Interior basin, whereas mudstone is predominant to the west. Crossbedding studies in west-central Kentucky, however, indicate westward and northwestward current directions (Potter and others, 1958, p. 1028) for sandstones in the Big Clifty. A sedimentary environment map (pl. 11, fig. 4) for the period of deposition of the Big Clifty purports to show eastward dispersal of sands through the Cumberland saddle area; sands were winnowed and in part retransported short distances westward by the dominant westerly sea current or "drift" postulated by Swann (1963) or by tidal currents. According to this interpretation, the Big Clifty contains current features aligned with its depositional strike in west-central Kentucky, a possibility recognized by Potter and others (1958, p. 1038).

Whether detrital material from Appalachian basin source areas contributed to interval D strata in the Eastern Interior basin has not been determined, but some contribution is suggested in late interval D time by lithologic similarity of the Pennington Formation of eastern Kentucky and the Leitchfield and Buffalo Wallow Formations of west-central Kentucky. Intervals D and C units below these formations were deposited across the Cumberland saddle area as shown by their gross lithic and faunal similarities, and their thicknesses indicate that the saddle was a negative trend across the Cincinnati arch. In eastern Kentucky, sandstones in the Pennington, which thin westward and southward, were transported southwestward along the Appalachian basin from a northeastern source area (Mitchum, 1954; Vail, 1959, p. 59). The northwest sea-current drift proposed by Swann, which moved muds discharged by the Michigan River west, explains the mudstone in the western part of the Eastern Interior basin but not the mudstone in the Leitchfield and Buffalo Wallow Formations in the eastern part of the basin. A sluiceway across the Cumberland saddle area (Cumberland Strait), however, could provide a passage for late interval D fine-grained sediments carried by westward-flowing marine currents from Appalachian sources into the Eastern Interior basin (fig. 15). Mixing of Appalachian sediments with those from the Michigan River would thus have taken place in west-central Kentucky.

Abundant marine faunas and oolitic limestones indicate subtropical or tropical conditions during interval D. Scattered coaly beds and fairly abundant fossil plant remains in sandstones and siltstones suggest humid conditions. Red shales are common in many detrital units; they may indicate the presence of deep residual soils in source areas (Vail, 1959, p. 52).

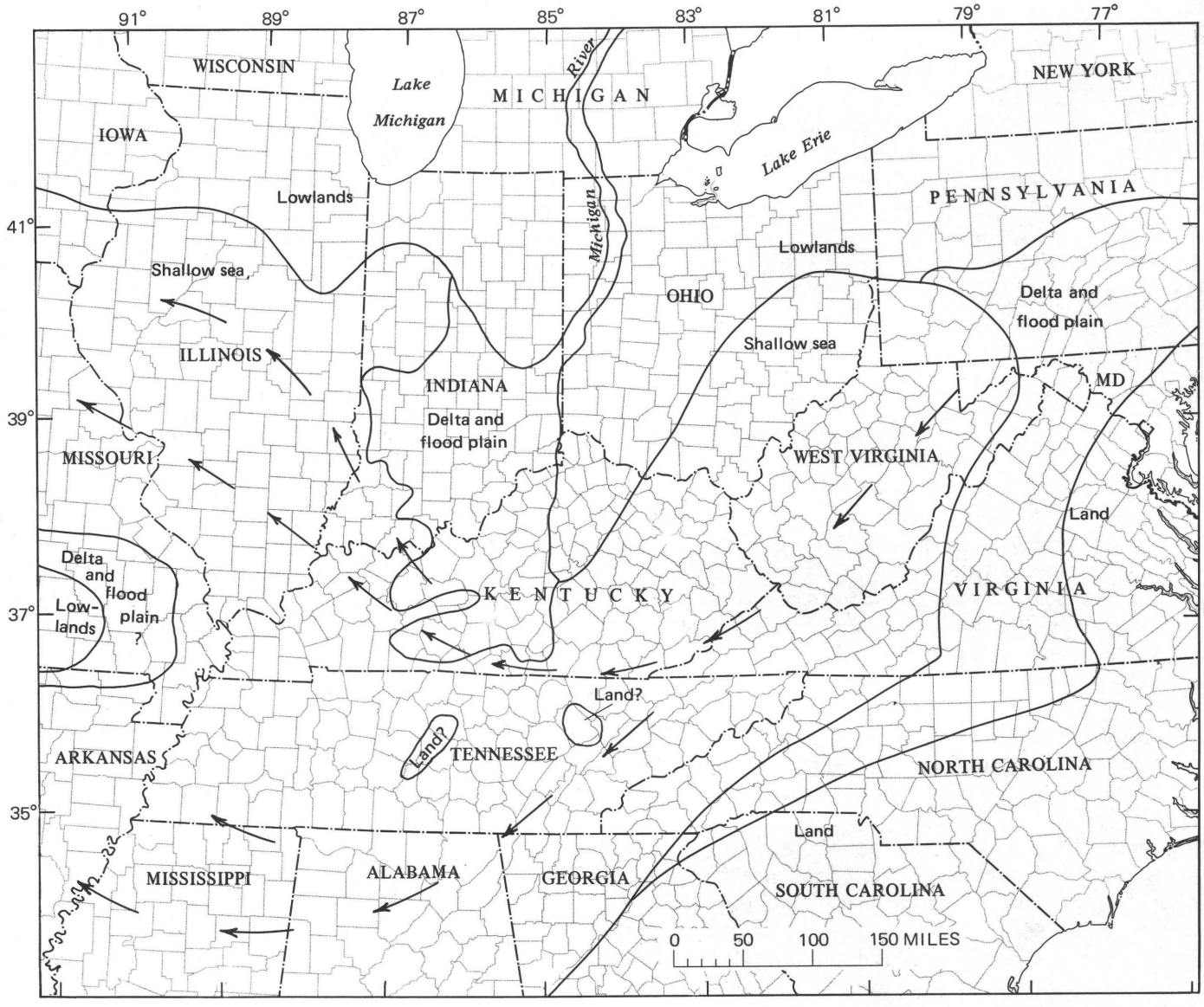


FIGURE 15.—Generalized sedimentary environments in eastern midcontinent region during mid-interval D (Big Clifty) time. Arrows show directions of hypothetical sea currents.

Causes for cyclic sedimentation in the Chester are unresolved. Swann (1964, p. 656–657) considered that climatic fluctuations, primarily changes in rainfall in the source region of the Michigan River combined with even basin subsidence, were the chief factors. He (Swann, 1964, p. 654) also cited sea-level changes as possible factors in some cycles in the upper part of interval D and suggested that alternation between pluvial and interpluvial stages corresponded to fluctuations of the margins of continental glaciers in the southern hemisphere.

The paleogeographic maps (pl. 12, figs. 4*A*, *B*, *C*) show in general the major stream and shoreline positions that existed, with local variations and oscillations, through much of interval D time. The Michigan River system, which had a Canadian source, was the dominant transporting agent. A delta complex at its mouth included both sheet sands and linear sand bodies, such as channel fillings, point bars, and distributary mouth bars. Sediment discharged by the Michigan River is believed to have been deposited as far south as northwestern Alabama and northeastern Mississippi and intermittently at least as far southeast as east-central Tennessee and northeastern Alabama. Coaly beds formed in interdistributary swamps on the delta plain (Swann, 1964, p. 656). Smaller river systems that headed in hilly land in the eastern part of the Transcontinental arch and possibly in land in the central

Ozark region transported detritus southward into the Eastern Interior basin and into the Arkoma basin, respectively. Highlands east of the Appalachian basin contributed detrital sediments, some of which late in interval D time were probably carried through the Cumberland Strait and intermingled with Michigan River sediments in western Kentucky. The northern part of the Cincinnati arch was a low-lying peninsula, or shoal area, which contributed little sediment but which acted as a barrier between the Eastern Interior and Appalachian basins.

Shallow seaways in which the water freely circulated connected the Eastern Interior basin with other basins to the east, south, and west. Widespread emergence and southward tilting across northern parts of the region terminated interval D deposition and produced a low-lying landmass in most or all of the Eastern Interior region.

PALEOTECTONIC IMPLICATIONS

The closely spaced reconstructed isopachs in southeastern Missouri and southern Illinois (pl. 10, fig 4) are based on the supposition that a hingeline existed in that region, forming a rather abrupt southeast-trending boundary between the Eastern Interior basin and the adjoining Ozark region. Proper reconstruction across Missouri depends on whether any of the numerous detrital units of the Eastern Interior basin were originally continuous with the Batesville Sandstone of northern Arkansas and the Batesville Formation of southwestern Missouri. The southward disappearance of the Batesville in northern Arkansas (E. E. Glick, oral commun., 1968) suggests a southerly transportation direction for its coarse components, but whether an Ozark source area or Eastern Interior basin sources were prime contributors is not known. Sedimentary trends used for restoring the isopachs for the interval suggest that areas in the Ozark region were sources for Batesville sediments. However, if the Batesville was continuous with an Eastern Interior basin interval D sandstone unit, a continuous southwestward-dipping paleoslope from the Eastern Interior basin across or around the east side of the Ozark region would be required.

Epeirogenic uplift affected the Transcontinental arch region and the previously stable shelves marginal to it. The Cincinnati arch was mildly uplifted, was generally emergent, and was flanked by an alternately submergent and emergent shelf along its eastern margin in eastern Kentucky. Uplift in the central and western Ozark region produced a platform which may have been structurally connected with the Transcontinental arch across northwestern Missouri and western Iowa, although this connection is not shown on the map. Much of Missouri and Iowa was a stable submerged platform.

Mild subsidence characterized broad areas of the basin region, and sediment thicknesses indicate that the greatest subsidence was in the southern part of the Eastern Interior basin. The lateral persistence of individual formations and their small thickness variations show that the basin subsidence was relatively even. Minor differential movements occurred along the La Salle anticlinal belt, along the folds in Illinois and Indiana (Siever, 1951, p. 569), and along the Rough Creek fault zone. A regional southwestward-dipping paleoslope and a mildly negative trough across Michigan controlled the trend of the Michigan River system. The area of the Cumberland strait was a mildly downwarped saddle between the Jessamine and Nashville domes. The saddle connected eastern and western basins.

Similarities of interval D detrital rocks in the Black Warrior basin of northwestern Alabama and northeastern Mississippi to those in the Eastern Interior basin suggest interbasin connections according to Swann (1964, p. 653). In the intervening areas, across western Tennessee, interval D rocks, if originally present, have been removed by erosion. A mildly negative trough across this area is shown on the interval D palotectonic map (pl. 10, fig. 4); it is restricted on the west by the Ozark shelf and on the east by the Nashville dome. The reconstruction takes into account the inferred presence of the eastern Ozark uplift as a barrier to westward clastic dispersal, eastward thinning of the interval D sequence, and the similarities between cyclically deposited rocks of the Black Warrior basin and those of the Eastern Interior basin. The trough is inferred to have been a locus of deposition of detritals from the Michigan River during times of maximum outpourings. Interval D carbonate units are not abundant in the northern part of the Black Warrior basin, which fact may indicate a trough too deep or waters too turbid for much carbonate deposition.

Post-interval D differential movements prior to Pennsylvanian deposition include subsidence of the basin in southern Illinois, southwestern Indiana, and western Kentucky relative to three major positive structures: the La Salle anticlinal belt, the Du Quoin monocline, and the east flank of the Ozark uplift (fig. 10) (Wanless, 1955, p. 1769–1772). Restored isopachs of interval D rocks, although less reliable than isopachs of the Lower Pennsylvanian, may suggest that those structures, except for the Du Quoin monocline, were also active during interval D deposition (pl. 10, fig. 4).

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS AND PALEOTECTONIC IMPLICATIONS

Factors affecting the total thickness of the Mississippian System include post-Mississippian erosion, variations in Mississippian sediment distribution, and Mississippian tectonic movements. Erosional effects include regional beveling and channel cutting after deposition of interval D but prior to deposition of Pennsylvanian strata, and post-Pennsylvanian, Mesozoic and Cenozoic erosion which has stripped Mississippian rocks from the Pascola arch (fig. 10) and other major positive elements. Thickness variations of Mississippian deltaic deposits in parts of the region, shifts in loci of detrital deposition, and filling of troughs peripheral to deltaic platforms by carbonate and detrital sediments appear to be less important than erosion in determining present thicknesses.

Total thicknesses of preserved Mississippian rocks (pl. 7) indicate some features which are also shown on the interval isopach maps: the general outline and depocenters of the Eastern Interior basin; areas of thicker sedimentary rocks at the sites of the Fairfield basin and Moorman syncline; areas of thin Mississippian rocks along the La Salle anticlinal belt; thin or absent Mississippian rocks on large positive elements, such as the Cincinnati and Kankakee arches and the Ozark uplift; and deposits of uniform thickness on a shelf area lying between the Cincinnati arch and the Appalachian basin to the east and on platform or shelf areas and margins of outlying basins in Missouri.

From their erosional edges along the Wisconsin, Kankakee, and Cincinnati arches, Mississippian rocks thicken to maximums of about 3,300 feet in the southern part of the Fairfield basin and about 2,800 feet in the Moorman syncline. Isopachs indicate an elongate northwest trend for the Eastern Interior basin. The fairly regular basinward thickening is locally interrupted by irregularities and crowding of isopachs along the La Salle anticlinal belt, the result both of depositional thinning or absence of some Mississippian units and of pre-Pennsylvanian and Pennsylvanian erosion. Mississippian rocks locally are thin or absent along the fault zones in southern Illinois, western Kentucky, and Missouri. Although mostly the result of post-Pennsylvanian erosion, some of this thinness or absence is due to internal thinning within the Mississippian rocks. The consistent offset and bending of the isopachs along the Rough Creek fault zone in western Kentucky suggest tectonic instability during Mississippian time. Although the Rough Creek and other fault zones are shown as post-Mississippian on the interval thickness maps (pls. 3-A, 4-A, 5-A, 6-A), movement may have occurred along them during the Early Mississippian, and quite likely during interval I time. Irregular thickness patterns and absence o Mississippian rocks along anticlinal or domal struc tures, such as the Lincoln anticline and Hicks dome, ar largely the result of post-Mississippian erosion. South westward-thickening sequences in the Illinois basi toward the Ozark uplift and Pascola arch have bee abruptly truncated near those positive features by pre Late Cretaceous and more recent erosion. Mississippia interval D rocks are not preserved in west-central Ten nessee and adjoining southern Kentucky, and apparen thinning toward these areas is at least partly the resul of post-Mississippian erosion.

East of the Cincinnati arch, Mississippian rocks i eastern Kentucky generally thicken southeastwar toward the Appalachian basin, but the region covere by this chapter includes only marginal areas of tha basin, and thicknesses generally are no more tha about 700 feet. Rocks having somewhat greate thicknesses (as much as 900 feet) are present in th Cumberland saddle area — an area of subsidence dur ing parts of middle and Late Mississippian time — and locally, in northeastern Kentucky where greate thicknesses are due to relatively thick Lower Mississip pian detrital rocks. Except for southeastern Kentucky pre-Pennsylvanian erosion has removed differing amounts of Mississippian rocks in the eastern part o the State.

Relatively thin Mississippian rocks, generally les than 500 feet, characterize Missouri. Northwest trend of isopachs in northern Missouri in part reflect thick ness trends within the lower part of Mississippian se quence. Rocks of interval D and the upper part of inter val C are absent in most of Missouri because of Penn sylvanian to Holocene erosion and, in part, probabl because of nondeposition. The apparent thinning ont the Ozark uplift is largely the result of erosion, but i also may indicate original depositional thinning of th Mississippian sequence toward this positive element.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Detrital rocks of Pennsylvanian or Cretaceous ag unconformably overlie rocks of all Mississippian inter vals in parts of the Eastern Interior basin. (See pl. 8. Pleistocene deposits of glacial origin or association, no shown on plate 8, overlie Mississippian strata in part of Indiana, Illinois, and northern Missouri.

Pennsylvanian rocks directly overlying Mississip pian rocks are of Morrow, Atoka, Des Moines, and locally, Missouri ages. Basal Pennsylvanian strata ap pear to be largely nonmarine and include sandstone conglomerate, mudstone, and clay. Morrow strata ar

represented in the Lee Formation in eastern Kentucky (pl. 15), the Caseyville Formation in western Kentucky and southern and southeastern Illinois, the Mansfield Formation in southern and western Indiana, and isolated outliers of the Hale Formation in southwestern Missouri. Atoka rocks include most of the Abbott Formation in Illinois, the Brazil Formation in Indiana, the *Cheltenham Formation* in central and eastern Missouri, the *McLouth* and *Riverton Formations* in northwestern Missouri, and sink-fillings of *Burgner Formation* rocks in southwestern Missouri. Des Moines strata overlying Mississippian rocks are represented by the Spoon Formation in Illinois and the *Krebs* and *Cabaniss Subgroups* of the *Cherokee Group* in Missouri. Channel-fill sandstone in the *Pleasanton Group* of Missouri age locally overlies Mississippian rocks in northeastern Missouri. Atoka and Des Moines strata are more widespread than Morrow rocks; in Illinois, the Abbott generally overlaps the Caseyville northward and in turn is overlapped by the Spoon Formation on some anticlinal structures (Kosanke and others, 1960, p. 30). In Missouri, the *Cherokee Group* is widely distributed and appears to overlap the *McLouth* and *Cheltenham Formations.*

A major erosional hiatus between Mississippian and Pennsylvanian deposition is well documented in much of the region and is marked by progressively deeper beveling of pre-Pennsylvanian strata toward the northern, eastern, and northwestern margins of the Eastern Interior basin. As much as 1,500 feet of Mississippian strata are estimated to have been removed in northern Illinois and Indiana. In many areas, channels cut into Mississippian rocks on the basal Pennsylvanian erosional surface produced local relief of 100–300 feet commonly and in parts of southeastern Illinois, as much as 450 feet (Siever, 1951, p. 561–562).

Studies of morphology, internal features, and petrology of Pennsylvanian channel sandstones (Siever, 1951; Wanless, 1955; Potter and Siever, 1956; Siever and Potter, 1956; Potter and Glass, 1958; Potter, 1963; Potter and Desborough, 1965) indicate that several integrated channel systems are incised into Mississippian rocks. In channel systems, the sediment transport was generally southwest to west in western Kentucky, Indiana, and southern and eastern Illinois. In western and west-central Illinois, transport directions appear to have been southeast (Potter and Siever, 1956, p. 242). Major source areas were the southeastern Canadian Shield, highlands east of the northern and middle Appalachian area, and the south flank of the Transcontinental arch (Potter and Siever, 1956, p. 243). Dominant flow patterns down a southwest-dipping paleoslope are similar to those shown by interval D sandstones.

In the eastern Kentucky outcrop belt, basal Pennsylvanian sandstone, conglomerate, mudstone, and clay of the Lee and Breathitt Formations overlie rocks of Mississippian intervals D, C, and B. Beveling and (or) channel cutting mark an unconformity below Pennsylvanian rocks along the margin of the Cincinnati arch. In south-central Kentucky, Pennsylvanian rocks rest on interval D strata; in northeastern Kentucky, intervals D and C rocks are thin or locally absent, and in parts of the area basal Pennsylvanian sandstones lie on interval B beds. Local absence of intervals D and C rocks in northeastern Kentucky, however, probably indicate intra-Mississippian erosion. In extreme southeastern Kentucky, the Pennington Formation (upper interval D) is laterally gradational and intertongues with the Lee Formation, which in its basal part is of Late Mississippian age (Englund, 1964). Apparently, marked pre-Pennsylvanian erosion may have been restricted in eastern Kentucky to areas along the Cincinnati arch, and erosional effects died out southeastward in the subsurface.

In much of northern and western Missouri, Atoka and Des Moines rocks overlie strata of Mississippian intervals B and C. Clay and other detrital rocks of the *Cheltenham Formation,* part of which may represent the residuum of preexisting rocks, overlie rocks as old as Ordovician in parts of the Ozark region. The distribution of the *McLouth Formation* and the *Krebs* and *Cabaniss Subgroups* of the Cherokee Group suggest successive eastward overlap of these units across the Forest City basin toward the Mississippi Valley. The absence of Mississippian interval D rocks in these areas probably is partly the result of pre-Atokan–post-Mississippian erosion. Interval D rocks, absent in east-central and the adjoining northern part of southeastern Missouri, probably were also removed prior to deposition of the *Cheltenham Formation* and *Cabaniss Subgroup.*

Detrital rocks of the Tuscaloosa and McNairy Formations of Late Cretaceous (Gulfian) age lie on an erosional surface which truncated Mississippian and older rocks in the Mississippi embayment area of southern Illinois, westernmost Kentucky, and western Tennessee (pl. 15). More than 3,000 feet of Mississippian strata was removed by pre-Cretaceous erosion in southern Illinois. Commonly, a well-defined weathered zone separates these Cretaceous beds from Mississippian and older rocks. In western Illinois, the *Baylis Formation* (Upper Cretaceous) overlies Mississippian and Pennsylvanian strata in a small area.

PALEOTECTONIC IMPLICATIONS

The distribution, contact relationships, and age of basal Pennsylvanian units in the region indicates emergence or near-emergence of most or all of the region after deposition of interval D sediments and before deposition of Pennsylvanian strata. Crustal in-

stability is indicated by unconformable relationships of basal Pennsylvanian rocks to beveled edges of older units along the margins of positive elements, by Pennsylvanian overlap relationships, by channel cutting in many areas, and by truncation of older units and thinning of Pennsylvanian strata on anticlinal structures.

Maximum relative uplift in the region at the close of Mississippian time was along the northern margins of the Eastern Interior basin and indicates widespread emergence there and southward tilting of the basin. Differential movements shown by variations in thickness of the basal Pennsylvanian units (Wanless, 1955, p. 1769–1772) include downwarp of the basin in southern Illinois and adjacent Indiana and Kentucky and growth of three positive structures, the La Salle anticlinal belt, the Du Quoin monocline, and the east flank of the Ozark uplift (fig. 11). Reconstructed thickness of Mississippian interval D rocks suggests that these structures, except for the Du Quoin monocline, also were active during interval D and earlier times. Detailed evidence for Mississippian movement along some of these trends has been cited (Whiting, 1959, p. 13–19; Brownfield, 1954, p. 20–23).

Channels filled by sandstone of Pennsylvanian age were incised in interval D rocks in southern Indiana and in intervals C and D strata in west-central Kentucky and along the outcrop belt in eastern Kentucky. If the channels represent subaerial erosion, the Cincinnati arch was emergent during the Mississippian-Pennsylvanian erosional episode.

Growth of the Pascola arch as a major positive feature between Mississippian and Late Cretaceous times is indicated by the pre-Cretaceous erosional edge in the Mississippi embayment area. Thickness patterns of Mississippian intervals B, C, and D show evidence of an actively subsiding trough in the embayment area which extended north into the Fairfield basin of Illinois, but subsequent uplift and erosion has destroyed evidence of its former southern extent. Fault systems in Illinois and western Kentucky appear to have moved mostly after Pennsylvanian time; data supporting Pennsylvanian movements seem to be lacking. Some of these faults, however, such as the Rough Creek fault zone in western Kentucky, probably were active during the Mississippian Period.

West of the Eastern Interior basin, maximum growth of the Lincoln anticline probably took place during the interval between the end of Mississippian and beginning of Pennsylvanian deposition (McQueen and others, 1941; Koenig, 1961, p. 76). In northern and western Missouri, the pattern of sub-Pennsylvanian units (Mississippian intervals B and C) suggests a system of northwest-striking and northwest-plunging folds which have been beveled to progressively deeper horizons toward the Ozark uplift. The Ozark uplift was an emergent positive element prior to Pennsylvanian deposition.

CONCLUSIONS AND SPECULATIONS

TECTONIC CONTROLS

Northeast- and northwest-trending structures constitute the present dominant structural grain in the region (fig. 11). Several less persistent linear trends, including anticlinal belts and fault zones, lie at angles to the major elements. Structural relief largely reflects post-Mississippian movements, but in a subdued to marked manner the major structures are also reflected in patterns of Late Devonian and Mississippian sedimentation.

The Michigan basin, Fairfield basin, and Mississippi embayment are negative bedrock elements which, when alined, are subparallel to the Appalachian Mountains front. During Late Devonian and much of Mississippian time a similar spatial relationship existed, and northeast-trending negative features in the eastern interior region paralleled the Appalachian basin. A southwest-trending trough provided interconnections between the Michigan and Eastern Interior basins during intervals B and D times and extended into the area of the present Mississippi Embayment. Sedimentary thicknesses and inferred deepwater environments indicate that subsidence along this trend far exceeded relative uplift on flanks of the trough. A subparallel linear trend involving the Forest City basin and a small basin in central Iowa also may have existed during Mississippian time.

Positive elements, such as the Cincinnati arch and the Ozark uplift, apparently were relatively passive elements among the actively subsiding troughs. Amplitudes of major troughs, as determined from thicknesses of strata, decrease westward, which fact suggests formation of troughs by westward-directed tangential stresses. In such a stress field, however, amplitudes of arches should be comparable to those of troughs. Another explanation is that tensile or vertical-couple stresses, perhaps the result of transfer of subcrustal material, resulted in grabenlike or downdrawn troughs.

Some Mississippian depositional patterns appear to be offset along some supposedly subordinate linear features which crosscut or lie at angles to the dominant structural grain of the region. These features include the Cap au Gres fault and associated Lincoln-Dupo anticline trend, the Rough Creek fault zone, and the La Salle anticlinal belt, all which may have been active during Late Devonian and Mississippian time. Although displacements along the fault trends seem to have been largely vertical, left-lateral movements are

also suggested by the apparent displacement of the Lincoln and Dupo anticlines along the Cap au Gres fault (Cole, 1961), and by the apparent displacement of the Fairfield basin and Moorman syncline on opposite sides of the Rough Creek fault zone (fig. 10). Southern limits of the Louisiana Limestone (Upper Devonian) lie along the trend of the Cap au Gres fault, and the lithofacies pattern of interval A (pl. 3-B) also seems to be influenced by this trend. Isopach trends of the total Mississippian System (pl. 7) and interval D (pl. 6-A) seem to be offset by the Rough Creek fault zone. Lateral fault movements may have been more important than generally recognized in the Paleozoic structural framework of the region.

Upper Devonian, Mississippian interval A (pl. 3-A), and interval B (pl. 4-A) rocks thicken irregularly along and east of the present La Salle anticlinal belt trend in eastern Illinois, Indiana, and west-central Kentucky. The irregularities indicate structural instability along the La Salle trend, which appears to be a southeastern continuation of the Wisconsin arch. Structural instability in northwest-trending zones marginal to or within the Ozark region is also suggested by development of the Upper Devonian Wittenberg trough (fig. 13) and distribution of the Sulphur Springs Formation south of the Cap au Gres fault (fig. 10).

SEDIMENTATION AND SUBSIDENCE

Sedimentary features in most Mississippian rocks in this region seem to indicate neritic environments; some features suggest very shallow to shallow neritic (0–120 feet) and deep neritic (120–600 feet) divisions, similar to the infraneritic and epineritic classification of Scott (1940).

Features which suggest relative water depths for Mississippian marine deposits in the region are:

1. Inclination and extent of interval B (Borden Formation) deltaic foreset slope beds, which indicate shallow to deep neritic environment;
2. Lithology, morphology, internal features, and biota of limestone beds in intevals B, C, and D strata (Salem, Ste. Genevieve, and Burlington Limestones), such as crossbedded oolitic and fossil-fragmental units, which indicate high current or wave energy and indigenous life forms of a very shallow neritic environment;
3. Dark, fine-grained aspect of carbonate and siliceous rocks in intervals B and C (Fort Payne Formation, Salem and St. Louis Limestones in southern parts of Eastern Interior basin), which suggests low energy and reducing conditions, probably indicating a fairly deep-neritic environment;
4. Presence of the evaporite facies in interval C, which, considering their distribution relative to the tectonic elements, are indicative of very shallow marine waters;
5. Shallow sandstone-filled channels, crossbedding, and marine invertebrate fauna of many interval D sandstone units, which denote very shallow neritic to littoral environments;
6. Abundant phosphatic constituents in interval A (Maury Formation), which are interpreted to indicate moderately deep neritic environment.

In addition, interval D coaly units, possibly associated red beds, and sandstone with fossil tree trunks in growth position are all interpreted to have been deposited in supratidal or continental environments at altitudes close to sea level.

Thickness of strata deposited, coupled with estimated depth of water, gives an approximation of the relative degree of subsidence to sediment accumulation during Late Devonian and most of Mississippian time. Maximum subsidence occurred in the southern part of the Eastern Interior basin, in southern Illinois and western Kentucky. At times, as in the Late Devonian, filling generally kept pace with subsidence. By interval A time, subsidence had almost ceased in the basinal area, and a fairly even thickness of interval A sediment accumulated under water on a topography of little relief. Temporary stability in the Eastern Interior basin and gradual southward deepening of water are inferred from the thickness and abundance of phosphatic constituents in the Maury Formation. A deep trough across central and southern Illinois developed during early interval B time and connected with the Michigan basin; a widespread stable shelf characterized by very shallow water existed to the west (Burlington time). Initial subsidence of the trough far exceeded sedimentation rates in Illinois; but in mid-interval B time, sedimentation rates increased, and the trough was partially filled by deltaic deposits of the Borden. Marginal to the delta, a deep neritic environment receiving very little sediment persisted in the southern part of the trough and adjacent areas to the south and east. Subsidence of the trough lessened in late interval B through mid-interval C time and deepwater prodelta areas were slowly filled by siliceous and carbonate rocks. Dark-hued fine-grained thick sediments (Salem and early St Louis times) deposited during early and mid-interval C time possibly reflect moderate to deep neritic depth and indicate that slow differential downsinking of the southern part of the trough continued into late interval C, at a time when detrital supply was low. A close balance between sedimentation and subsidence over large areas of the Eastern Interior basin in late interval C time and in interval D time is demonstrated by the even thicknesses of individual units (upper St. Louis, Ste. Genevieve, and Chester strata), all deposited in very

shallow neritic to continental environments. Large volumes of interval D detrital rocks were shed intermittently into the region, but no other evidence is available to indicate upwarping in the source regions. Possibly, source areas for interval D detrital rocks were so distant that movements in those areas were not related to movements in the Eastern Interior basin. Another possibility is that the cyclic units represent alternating wet and dry climates in the source area, as proposed by Swann (1964).

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Nebraska-Iowa Region

By MARVIN P. CARLSON

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–F

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

THE NEBRASKA-IOWA REGION[1]

By MARVIN P. CARLSON[2]

ABSTRACT

Mississippian rocks of Kinderhook, Osage, and Meramec age are present in the subsurface of southern and eastern Nebraska and continue eastward to the outcrop belt in eastern Iowa. Interval A (Kinderhook) disconformably overlies Upper Devonian shales in much of Iowa and transgressively overlaps older Paleozoic rocks westward across Nebraska. Older interval A claystone, siltstone, and sandstone were deposited in shallow basins in southeastern Iowa and in southeastern Nebraska. Younger interval A rocks are normal marine carbonates.

The Mississippian seas reached their maximum extent during interval B (Osage) and normal marine carbonate deposition continued. The resulting cherty limestone and dolomite thicken toward southeastern Iowa. During early interval C (Meramec), silts and clays were deposited in southeastern Iowa while normal marine carbonate deposition continued over much of the region. The seas became restricted during interval C as shown by the absence of lower interval C rocks in the northern part of the region and the presence of interval C evaporites in southeastern Iowa. Carbonate deposition was again widespread during upper interval C, although a northeastern source furnished large volumes of clastics.

Interval D (Chester) has not been recognized in the area, although some deposition and later erosion of the interval may have occurred in southeastern Iowa. Epeirogenic movement began in Late Mississippian time and continued into the Pennsylvanian. Mississippian rocks were removed from the Chadron arch, the Cambridge arch, and the Nemaha anticline in Nebraska and thinned over the Redfield arch and the Lincoln fold in Iowa. This truncated surface is disconformably overlain by the Cherokee Group of Middle Pennsylvanian age in most of the region. Cretaceous and Quaternary deposits overlie the Mississippian rocks in small areas of northern and eastern Iowa.

REGION DEFINED

This region includes the States of Iowa and Nebraska excepting the Counties of Sioux, Dawes, and Box Butte in extreme northwestern Nebraska (fig. 16). Mississippian rocks are present in the subsurface of southern and eastern Nebraska and continue eastward to the outcrop belt trending from north-central to southeastern Iowa (pl. 1). The interval boundaries within this region coincide, for the most part, with the accepted series terminology: interval A (Kinderhook), interval B (Osage), and interval C (Meramec). There are no rocks of interval D (Chester) recognized in this region.

The Mississippian rocks in northwestern Nebraska have been included within interval B of the adjoining region. (See Mallory, chap. M.) Previous studies by Carlson (1963) had correlated the Mississippian rocks in northwestern Nebraska as Kinderhook (interval A) by comparison of lithology and depositional trends with those of eastern Nebraska. This apparent conflict is largely due to the present uncertainty regarding the position of the Kinderhook-Osage boundary in parts of the northern midcontinent.

PALEOGEOLOGY

LOWER BOUNDARY OF THE MISSISSIPPIAN

The base of the Mississippian in the Nebraska-Iowa region is, for the most part, a transgressive disconformable contact. However, in extreme southeastern Iowa, conodont studies suggest nearly continuous deposition from Devonian into Mississippian within the *English River Sandstone* (Collinson, 1961). Throughout the remainder of this region the lower boundary is a disconformable surface of low relief.

The earliest Mississippian rocks in the region are confined to southeastern Iowa (*English River Sandstone*) and to southwestern Iowa and southeastern Nebraska (*Boice Shale*). The *North Hill Group* overlies and ovarlaps the *English River* in eastern Iowa. The *Chapin Formation* constitutes the base of the Mississip-

[1] This research was supported by the Conservation and Survey Division, University of Nebraska.

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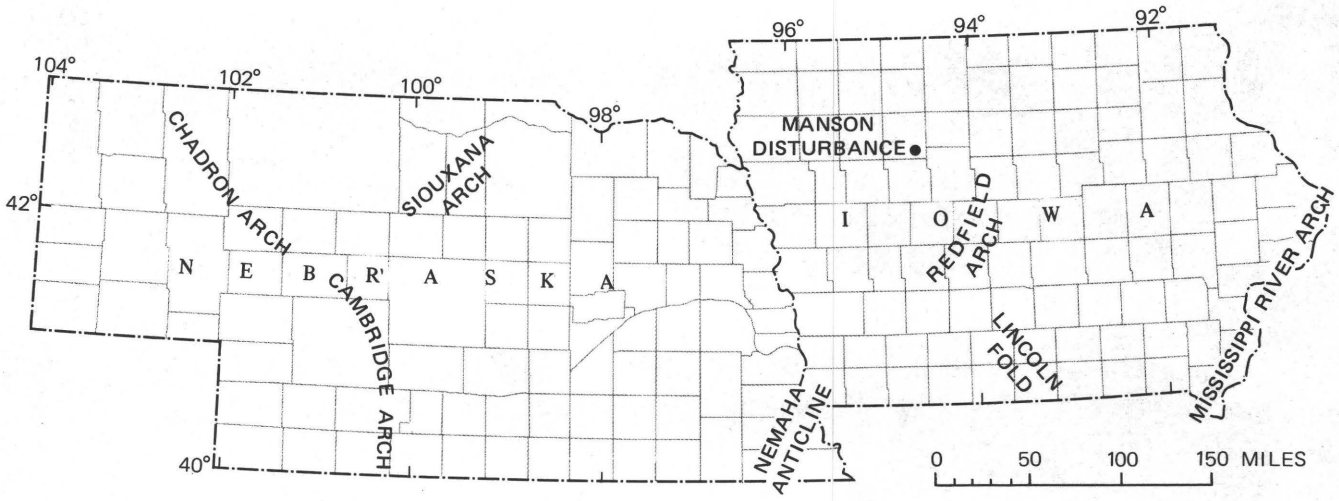

FIGURE 16. — Map of the Nebraska-Iowa region showing post-Mississippian tectonic features and geographic features mentioned in the text.

pian in western Iowa and extreme eastern Nebraska. A general westward to northwestward transgressive overlap occurs in Nebraska, within interval A, with progressively younger formations overlying pre-Mississippian rocks.

UNITS UNDERLYING THE MISSISSIPPIAN

The pre-Mississippian paleogeology is shown on plate 2. The Mississippian conformably overlies the Devonian *Maple Mill Shale* in extreme southeastern Iowa (Harris, 1947). This shale, throughout the remainder of its extent in Iowa, and the Devonian "*Chattanooga Shale*" in easternmost Nebraska have an upper, weathered surface of low relief.

Upper Devonian shales are absent in northwestern Iowa and the remainder of eastern Nebraska, and the Mississippian rests directly on Middle Devonian carbonate rocks. The Mississippian in Nebraska progressively overlaps westward the Devonian and Ordovician to rest upon Upper Cambrian carbonate rocks and sandstone in southwestern Nebraska.

PALEOTECTONIC IMPLICATIONS

Regional uplift had greatly restricted the seas in the Nebraska-Iowa region at the beginning of Mississippian time (Lee and others, 1946). Deposition from Devonian into Mississippian time was nearly continuous in basins centered in southeastern Iowa and southeastern Nebraska. The sea transgressed west and northwest, and Mississippian sediments were deposited over progressively older Paleozoic units. Erosion in post-Mississippian time has removed rocks in central and western Nebraska which would delineate the maximum extent of the Mississippian seas (Reed, 1948). Stratigraphic relations of the existing rocks suggest that Mississippian rocks formerly lapped upon, if not over, the Precambrian rocks along the Siouxana arch (fig. 16).

INTERVAL A

FORMATIONS INCLUDED

The formations included within interval A in the Nebraska-Iowa region are shown on plate 15, columns 58 to 64. In southeastern Iowa they consist of, in ascending order, the *English River Sandstone*, the *North Hill Group* (including the *McCraney Limestone*, *Prospect Hill Siltstone*, and *Starrs Cave Limestone*), and the *Wassonville Limestone Member* of the *Hampton Formation*. Interval A in central Iowa (pl. 9-A, sec. *b*–*b*′) consists of the *North Hill Group* (including the *Prospect Hill Siltstone* and the *Chapin Formation*), the *Hampton Formation* (including the *Maynes Creek, Eagle City,* and *Iowa Falls Members*) and the Gilmore City Limestone. A similar sequence is present in the southwestern Iowa with the addition of the *Boice Shale* at the base.

Equivalent lithostratigraphic units are recognized in Nebraska with some change in terminology (fig. 17). Interval A in eastern Nebraska consists of the formations of the *Hampton* (or *Chouteau*) *Group* and the Gilmore City Limestone. The *Boice Shale* is present in the base of the interval in southeastern Nebraska and southwestern Iowa (Reed, 1946). Only the *Eagle City* and *Iowa Falls Members* undivided and the Gilmore City Limestone are present in central and southwestern Nebraska (pl. 9-A, sec. $c - c'$).

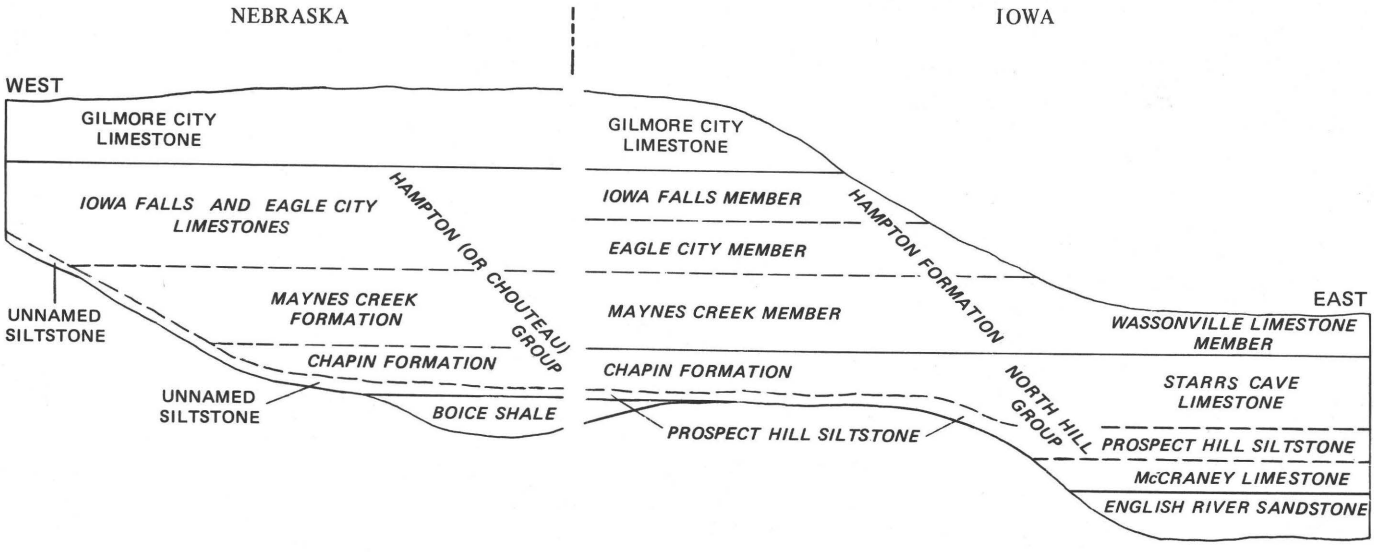

FIGURE 17. — Diagrammatic cross-section illustrating the terminology utilized for interval A in the Nebraska-Iowa region.

STRATIGRAPHIC RELATIONS

The *English River Sandstone* is recognized as a formation only in southeastern Iowa. However, a thin sandstone-siltstone unit has been recognized at the base of the carbonate rocks of interval A throughout Iowa (*Prospect Hill*) and Nebraska (unnamed siltstone, see fig. 17). Conodont faunas from outcrops of the *English River* in southeastern Iowa suggest that the formation may be transitional between the Devonian and the Mississippian (Collinson, 1961). The *English River* conformably overlies the Devonian *Maple Mill Shale* and is disconformably overlain and overlapped by the *North Hill Group.*

The *Boice Shale* is similar in stratigraphic position to the *English River* but is restricted to southeastern Nebraska, southwestern Iowa, and adjoining areas in Kansas and Missouri. The *Boice* disconformably overlies the "*Chattanooga Shale*" in Nebraska and the *Maple Mill Shale* in Iowa, both of Devonian age. It is overlain by the basal siltstone of the *Hampton Group* in Nebraska and by the *Prospect Hill Siltstone* in Iowa.

The *North Hill Group* is most completely represented in southeastern Iowa. In central and western Iowa, this group contains only the *Chapin Formation* and the *Prospect Hill Siltstone* and is overlain by the *Hampton Formation* (Laudon, 1931). In Nebraska, the *Chapin* and the members of the *Hampton* are considered as formations of the *Hampton* (*Chouteau*) *Group* (Carlson, 1963). The units within the *North Hill,* although a transgressive sequence, are all conformable.

The *Maynes Creek* and *Wassonville Limestone Members* of the *Hampton Formation* are lateral equivalents and conformably overlie the *North Hill Group* in Iowa and the *Chapin Formation* in Nebraska (fig. 17). The *Eagle City* and *Iowa Falls Members* are lithologically similar limestones and are combined in this report. They conformably overlie the *Maynes Creek* and overlap this unit in Nebraska. The Gilmore City Limestone, the youngest unit in interval A, conformably overlies the *Hampton.*

UPPER BOUNDARY OF INTERVAL A

The Gilmore City Limestone is the uppermost unit of interval A throughout much of the Nebraska-Iowa region. This formation in Nebraska is overlain by interval B or with regional unconformity by post-Mississippian rocks. In north-central Iowa, the Gilmore City is disconformably overlain by the St. Louis Limestone (interval C).

A disconformable contact between intervals A and B in southeastern Iowa is illustrated on plate 9-A, section $a - a'$. This disconformity was suggested by Laudon (1937) to explain the transgressive overlap of crinoid zones within the Burlington Limestone (interval B). The older crinoid zones are present in southeastern Iowa, and younger zones extend progressively northwestward. A parallel relationship based on lithologic criteria was described by Harris and Parker (1964). Successively younger members of the Burlington Limestone overlie successively younger units of the Kinderhook Series north and west from southeastern Iowa. Uplift and erosion in southeastern Iowa during late interval A time were suggested to explain this unconformable relationship.

Moore (1935) suggested a facies relationship between the lower Burlington (interval B) and the *Iowa*

Falls–Eagle City (interval A). The regional distribution and thickness variations which were observed during the present study also allow this interpretation. However, definitive faunal evidence is not available to support the facies interpretation. The interpretation of a pre-interval B unconformity in southeastern Iowa suggested by Laudon (1937) is accepted for the present report.

THICKNESS TRENDS

A maximum thickness of interval A of more than 300 feet occurs in north-central Iowa. A belt of thick interval A rocks extends southwestward from this maximum into extreme eastern Nebraska (pl. 3-A). The configuration of this belt as originally deposited was altered in post-Mississippian time by erosion brought on by the Nemaha uplift in southeastern Nebraska and the Manson disturbance in north-central Iowa. The *Hampton* is completely represented in west-central Iowa, but thins westward in Nebraska by transgressive overlap during its deposition.

The upper *Hampton* and Gilmore City thin southeastward across Iowa under interval B. This thinning trend for interval A is interrupted by thickening of lower interval A in southwestern Iowa (*Boice Shale*) and southeastern Iowa (*English River* and *North Hill*). Pre-interval B erosion has removed most of interval A in extreme southeastern Iowa (Harris and Parker, 1964).

LITHOFACIES TRENDS

The lithofacies of interval A is dominantly carbonate rocks (pl. 3-B). A thin sandstone-siltstone unit at the base of the interval A carbonate rocks becomes apparent along parts of the eroded edges of the interval. The presence of the *Boice Shale* is indicated by a more shaly lithofacies pattern in southeastern Nebraska and southwestern Iowa. The detrital pattern in southeastern Iowa is due to the *English River Sandstone* and thin beds of similar lithology in the overlying *North Hill Group.*

Chert is present in a nearly constant volume within the *Maynes Creek Member* in central and western Iowa. The absence of a chert percentage overprint for the thicker parts of interval A in this area is caused by the thickening of the noncherty units. The chert content decreases toward southern Iowa, and chert is absent westward of the depositional overlap of the *Maynes Creek.*

SOURCES OF SEDIMENT

The *Hampton Formation* and Gilmore City Limestone are nearly pure, normal marine carbonate rocks. At the base is a quartz detrital zone representing an initial transgressive deposit. The mudstones of the *Boice Shale* were locally derived from the underlying and more extensive "*Chattanooga*" and *Maple Mill Shales* of Devonian age. The quartz sands and silts of the *English River* and *North Hill* were probably derived from lower Paleozoic and Precambrian sedimentary rocks exposed to the northeast, and some material was derived locally from the *Maple Mill Shale.* A part of the Siouxana arch remained emergent in western Nebraska throughout interval A but contributed little sediment.

ENVIRONMENTS OF DEPOSITION AND CLIMATE

The quartz sands and silts of the *English River Sandstone* represent nearshore deposits within a seaway extending into Iowa from the southeast. Red and gray muds of the *Boice Shale* were deposited in shallow water in a basin restricted to southeastern Nebraska and adjoining areas. A deeply weathered land area of low relief occupied most of Iowa and Nebraska marginal to these seas. Red soil at the top of the *Maple Mill* in central Iowa suggests a warm humid climate.

Carbonate deposition began in southeastern Iowa within the *North Hill Group* in a sea that transgressed to the north and west across the region. The extent of transgression and the sedimentary environments near the middle of late interval A time are shown on plate 11, figure 1. Carbonate rocks of the *Hampton* and Gilmore City represent a period of warm shallow marine deposition, as evidenced by pure limestone containing numerous crinoidal remains and oolitic beds. A dominance of dolomite over limestone in the thicker part of this sequence in central and western Iowa (fig. 18) may indicate a slight restriction of circulation in the deeper part of the basin.

PALEOTECTONIC IMPLICATIONS

A period of general emergence at the beginning of the Mississippian is evidenced by the weathered zone on the *Maple Mill Shale* in central Iowa (Harris, 1947) and on the Devonian carbonate rocks in central Nebraska. Seas were restricted to southeastern Iowa (*English River*) and southeastern Nebraska (*Boice*). Slow downwarping allowed the seas to transgress westward and they were widespread in the northern midcontinent late in interval A time (pl. 10, fig. 1). Sinking was somewhat more pronounced in western Iowa than elsewhere.

Uplift and erosion in latest interval A time removed the Gilmore City and upper part of the *Hampton* in southeastern Iowa (Laudon, 1937). The absence of interval A in southeastern Nebraska is due to erosion on the post-Mississippian Nemaha anticline.

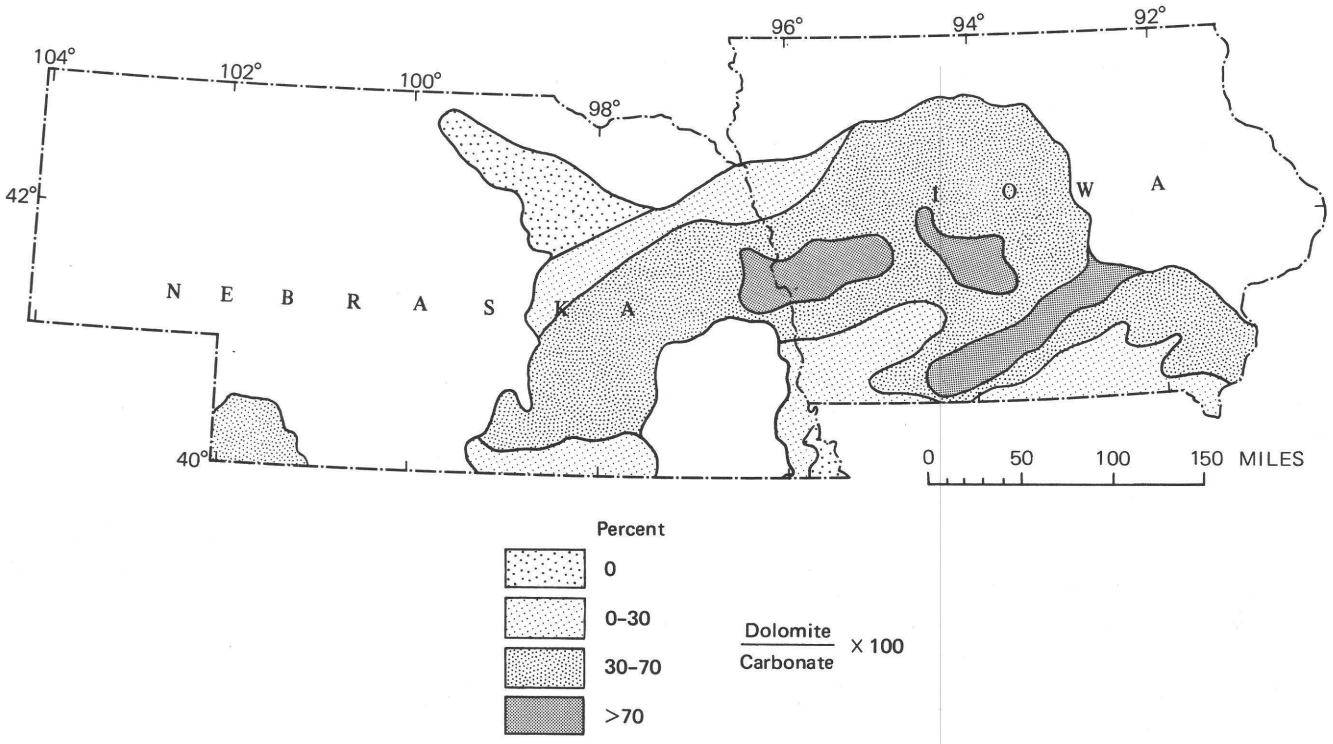

FIGURE 18. — Dolomite/total-carbonate percentage map for interval A in the Nebraska-Iowa region.

PALEOGEOGRAPHIC INTERPRETATIONS

The paleogeography of interval A is summarized on plate 12, figure 1. Most of the Nebraska-Iowa region was covered by a shallow, normal marine sea with slightly deeper water in a western Iowa basin. A low landmass is postulated in western Nebraska.

INTERVAL B

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Interval B throughout the Nebraska-Iowa region consists of the Burlington Limestone and overlying Keokuk Limestone (pl. 15, cols. 58–64). A regional unconformity at the base of the Burlington in southeastern Iowa was suggested by Laudon (1937) to account for transgressive overlap of crinoid zones within the formation toward the north and west. The detailed stratigraphy of, and conformable relationships between, the Burlington and Keokuk in southeastern Iowa were described by Harris and Parker (1964). In Nebraska and western Iowa these formations are considered to be conformable and are described as the Burlington and Keokuk Limestones undivided.

UPPER BOUNDARY OF INTERVAL B

Interval B is conformably overlain by the Warsaw Limestone in southern Iowa and eastern Nebraska. In southeastern Iowa the Keokuk (upper interval B) also appears to grade upward into the overlying Warsaw (Harris and Parker, 1964). The Warsaw is absent in north-central Iowa, and interval B is unconformably overlain by the St. Louis Limestone of interval C. Post-Mississippian erosion has removed younger Mississippian units in parts of both Nebraska and Iowa and interval B is unconformably overlain by Pennsylvanian rocks.

THICKNESS TRENDS

The thickness map of interval B (pl. 4-A) shows a general thickening southeastward toward the Illinois basin. A little more than 200 feet of interval B is present where it is thickest in southern and southeastern Iowa. Thinning to the northwest from this area of thick interval B may be due to the transgressive character of the Burlington Limestone.

Because of post-Mississippian erosion, interval B is less widely distributed than interval A (pls. 3-A, 4-A). In Nebraska, interval B is restricted to the deeper basins along the southern and eastern margins of the State where the maximum thickness is slightly more than 100 feet (pl. 4-A; pl. 9-B, sec. $c - c'$).

LITHOFACIES TRENDS

Interval B everywhere is more than 80 percent carbonate rock, as shown by the interval B lithofacies

map (pl. 4-B). A lithofacies map of the combined Keokuk (interval B) and Warsaw (interval C) of southeastern Iowa (Harris and Parker, 1964, pl. 7) shows that the combined units are dominantly a cherty shaly dolomite with a decrease in argillaceous content toward the northwest. The Burlington of southeastern Iowa is cherty crinoidal limestone, and the Keokuk is cherty argillaceous limestone and dolomite.

A percentage map of dolomite to total carbonate for interval B in the Nebraska-Iowa region is illustrated in figure 19. The thicker limestones of the Burlington result in a high carbonate–low dolomite percentage for southern Iowa. In a discontinuous east-trending band in central Iowa and in south-central Nebraska, both the Burlington and Keokuk are dolomite containing a little argillaceous material. Nearly pure limestone makes up the interval north of this dolomite zone in the north-central part of the region and in southwestern Nebraska. Argillaceous material increases upward within the interval and toward the eastern part of the region. Chert decreases toward the northwest across the region (pl. 4-B; pl. 9-B, secs. $a - a'$, $b - b'$).

SOURCES OF SEDIMENT AND ENVIRONMENT OF DEPOSITION

Interval B in the Nebraska-Iowa region was deposited in shallow water in a normal marine environment (pl. 11, fig. 2). The original sediments were mostly crinoid fragments and other reworked fossil debris. Fine clastic material gradually encroached from sources to the east and northeast.

PALEOTECTONIC AND PALEOGEOGRAPHIC INTERPRETATIONS

The Mississippian seas reached their maximum extent during the deposition of interval B (pl. 12, fig. 2), and the lowland area of interval A time in the vicinity of the Siouxana arch in western Nebraska was probably entirely submerged. Sediments in southern Iowa were deposited in a slowly subsiding area that was an extension of the larger Illinois basin to the southeast. Most of Nebraska was a stable shelf area receiving sediment; the shelf separated basins to the north, south, and east. The sediment has subsequently been removed except for remnants in the southern part of the State.

INTERVAL C

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Interval C in the Nebraska-Iowa region includes in ascending order the Warsaw Limestone, the *Salem* (or *Spergen*) *Limestone,* the St. Louis Limestone, and the Ste. Genevieve Limestone (pl. 15, cols. 58 – 63). The Warsaw is conformable with both the underlying rocks of interval B and the overlying *Salem.* The lower part of interval C (Warsaw and *Salem*) is restricted to the southern part of the region because of a disconformity at the base of the overlying St. Louis (pl. 9-C, secs. $a-a'$ $b - b'$). The St. Louis overlaps lower interval C and interval B, and in north-central Iowa it rests upon interval A. The Ste. Genevieve is conformable upon the St. Louis but varies greatly in thickness and distribution owing to post-Mississippian erosion.

UPPER BOUNDARY OF INTERVAL C

Interval C contains the youngest Mississippian rocks in the Nebraska-Iowa region. The upper boundary is a post-Mississippian erosional surface above which are the Cherokee Group of Pennsylvanian age, the Dakota Sandstone of Cretaceous age, or Pleistocene deposits.

THICKNESS TRENDS

The present distribution and thickness of interval C (pl. 5-A) in large part shows the effect of post-Mississippian erosion. Some northward thinning within the interval is the result of truncation and overlap at the base of the St. Louis. Thinning in extreme south-central Iowa is the result of erosion on the crest of the post-Mississippian Lincoln fold (fig. 16). The maximum thickness of interval C is more than 200 feet in southeastern Iowa near the axis of the depositional basin in the region.

LITHOFACIES TRENDS

The lithofacies of interval C is mostly carbonate rock that contains differing amounts of detrital material (pl. 5-B), owing both to facies changes and to the present distribution of the included formations. The Warsaw Formation, which consists of shale and shaly limestone in southeastern Iowa, becomes less argillaceous to the northwest. The *Salem* (or *Spergen*) consists of dolomitic limestone containing minor amounts of quartz sand and is present only in southern Iowa and extreme southeastern Nebraska. The St. Louis Limestone in southeastern Iowa contains an upper zone of shaly limestone and a lower zone of limestone, anhydrite, and gypsum. Gypsum and anhydrite are absent in outcrops and the limestone is brecciated probably because of solution. Shale increases to the west in the St. Louis across southern Iowa, and a sandy limestone facies becomes prominent toward the north. The Ste. Genevieve is limestone containing a high percentage of quartz sand in southern Iowa and is mainly shale in central Iowa.

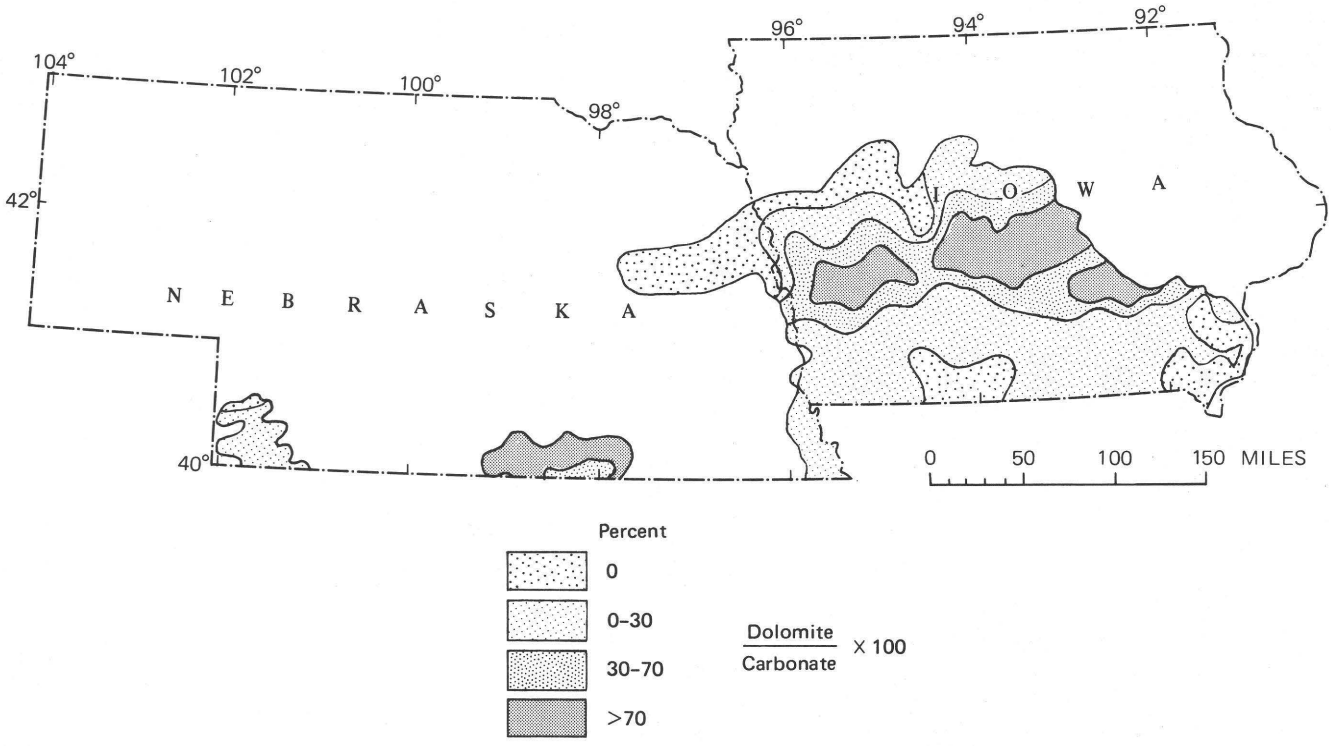

FIGURE 19. — Dolomite/total-carbonate percentage map for interval B in the Nebraska-Iowa region.

Anhydrite and gypsum in the St. Louis coincide with the thickest part of interval C in southeastern Iowa. Detrital patterns (pl. 5-B) in southeastern Iowa reflect sandstone and sandy limestone in the St. Louis and the shaly facies of the Warsaw. Sandy carbonate patterns in south-central Iowa show the content of detrital material in the Ste. Genevieve. An increase of shale in the St. Louis accounts for a shaly carbonate pattern in southwestern Iowa. Detrital patterns in central Iowa result from erosional thinning of the Warsaw and *Salem* carbonate rocks and an increase of sand in the St. Louis and of shale in the Ste. Genevieve.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Carbonates were deposited in freely circulating water of normal salinity from interval B time into the early part of interval C in most of western and central Iowa. In southeastern Iowa, however, an influx of fine detrital material came from the east to form the shaly facies of the Warsaw. Some restriction of the seas is suggested in *Salem* time; quartz sand in the *Salem* was derived from a positive element to the north and northeast. In early St. Louis time the shelf emerged and erosion removed the Warsaw in northern Iowa and the *Salem* in northern and central Iowa. Some restriction of circulation is also shown by the presence of anhydrite and gypsum in the lower St. Louis in southeastern Iowa. The area was submerged and marine carbonate deposition was resumed during later St. Louis time. The northern positive element that furnished sand during deposition of the St. Louis became subdued and yielded mainly finer detritals to the Ste. Genevieve. Environments of deposition near the middle of interval C time are shown on plate 11, figure 3.

PALEOTECTONIC AND PALEOGEOGRAPHIC INTERPRETATIONS

The submerged shelf of earlier Mississippian time was slightly elevated and became emergent during mid-interval C time. Rocks at least as old as interval B were removed by erosion in northern Iowa and adjoining areas. A positive element rose in Minnesota and Wisconsin (pl. 10, fig. 3) and land areas of low relief extended south from it into Iowa and Nebraska. During later St. Louis time most of Nebraska and Iowa was again a slightly negative shelf.

Most of Nebraska and Iowa were covered by shallow marine water during most of interval C time (pl. 12, fig. 3); to the northeast in Wisconsin an area of low hills surrounded by a coastal plains belt is postulated, and in western Nebraska an area of low hills may have been emergent during most of interval C time.

INTERVAL D

Interval D (Chester) is absent in the Nebraska-Iowa region. It is possible that some rocks of this interval were deposited in parts of the region and subsequently were eroded during Late Mississippian and Early Pennsylvanian time. Because of this possibility, interval D is shown projecting into south-central Iowa on the paleotectonic map (pl. 10, fig. 4) for this interval.

Some sandstones assigned to interval D (Yankeetown Chert and Degonia Sandstone) are present in the western part of the Illinois basin, southeast of the Nebraska-Iowa region, and crossbeds suggest southeastward transport of the sand (chap. E, p. 93; Potter and others, 1958; Potter, 1962). These data are the basis for reconstructing a river system flowing southward across a delta and into a marine embayment in southern Iowa (pl. 11, fig. 4; pl. 12, fig. 4) during interval D time.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS AND PALEOTECTONIC IMPLICATIONS

The present thickness of Mississippian rocks in the Nebraska-Iowa region (pl. 7) is mainly a function of post-Mississippian erosion. The most complete sequence is preserved in a basin extending from west-central to southeastern Iowa in which the Mississippian rocks attain a maximum thickness of about 500 feet. Thinning to the northwest reflects a northwestward transgression of the sea in the Early Mississippian.

Post-Mississippian erosion has removed the Mississippian rocks from the Cambridge arch (fig. 16) in south-central Nebraska, from the Nemaha anticline in southeastern Nebraska, and from the area of the Manson disturbance in north-central Iowa. Erosional thinning has occurred over the Redfield arch in central and southwest Iowa and the Lincoln fold in central southern Iowa. Thickness and lithofacies trends in northeastern Nebraska suggest deposition occurred across the Siouxana arch in at least part of Mississippian time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

The Mississippian is unconformably overlain by rocks of Middle Pennsylvanian, Des Moines, age (Cherokee Group) in most of Nebraska and Iowa (pl. 8). Depositional basins were modified in the Early Pennsylvanian by movement along the Cambridge arch, Nemaha anticline, and Mississippi River arch. Pennsylvanian sedimentation began, possibly during Atoka time, in the intervening basins and was gradually expanded until Pennsylvanian rocks completely covered the eroded surface of the Mississippian.

Erosion in Mesozoic time removed Pennsylvanian beds in northern Iowa. Subsequently, the Lower Cretaceous Dakota Sandstone and overlying Cretaceous and Tertiary rocks were deposited. Pre-Pleistocene erosion exposed Mississippian rocks in eastern Iowa. This area is now mantled by glacial deposits and outcrops of the Mississippian are dependent on recent erosion.

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Kansas

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–G

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

KANSAS[1]

By E. D. GOEBEL[2] and G. F. STEWART[3]

ABSTRACT

In eastern and central Kansas the Mississippian rocks rest disconformably on the Chattanooga Shale of Late Devonian and Early Mississippian age; in western Kansas they rest unconformably on rocks of Ordovician age and locally on rocks of Late Devonian age.

Rocks of interval A were deposited in a single cycle of marine transgression and regression. The earliest deposits, the *Boice Shale*, indicate that transgression began in northeastern Kansas; unnamed rocks of Kinderhook age in southwestern Kansas also mark an early Kinderhook marine invasion. Maximum thickness of interval A is more than 150 feet. Detrital rocks, which characterize the oldest interval A deposits, were succeeded by widespread carbonate rocks in the later part of interval A time. Interval A probably covered most if not all of the State. Minor uplift, epeiric tilting, and regression of the strandline took place at the end of interval A.

Interval B records a gradual marine transgression beginning in southeastern and south-central Kansas (*Fern Glen Limestone*); the sea then spread to cover the entire eastern and central part of the State (Burlington Limestone) and eventually covered the entire State (Keokuk Limestone). Although a disconformity has been reported between Osage and Meramec rocks in Kansas, no major hiatus is thought to mark this contact. Southwest Kansas subsided during Osage time to receive a maximum of almost 600 feet of interval B. Interval B is dominantly cherty carbonate rock throughout the State.

Interval C is also dominated by carbonate rocks and represents a continuation of depositional conditions of interval B. In late interval C time minor quantities of detrital siliceous materials were deposited with the carbonate. This detrital influx and the clastic textures in upper interval C suggest an increase in energy of the environment and perhaps a shoaling of the sea. Emergence of the region marked the end of interval C. In southwest Kansas a maximum of about 850 feet of interval C is preserved.

Interval D is preserved in southwest Kansas, where it has a maximum thickness of more than 400 feet. It is also preserved locally as fillings in sinkholes in southeast Kansas. In contrast to the underlying Mississippian, the interval contains appreciable amounts of detrital rocks. Northern Kansas may have been emergent during much of interval D time and some of the detritus in interval D may have been derived from local erosion of older Mississippian rocks, as well as from a source area outside the region.

The end of interval D time marks a change from relatively quiescent tectonic conditions during the Mississippian to more active tectonic conditions that prevailed during the Early Pennsylvanian. Kansas was uplifted and emergent at the end of the Mississippian, the Central Kansas uplift and Nemaha anticline became clearly marked positive features, and erosion stripped many lower Paleozoic beds from the crests of these uplifts and beveled and partially removed Mississippian rocks from the basins. The Mississippian is overlain by rocks of Pennsylvania age throughout the State except in the small area of Mississippian outcrop in extreme southeastern Kansas; Lower Pennsylvanian (Morrow) rocks are limited to the western third of the State, and Middle Pennsylvania (Atoka or Des Moines) lie on the Mississippian across the eastern two-thirds of the State.

REGION DEFINED

This region is restricted to the State of Kansas. Mississippian rocks crop out only in an area of less than two townships in Cherokee County in southeasternmost Kansas (pl. 1). In the subsurface, Mississippian rocks underlie all of Kansas except on the crests of the Cambridge arch, Central Kansas uplift, and Pratt anticline in the central part of the State, and the Nemaha anticline in the northeastern part (fig. 20). With exception of the lower boundary of interval A, the interval boundaries used in this chapter are the same as boundaries of the series recognized by the Kansas Geological Survey: interval A (Kinderhook), interval B (Osage), interval C (Meramec), and interval D (Chester) (pl. 15). Some rocks of each interval are preserved in Kansas.

PALEOGEOLOGY

UNITS UNDERLYING MISSISSIPPIAN

Through most of eastern Kansas (pl. 2) the rocks included in this study are underlain by the Chattanooga Shale of Late Devonian–Early Mississippian age. The Mississippian part of the Chattanooga is indistinguishable lithologically from the Devonian part

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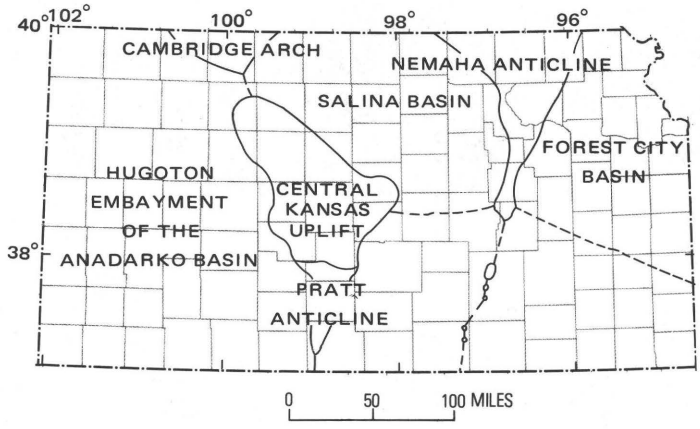

FIGURE 20. — Mississippian and post-Mississippian structures of Kansas mentioned in the text. From Merriam (1963).

almost everywhere in the State. Locally, on the southern part of the Nemaha anticline (fig. 20), the Chattanooga is absent and Mississippian rocks overlie Ordovician rocks, but these areas are so small that they are not shown on plate 2. The term "Kinderhook" has been used by petroleum geologists for both the Chattanooga Shale and the overlying *Boice Shale*, where both are present, in spite of the fact that a disconformity separates these two units (Lee, 1956, p. 67). Erosion of the Chattanooga prior to deposition of the *Boice* is suggested by inverse thickness relations of the two formations.

In western Kansas, Mississippian rocks overlap older rocks northward; they rest successively on locally occurring unnamed Upper Devonian rocks (Goebel and others, 1969) in southwestern Kansas (not shown on pl. 2) and on the Middle and Upper Ordovician Viola Limestone and the Lower Ordovician and Upper Cambrian Arbuckle Group farther north. Mississippian rocks are not preserved on the crest of the Central Kansas uplift (fig. 20), but along the entire west flank of the uplift, Mississippian rocks overlie Ordovician rocks.

LOWER BOUNDARY OF MISSISSIPPIAN

The lower boundary of the Mississippian, as used in this study, is a disconformity throughout the State. The minimum hiatus at this contact is probably in northeastern Kansas where the *Boice Shale*, the oldest unit assigned to interval A, rests on the Chattanooga Shale of Devonian and Mississippian age (pl. 15, col. 67). The hiatus is greatest in northwestern Kansas where the Gilmore City Limestone, the uppermost unit of interval A, overlaps the upper member of the Sedalia Dolomite and rests on the Arbuckle Group of Early Ordovician and Late Cambrian age.

In southeastern Kansas the Compton Limestone (pl. 15, col. 68) rests on the Chattanooga, in central and north-central Kansas the upper member of the Sedalia Dolomite (pl. 15, col. 66) rests on the Chattanooga, and in southwestern Kansas an unnamed detrital unit assigned to interval A (pl. 15, col. 65) rests on the Viola Limestone of Middle Ordovician age. In south-central Kansas interval A is missing and the *Fern Glen Limestone* (interval B) rests directly on the Chattanooga Shale.

PALEOTECTONIC IMPLICATIONS

The Chattanooga Shale masks pre-Mississippian structural features. The pre-Mississippian Ellis arch in northwestern Kansas and the Chautauqua arch in southeastern Kansas probably were connected by the Central Kansas arch during part of pre-Chattanooga time (fig. 21) and acted as a northwest-trending positive element separating the Southwest Kansas basin and the North Kansas basin.

The western edge of the Chattanooga Shale appears to be an erosional edge and has a north trend across central Kansas (pl. 2). The Upper Devonian part of the Chattanooga probably was more widespread than at present and may have extended across western Kansas onto the east flank of the Transcontinental arch in eastern Colorado (pl. 10, fig. 1). Reworked Late Devonian conodonts in the Mississippian rocks in part of western Kansas are supporting evidence to the former presence of Upper Devonian strata in this area (Goebel, 1968), and, locally, in southwestern Kansas sandy carbonate rocks of Late Devonian age are preserved beneath the Mississippian rocks (Goebel and others, 1969).

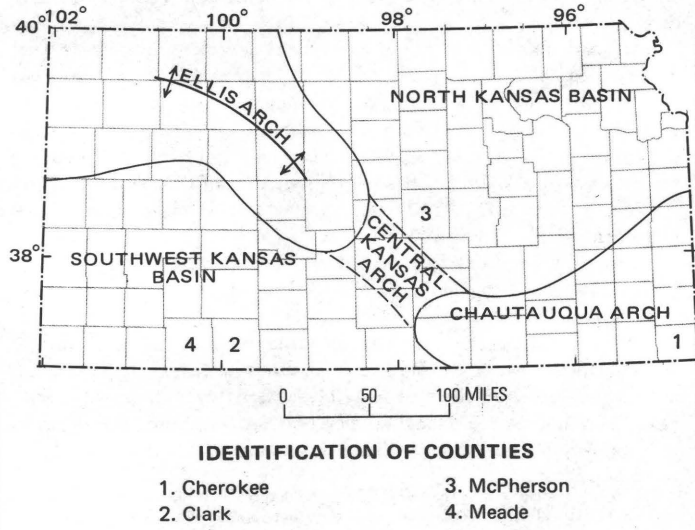

FIGURE 21. — Geographic features and pre-Mississippian structures of Kansas mentioned in the text. Modified from Merriam (1963) and Goebel (1968).

Uplift of the Transcontinental arch in early Kinderhook time may have allowed erosion and removal of the Chattanooga from western Kansas and may have provided an eastward-sloping surface upon which the beds of interval A were deposited, first in northeastern Kansas, but later across most of the State.

INTERVAL A

FORMATIONS INCLUDED

The formations included in interval A are given on plate 15 (cols. 65–68). In southeastern Kansas interval A consists of, in ascending order, the Compton Limestone and the lower member of the Sedalia Dolomite or the Northview Shale. In northeastern Kansas interval A consists of the *Boice Shale*, the Chouteau Limestone (equivalent of the Compton Limestone), the upper member of the Sedalia Dolomite, and the Gilmore City Limestone. In central and northern Kansas interval A consists only of the upper member of the Sedalia overlain by the Gilmore City Limestone. In western Kansas, unnamed rocks of Kinderhook age are overlain by the upper member of the Sedalia(?). These in turn are overlain by the Gilmore City Limestone.

STRATIGRAPHIC RELATIONS

The *Boice Shale*, recognized only in northeast Kansas and adjacent States, unconformably, overlies the Chattanooga Shale and seemingly is conformable below the Chouteau of Moore (1928). Moore and others (1951) identified the Compton Limestone as the lowest limestone formation of the Kindherook in Kansas, and they correlated the Northview Shale with the lower member of the Sedalia Dolomite. The upper member of the Sedalia Dolomite as defined by Lee (1943) is recognized only west of the Nemaha anticline.

Unnamed rocks of Kinderhook age, mostly beds of sandstone and shale, occur on the southwest flank of the Central Kansas uplift (Goebel, 1968). The rocks contain a mixed fauna of Kinderhook and Late Devonian conodonts. The unit occurs unconformably above Ordovician and possibly Devonian carbonate units and below well-established Kinderhook and Osage strata. Goebel (1968) regarded the unit as late Kinderhook in age, but older than the upper member of the Sedalia. The stratigraphic relations with *Boice*, Compton, Chouteau, and lower member of the Sedalia are undetermined.

The upper member of the Sedalia Dolomite overlain by the Gilmore City Limestone is widespread in northeastern and north-central Kansas and locally in far northwestern Kansas. In the Salina basin the upper member of the Sedalia Dolomite is separated by disconformities from overlying and underlying units (Lee, 1956).

The *Misener sand* is distributed erratically through central and eastern Kansas at the base of the Chattanooga Shale. West of the Central Kansas uplift where the black shale of the Chattanooga is absent, the term *Misener* is applied, probably incorrectly, to sandstone below rocks of Mississippian age and above rocks of Ordovician age. This sandstone may have been deposited at any time from Middle or Late Ordovician to Early Mississippian, and the extension of the name *Misener* does not seem justified. This sandstone, where present, has been included in interval A of western Kansas.

The stratigraphic relations and age of the Gilmore City Limestone are better known than those of other Kinderhook units. Late Kinderhook conodont faunas are present in both the upper member of the Sedalia and Gilmore City in west and northwest Kansas (Goebel, 1968; Klapper, 1971). Supporting the physical correlation of the Gilmore City of western Kansas with the type Gilmore City of Iowa is the presence of *Patrognathus andersoni*. This genus has been found in the United States only in this formation and in the Lodgepole Limestone of Montana.

UPPER BOUNDARY OF INTERVAL A

The Gilmore City Limestone is the uppermost unit of interval A throughout much of Kansas; several different units of interval B rest on it or on lower beds in interval A, indicating a disconformity at the contact. Only in McPherson County in central Kansas (Lee and Girty, 1940) does the section approach completeness; there, the Gilmore City underlies the oldest unit in interval B, the *St. Joe Limestone Member* of the *Fern Glen Limestone*. In northeastern Kansas the Gilmore City Limestone is overlain by the Burlington and Keokuk Limestones undifferentiated; in western Kansas the Keokuk Limestone of interval B rests disconformably on the Gilmore City Limestone (Goebel, 1968); and in parts of southwestern Kansas, west of the Central Kansas uplift and the Pratt anticline, the *Fern Glen Limestone* rests on the lowest Kinderhook unit, the unnamed beds of Kinderhook age.

The area in which the oldest formation of interval B, the *Fern Glen*, is recognized almost excludes the area in which the youngest units of interval A, the Gilmore City and upper member of the Sedalia, are recognized (Goebel, 1968, figs. 14, 16); and this fact suggests that they might be facies equivalents. The presence of an Osage Bachtrognathodid conodont fauna in the *Fern Glen Limestone* in south-central Kansas and a Kinderhook Siphonodellid fauna in the Gilmore City argues against this interpretation.

THICKNESS TRENDS

The maximum thickness of interval A is more than 150 feet in western Kansas, almost 150 feet in northeastern Kansas, and more than 100 feet in several small areas in north-central Kansas (pl. 3-A).

In the southern Kansas area, except at its west end, interval A is less than 100 feet thick (pl. 9-E, sec. *D–D'*). In western Kansas the interval is more than 150 feet in thickness (pl. 9-E, sec. *E–E'*).

LITHOFACIES TRENDS

Interval A is dominantly carbonate rock. In north-central and parts of western Kansas, however, terrigenous detritus is contained locally in interval A (pl. 3-B).

The Gilmore City is a carbonate unit in northwestern Kansas and a detrital unit in southwestern Kansas (pl. 9-E, sec. *E–E'*), which fact accounts for the carbonate lithofacies pattern in the central and northern parts of western Kansas and the detrital patterns in southwestern Kansas (pl. 3-B). The shaly lithofacies pattern in southeastern Kansas is evidence of the Northview Shale, a unit that was probably derived from a nearby source in Missouri.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The unnamed rocks of Kinderhook age in western Kansas and the overlying upper member of the Sedalia(?) Dolomite and Gilmore City Limestone are a lithologic sequence which illustrates deposition during a marine transgression, as described by Visher (1965, p. 56). Detrital sediment was deposited as a thin layer; reworking by wave action was probably effective to a depth of only a few feet. Little of the sediment was derived from underlying sediments. Instead, sources of the sand- and silt-sized detrital particles of interval A were probably from Paleozoic quartz-rich rocks exposed on the Ellis and Chautauqua arches. In western Kansas the basal part of the sandstone sequence contains coarse particles, and the sand is well sorted. Above the basal part of the sandstone the unit grades abruptly into finer grained and more poorly sorted rocks, which in turn grade abruptly into the overlying carbonate unit without intermediate shaly beds.

The Gilmore City Limestone of western and north-central Kansas consists of units composed of brachiopods, crinoids, and bryozoan fragments mixed with oolites and pellets and lenticular beds of oolitic limestone. The intercalated beds of oolitic and fossil-fragmental limestone indicate current and wave action in a shallow marine environment (pl. 11, fig. 1). The water was saturated with calcium carbonate. The presence of glauconite, phosphates, and chemical precipitates indicates slow sedimentation. The shallow warm seas prevailed to the end of the Kinderhook. The primary sediment source for the carbonate rocks of interval A was organic.

PALEOTECTONIC IMPLICATIONS

Pre-Mississippian positive elements (fig. 21), the Ellis arch, Central Kansas, and Chautauqua arches, probably were covered by the Chattanooga Shale (Goebel, 1968, p. 1738). After a period of uplift and erosion that stripped the Chattanooga to its present erosional limit, eustatic rise of the sea or epeiric subsidence of the land permitted the marine transgression of interval A time (pl. 10, fig. 1). The sea advanced progressively westward across a surface that sloped gently to the east and was widespread late in interval A time. The Central Kansas uplift may have risen slightly and remained above sea level during early interval A time; at least, rocks of only late Kinderhook age are preserved in northern and central Kansas. To the west in Colorado, the Transcontinental arch probably was neutral or slightly positive and, possibly, was emergent (pl. 12, fig. 1) during interval A time (Goebel, 1968, p. 1738).

At the end of interval A time, the entire region was raised above sea level and interval A was beveled. The beveled surface may have been tilted southward at the beginning of interval B deposition (Lee, 1956, p. 140) because deposits of early Osage age thin northward and later Osage and Meramec units (intervals B and C) overlap in succession northward on Kinderhook rocks.

INTERVAL B

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Interval B consists of the *Fern Glen Limestone* and the Burlington-Keokuk Limestones undifferentiated (pl. 15, cols. 65–68). The *Fern Glen* is restricted to southern and southeastern Kansas and consists of two members, the *St. Joe Limestone Member* and the *Reeds Spring Limestone Member*. The *St. Joe* is less widespread and is overlapped by the *Reeds Spring*, which in turn is overlapped by the Burlington in south-central Kansas (Lee and Girty, 1940).

The Burlington and Keokuk have not been differentiated in much of Kansas. In the Salina basin, Lee (1956) mapped the Keokuk unconformably on the Burlington. The hiatus at this disconformity increases in magnitude westward in the State. Although the western limit of the Burlington is unknown, only the

Keokuk has been identified west of the Central Kansas uplift (Goebel, 1968), where the Keokuk rests directly on the Gilmore City Limestone (pl. 9-E, sec. *E–E'*).

The lower part of the Cowley Formation in southwestern and south-central Kansas is included in interval B and is regarded as a facies equivalent to Keokuk, Burlington, and *Fern Glen* Limestones, and also to the Chattanooga Shale (Goebel, 1968, p. 1763). Lee (1956) interpreted the Cowley as deposited in a basin in south-central Kansas and derived by erosion from rocks of Osage age and older; an erosion surface was reported at the base of the Meramec in the Salina basin, and no break was recognized between the Cowley and Warsaw (interval C) in western Kansas (Lee, 1956). Later studies provide both lithologic and paleontologic data that support the interpretation that the Cowley of southwestern and south-central Kansas is a facies equivalent to rocks of the Osage and much of the Meramec elsewhere in the State (Goebel, 1968; Thompson and Goebel, 1968).

UPPER BOUNDARY OF INTERVAL B

The contact between intervals B and C is the contact of the Warsaw and the Keokuk Limestones over a large part of the State. In this study the top of interval B is placed within the Cowley-like rocks of south-central and southwestern Kansas. Lee and Girty (1940) and Lee (1953, 1956) identified a discontinuity at the top of the Osage in parts of the State, and Maher and Collins (1949) reported disconformity in western Kansas. Although local concentrations of glauconite at the base of the Warsaw suggest a disconformable relationship of the Warsaw on the Keokuk, we believe that no major disconformity is present at the interval B–C boundary in Kansas.

THICKNESS TRENDS

In Kansas, interval B rocks are thickest, almost 600 feet, in the Hugoton embayment (Maher and Collins, 1948; Goebel, 1968, fig. 2) in Meade and Clark Counties (pl. 4-A; pl. 9-E, sec. *D–D'*), and they thicken southward into the Anadarko basin of Oklahoma. In this area of thick interval B in southwest Kansas, interval A is very thin or absent. Presumably, interval A was removed by erosion prior to deposition of interval B.

In south-central Kansas interval B thickens to a maximum of more than 300 feet. The interval thins gradually northward and extends a short distance into southwestern and south-central Nebraska. The interval thins markedly around the margins of the Central Kansas uplift and Nemaha anticline; it has been removed by post-Mississippian erosion from these Early Pennsylvanian positive features.

LITHOFACIES TRENDS

The Burlington-Keokuk unit in Kansas is mostly cherty siliceous carbonate. Because the Burlington-Keokuk Limestones are thicker and more widespread than the *Fern Glen Limestone*, the lithofacies map shows (pl. 4-B) cherty carbonate. The formations of interval B, in large part, have been differentiated according to the kinds of chert and differences in the insoluble residues. Where interval B has been eroded during Late Mississippian and Early Pennsylvanian time, a thick residuum of weathered chert called "chat" has formed. These deposits, commonly included in thickness measurements of Mississippian rocks, form a broad band bordering the major tectonic or structural features of Kansas. The lithofacies map (pl. 4-B) shows a discontinuous belt with abundant chert surrounding the Central Kansas uplift.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

In Kansas, interval B rocks were deposited in a shallow sea of normal salinity (pl. 11, fig. 2). Shells and other calcareous hard parts of benthonic organisms composed much of the sediment, and relatively uniform and widely distributed carbonate sediments were deposited. Periodic slight fluctuations of sea level caused local mild reworking of calcareous materials, resulting in calcarenite interspersed with unworked crinoidal limestone beds. Only in southern Kansas, near the southern end of the Central Kansas uplift, is there evidence of land-derived sediment. Here, interbedded red and green partly crinoidal limestone beds make up the lower part of interval B, the *Fern Glen Limestone.*

Shale partings and mud matrix in limestones occur throughout interval B. Fossiliferous and nonfossiliferous chert are present in interval B, and some of the chert is clearly of secondary origin. Incompletely silicified crinoidal debris is common both in the chert and in the limestone.

PALEOTECTONIC IMPLICATIONS AND PALEOGEOGRAPHIC INTERPRETATIONS

At the beginning of Osage time the Kansas region sloped southward. The oldest rocks of interval B, which were deposited in a northward-transgressing sea, pinch out northward. With additional subsidence or eustatic rise, younger parts of interval B probably were deposited throughout the State. The incipient Central Kansas uplift was covered, and deposition extended westward onto the east flank of the Transcontinental arch (pl. 10, fig. 2). The present distribution of interval

B, restricted to structural depressions, is the result of Late Mississippian and Early Pennsylvanian erosion along the Central Kansas uplift and Nemaha anticline.

The Hugoton embayment subsided actively during interval B time and connected southward with the subsiding Anadarko basin in Oklahoma. The Southwest Kansas basin of Devonian time may be considered ancestral to the Hugoton embayment of Mississippian and later Paleozoic time.

The paleogeographic map of interval B (pl. 12, fig. 2) indicates the wide shallow marine transgression of Osage time and shows that the State probably was entirely submerged.

INTERVAL C

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

The Warsaw, Salem, St. Louis, and Ste. Genevieve Limestones compose interval C and are preserved in both the Hugoton embayment and the Forest City basin (pl. 15, cols. 65 and 67). Only the Warsaw and Salem Limestones are recognized in central and northern Kansas. In the central part of southern Kansas, some questionably identified St. Louis is preserved. The Warsaw is conformable below the Salem in western Kansas.

The Salem is less widely distributed than the Warsaw; it is generally preserved basinward from the Warsaw subcrop. Both units were eroded during the Late Mississippian and Early Pennsylvanian. They underlie Middle Pennsylvanian rocks with angular unconformity. Remnants of the Salem normally overlie the Warsaw in the Salina basin, but, locally, in the absence of the Warsaw they overlap unconformably on the Keokuk (Lee, 1956).

In western Kansas the St. Louis is conformable with the underlying Salem and the overlying Ste. Genevieve. Within the St. Louis in western Kansas, however, an intraformational erosion surface marks the initial change from carbonate to detrital (quartz sandstone) sedimentation. All the formations in interval C probably extended across the Central Kansas uplift and Nemaha anticline but were removed, along with Osage and Kinderhook rocks, by pre-Chester and Early Pennsylvanian erosion. The Salem and St. Louis increase slightly in thickness basinward. Ste. Genevieve rocks possibly were nearly statewide in their original distribution.

The upper part of the Cowley in south-central Kansas is included in interval C and is regarded as a diagenetic alteration of the Warsaw, Salem, and part of the St. Louis Limestones. Dolomitization, chertification, and pressure solution are thought to be responsible in large part for this facies.

UPPER BOUNDARY OF INTERVAL C

The upper boundary of interval C is marked by a pre-Chester and (or) Early Pennsylvanian erosion surface. This interval is overlain unconformably by Chester rocks in southwest Kansas where sandstone units of interval D locally fill erosional depressions in the upper part of interval C. Lower to Middle Pennsylvanian rocks overlie rocks of interval C elsewhere (pl. 8).

A karst surface has been identified at the upper surface of interval C in the midcontinent; it is exposed locally in Cherokee County in southeast Kansas, and is recognizable in the subsurface of Kansas (Hiestand, 1938; Lee and Girty, 1940; Lee 1943, 1956; Shenkel, 1955; Goebel, 1968).

THICKNESS TRENDS

Interval C attains a maximum thickness of about 850 feet near the southwest corner of the State in the Hugoton embayment (pl. 5-A). It thins to the north and northeast to a zero line that generally parallels the southwestern margin of the Central Kansas uplift. In eastern and central Kansas interval C is thin and irregularly distributed; it attains a maximum of about 160 feet in two small areas but it is mostly less than 100 feet thick (pl. 9-E, sec. D–D').

The thickness and distribution of interval C are largely the result of post-Meramec erosion, but in a few places part of the thinning of upper interval C units may be the result of depositional thinning on the flanks of contemporaneous positive structures. The Ste. Genevieve may have been less widely distributed than older units of the interval; thinning of the Ste. Genevieve locally may be depositional and may mark pre-Chester regression of the strandline.

LITHOFACIES TRENDS

Interval C is dominated by carbonate rock and chert in different amounts (pl. 5-B). The chert is difficult to distinguish from chert in interval B. The chert of interval C decreases in amount from bottom to top and forms only a minor constituent of the Ste. Genevieve. It seems to be more abundant in areas where the interval is relatively thin, partly because in these areas only the lower beds of the interval are preserved and partly because chert residuum is included with the interval and tends to be concentrated where the interval is thin. This residuum was produced by post-Meramec weathering and was concentrated marginal to the major positive structures.

Interval C contains less shale and claystone than does either interval B or interval D. Small amounts of sand and silt are present from about midway in the St. Louis upward in interval C, especially in far western

Kansas where the greatest thicknesses of the upper part of interval C are preserved.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Deposition of carbonate rocks, largely biogenetic in origin, continued from interval B through more than half of interval C (pl. 11, fig. 3). This was followed by deposition of intraformational limestone conglomerates, oolite, and calcarenite beds and increasing amounts of siliceous detritus. The widespread oolite and calcarenite beds are interpreted as indicating current activity in shallow warm seas. Although the limestone was the product of direct precipitation of carbonate and the accumulation of fossil debris, the siliceous sediment probably was transported into the region from an external source, perhaps southeastward from the Transcontinental arch, or from the Central Kansas uplift, which might have been exposed during part of late interval C time. The change in character of deposits in mid-St. Louis time appears to represent an increase in energy in the environment. The Ste. Genevieve probably was deposited in a regressing sea.

PALEOTECTONIC IMPLICATIONS

Subsidence of the region continued from interval B through most of interval C time, and in southwestern Kansas it was sufficient to accommodate more than 800 feet of sediment (pl. 10, fig. 3.). The erosion surface within the St. Louis in western Kansas may mark a short period of uplift in the region accompanied by brief partial withdrawal of the sea. Subsidence may have slowed following this episode. The influx of detrital materials in late interval C time may indicate minor uplift of a distant terrigenous source, perhaps the Transcontinental arch to the northwest (pl. 12, fig. 3) or perhaps the Central Kansas uplift. The entire region was uplifted slightly and probably was emergent at the end of interval C time. Interval D rests disconformably on interval C in the southwest part of the State, and a pre-interval D karstlike surface developed on interval C and older Mississippian rocks in parts of central and eastern Kansas.

INTERVAL D

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

Rocks assigned to interval D are an unnamed sequence of dominantly clastic rocks of Chester age and are limited to a relatively small part of southwestern Kansas (pl. 15, col. 55). They consist of variegated dominantly green, shale, lenticular sandstone beds, and lesser amounts of lithographic to crinoidal limestone. They probably include both marine and nonmarine beds.

In southeastern Kansas, some dominantly detrital material preserved as sink fillings might have been deposited during interval D time, but these are of such limited extent that they are not shown on any of the maps or sections.

UPPER BOUNDARY OF INTERVAL D

The upper boundary of interval D is a disconformity. Over most of the area of interval D the *Kearny Formation* of Morrow age rests on interval D; but at the eastern extremity of interval D it is overlain by unnamed beds of Atoka age, and in a small area in Clark County interval D is overlain by the Cherokee Group of Des Moines age.

Where the *Kearny Formation* rests on the unnamed rocks of interval D, the two units are difficult to distinguish because of lithologic similarities. The upper part of interval D contains crinoidal limestone beds similar to crinoidal limestone beds in the *Kearny*, but the Pennsylvanian rocks are more arenaceous and glauconitic. The lower beds of interval D are variegated shale and are distinct from the *Kearny*.

THICKNESS TRENDS

A maximum of more than 400 feet of interval D is preserved in southwest Kansas (pl. 6-A). Thinning of interval D on uplifts and local absence as a result of truncation were reported by Beebe (1959). The updip edge of interval D forms an irregular arcuate pattern which suggests that the pinchout is erosional.

LITHOFACIES TRENDS

Interval D is most sandy and muddy near its margins in southwestern Kansas and most calcareous in the central and thicker part (pl. 6-B). In part this is caused by post-interval D erosion. Around the margin only the dominantly detrital lower part of interval D is preserved; where the interval is thicker, however, the dominantly carbonate upper part is also preserved beneath the erosion surface.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Some of the clay, silt, and sand of interval D was probably derived from erosion and reworking of rocks of intervals C and B in northwestern Kansas and the Central Kansas uplift; but it also seems possible that much of the material was derived from outside the

region, perhaps the Transcontinental arch to the northwest.

Deposition was probably in a relatively shallow sea (pl. 11, fig. 4) in which considerable reworking of the detrital sediment took place. The strandline periodically may have migrated long distances across the region, but the northern part of Kansas probably was emergent throughout interval D (pl. 12, fig. 4).

PALEOTECTONIC IMPLICATIONS AND PALEOGEOGRAPHIC INTERPRETATIONS

The 400 feet of preserved interval D in the Hugoton embayment indicates that marked subsidence took place in this part of the State during interval D time (pl. 10, fig. 4). The absence of interval D in the northern part of the State, the possibility that some of the materials contained in interval D were derived from intervals B and C, and the suggestion that the strandline migrated widely during interval D time support the possibility that northern Kansas may have been uplifted slightly and was emergent during much of Chester time. It is also possible that the Central Kansas uplift was a relatively positive feature at this time.

The end of interval D time marks a change from relatively quiescent tectonic conditions during the Mississippian to more active tectonic conditions that prevailed during the Pennsylvanian. Much of Kansas was uplifted before the invasion of Pennsylvanian seas, and an unconformity was formed that in some places is marked by a karst surface.

Geographically (pl. 12, fig. 4), the record of interval D suggests that the southern part of the State may have been covered by a shallow sea; a general east-trending strip of coastal plain may have extended across the northern part of the State which, in turn, may have passed northward into a hilly terrain on the southern flank of the Transcontinental arch.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

Mississippian rocks attain a maximum thickness of more than 1,700 feet in the Hugoton embayment of southwestern Kansas (pl. 7), and they thin to the north to about 100 feet in the northwest corner of the State. In central and in eastern Kansas, the system attains maximum thicknesses between 400 and 500 feet at several scattered places. The zero isopach outlines the Central Kansas uplift in central and north-central Kansas, and the Nemaha anticline in eastern Kansas.

PALEOTECTONIC IMPLICATIONS

Minor movement on the Central Kansas uplift may have taken place during the Mississippian Period, but the outline of the uplift, as well as the Nemaha anticline, shown by the pinchout of the Mississippian, is mainly a result of post-Meramec and pre-Middle Pennsylvanian uplift and erosion.

The total-thickness map (pl. 7) combines a relatively complex history of widespread marine transgressions and deposition, followed by regressions and erosion, and climaxed by widespread erosion at the end of the Mississippian.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Rocks of Pennsylvanian age rest unconformably on Mississippian rocks throughout Kansas (pl. 8). The *Kearny Formation* of Morrow age overlies Mississippian intervals C and D along the western boundary of the State. It is overlapped by unnamed beds of Atoka age, which rest on Mississippian rocks in a narrow belt extending south-southeast from near the northwest corner of the State. The Cherokee Group of Des Moines age overlaps the beds of Atoka age and rests on Mississippian rocks throughout the rest of the State.

The base of the Pennsylvanian generally is marked by a unit of coarse or reworked material. In the western and central parts of the State this unit is a basal conglomerate or sandstone and in the eastern part it commonly is a reworked concentration of chert derived from the residuum at the top of the Mississippian.

PALEOTECTONIC IMPLICATIONS

The unconformable contact between Pennsylvanian and Mississippian rocks marks a period of uplift and erosion that was mainly responsible for the present distribution of the Mississippian intervals. The Central Kansas uplift and Nemaha anticline were sharply uplifted and eroded. Between Meramec (Late Mississippian) and Des Moines (Middle Pennsylvanian) time, all the Mississippian and older Paleozoic strata were stripped from the highest parts of these structures, and Pennsylvanian rocks of Des Moines age were deposited directly on Precambrian.

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Arkansas

By ERNEST E. GLICK

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010-H

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

ARKANSAS

By ERNEST E. GLICK

ABSTRACT

About three-fourths of Arkansas is underlain by rocks of Mississippian age either at the surface or in the subsurface. The entire Mississippian sequence has been eroded from northeastern Arkansas and from central Arkansas. Surface and subsurface data are excellent from the northern third of the State but only fair to very poor from the southern two-thirds.

The Mississippian and Devonian Systems are conformable and lithologically indistinguishable at their boundary in much of Arkansas. A black shale unit that encompasses the systemic boundary covers much of the State. Beyond the northeastern limit of the shale, the systemic boundary is marked by a thin "lag" sand of Devonian age that was in part exhumed in Early Mississippian time and exposed on the sea floor until it was buried again by limestone of Kinderhook age. That sand is underlain by rocks of Devonian, Silurian, and Ordovician age.

The black shale unit is excluded from interval A by placing the boundary arbitrarily at its upper surface. In this study interval A of Arkansas is restricted to the Ozark shelf area where approximately the lower one-third of the St. Joe Member of the Boone Formation — locally as much as 29 feet of crinoidal limestone — is of Kinderhook age and makes up interval A. Equivalent beds in southern Arkansas cannot be distinguished separately and are included with interval B.

Interval B consists of nearly 350 feet of chert, siliceous limestone, and limestone in the Ozark region, less than 200 feet of novaculite in the Ouachita Mountains area, and a thinner layer of dark-gray shale elsewhere. The Ozark shelf facies is unconformably overlain by interval C. During interval B time, the basins were "starved" and source areas were weathering chemically. Any calcium carbonate that was deposited tended to become silicified unless it was covered rather quickly by the next layer of organic debris.

Interval C rocks in the Ozark region accumulated after a slight regression of the Boone Sea. Beds of oolitic limestone, which are herein correlated with the Short Creek Oolite Member of the Boone Formation of northeastern Oklahoma and southeastern Kansas, were deposited in the energetic readvancing sea and were overlain by about 40 feet of "typical" Boone — cherty limestone — in the central and western Ozark region of Arkansas. Around the edge of the shelf and in the Batesville channel the upper part of the Moorefield Formation and the Ruddell Shale were deposited as a basin facies and an offlapping facies, respectively, of interval C.

Interval C time marked the beginning of the accumulation of a thick sequence of clay and sand turbidites in the Ouachita trough. Volcanic debris was deposited in a few beds of tuff, mostly in the Ouachita Mountains area of western Arkansas and Oklahoma. Interval C grades upward into interval D in all of Arkansas except in the northwestern and north-central shelf area where the contact is an unconformity.

Shale is by far the most abundant rock of interval D in Arkansas. Sandstone is second in abundance and limestone is third. All the limestone is retricted to the shelf area; sand making up the sandstone was deposited on the shallowest part of the shelf and in the deepest part of the trough. A sequence of limestone and limy shale deposited on the shelf gradually thickens and grades into nonlimy shale southward in central Arkansas; the sequence thickens abruptly at the trough margin in western Arkansas along a postulated fault. The thickness of interval D is less than 200 feet on the western Ozark shelf and more than 5,000 feet in the Ouachita trough.

Rocks of Mississippian age in Arkansas were deposited in two cycles — a Chattanooga-to-Ruddell transgression and regression during intervals A, B, and C time, and a Batesville-to-Pitkin transgression and regression in interval D time. Maximum regression was reached toward the end of interval C time. In each cycle, mud was deposited from the transgressing sea on the sloping shelf. The mudstone was overlain by biogenic limestone deposited in a moderate- to high-energy marine environment on the subsiding shelf. Reefs formed during the second cycle and, perhaps, during the first cycle. The marine waters of the deep basin to the south furnished nutrients to the limestone-producing shelf fauna more freely during the middle part of each cycle — partly through upwelling.

The maximum withdrawal of the sea at the end of Mississippian time uncovered only the Ozark shelf area. A shoreline coal swamp developed along the slightly emergent coastal plain while erosion of the central shelf was in progress. As the Illinois basin and areas to the east were also subjected to post-Mississippian erosion, sand and clay that had been deposited there was carried into Arkansas via the ancestral Mississippi embayment. In earliest Pennsylvanian time the sand was spread as a blanket across the Ozark region, covering the slightly truncated offlapping units of interval D. Most of the clay and some of the sand was dispersed farther south across the basin.

REGION DEFINED

The region treated in this chapter is the State of Arkansas. Rocks of Mississippian age are exposed in the Ozark region and Ouachita Mountains of Arkansas (pl. 1). They are present in the subsurface of the southern part of the Ozark region, in the Arkansas Valley, in the Gulf Coastal Plain, and in the Mississippi embayment (fig. 22). They have been removed by ero-

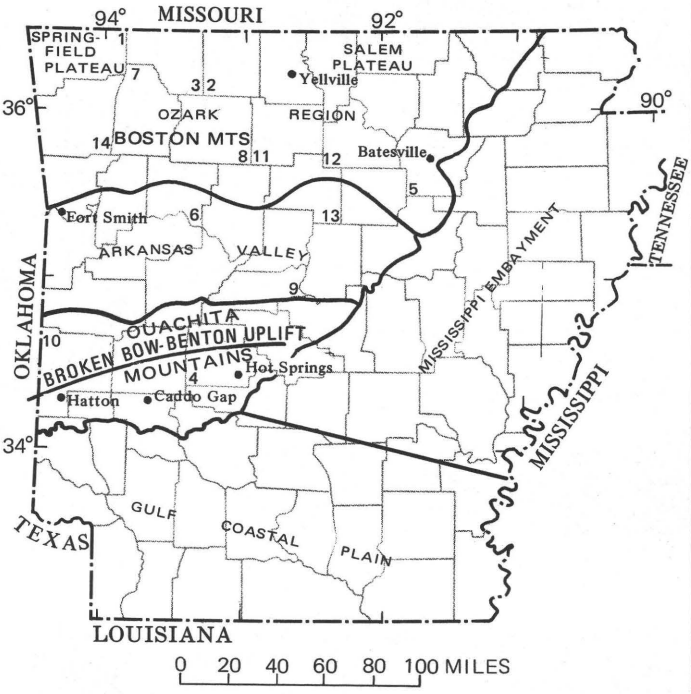


FIGURE 22. — Localities and geographic features of Arkansas referred to in text.

sion — mostly pre-Cretaceous erosion — from a large area in the northeastern Ozark region of Arkansas and its southeastern extension beneath the northern part of the Mississippi embayment and from the core area of the Ouachita Mountains and its extension beneath the Mississippi embayment. Probably at least three-fourths of Arkansas contains rocks of Mississippian age either on the surface or in the subsurface (Arkansas Geological Survey, 1929).

The exposures in much of the northern Ozark region are excellent. There, the resistant Boone Formation underlies the Springfield Plateau (fig. 22) and is exposed along the Eureka Springs escarpment which separates the Springfield Plateau from the Salem Plateau. The Salem Plateau is underlain by rocks of Ordovician age. The younger rocks of Mississippian age are well exposed in cliffs along the Boston Mountains escarpment which is capped by rocks of Pennsylvanian age. In the northern Ozark region the rocks are mostly flat lying.

In the southern part of the Ozark region, streams of a southward-flowing drainage system have cut through the Pennsylvanian cover locally and have exposed rocks of Mississippian age. In that area the rocks have gentle regional southward dip.

In the northern part of the Arkansas Valley, rocks of Mississippian age have been drilled extensively and are well known. Drilling in the southern part of the valley has been too shallow to reach the Mississippian except in a few very deep tests.

The exposures of rocks of Early Mississippian age in the Ouachita Mountains of Arkansas are excellent on the south flank of the core area, which is the Broken Bow–Benton uplift (Miser, 1959, and fig. 22). The thick younger part of the Mississippian (Stanley Shale) is contorted and weathered on both flanks of the uplift to such an extent that few sections have been published.

In the Gulf Coastal Plain and in the southern part of the Mississippi embayment, drilling generally stops at the top of the Paleozoics. The wells that have penetrated part of the subsurface Mississippian sequence are widely spaced; data from them are difficult to interpret because of structural complexities and regional lithologic changes.

PALEOGEOLOGY

LOWER BOUNDARY OF MISSISSIPPIAN

The Mississippian and Devonian Systems are conformable and lithologically indistinguishable at their boundary in much of Arkansas. In the Ouachita Mountains area, the systemic boundary falls within the upper part of the middle division of the Arkansas Novaculite (pl. 2; fig. 23; Hass, 1951). At Caddo Gap (Ark loc. 306A) the systemic boundary is 318 feet above the base and 29 feet below the top of the middle division (fig. 23; Hass, 1951). In the Arkansas Valley and Ozark region, the lower part of the Chattanooga Shale is of Devonian age and the upper part is of Mississippian age (Frezon and Glick, 1959, p. 174; and Freeman and Schumacher, 1969, fig. 1). For consistency in mapping, the Chattanooga — where the shale facies is present — and the middle division of the Arkansas Novaculite are excluded from interval A. In north-central Arkansas the Chattanooga is represented only by a thin sandstone, commonly less than 1 foot thick, that locally contains Devonian conodonts in its lower part and Mississippian conodonts in its upper part (Freeman and Schumacher, 1969, fig. 1; and Thompson and Fellows, 1969, fig. 3). This thin unit of sandstone — which is comprised of "lag" sand or marine regolith that accumulated on the sea floor during Late Devonian time (Sylamore regolith), was in part reworked during early Mississippian time (Bachelor regolith), and has since been indurated — is included with the Mississippian rocks where the shale facies of the Chattanooga is absent in the Ozark region (fig. 24).

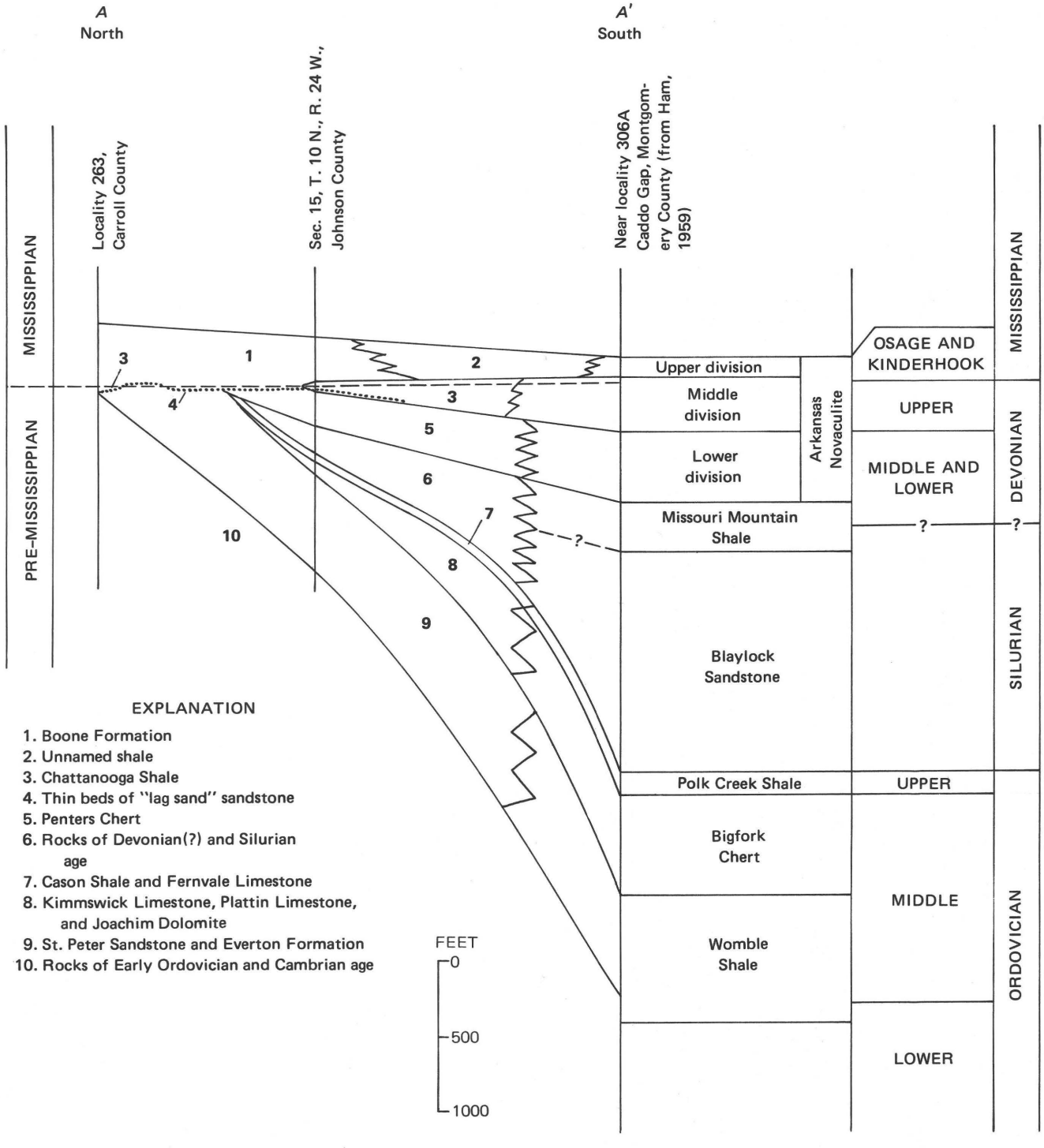
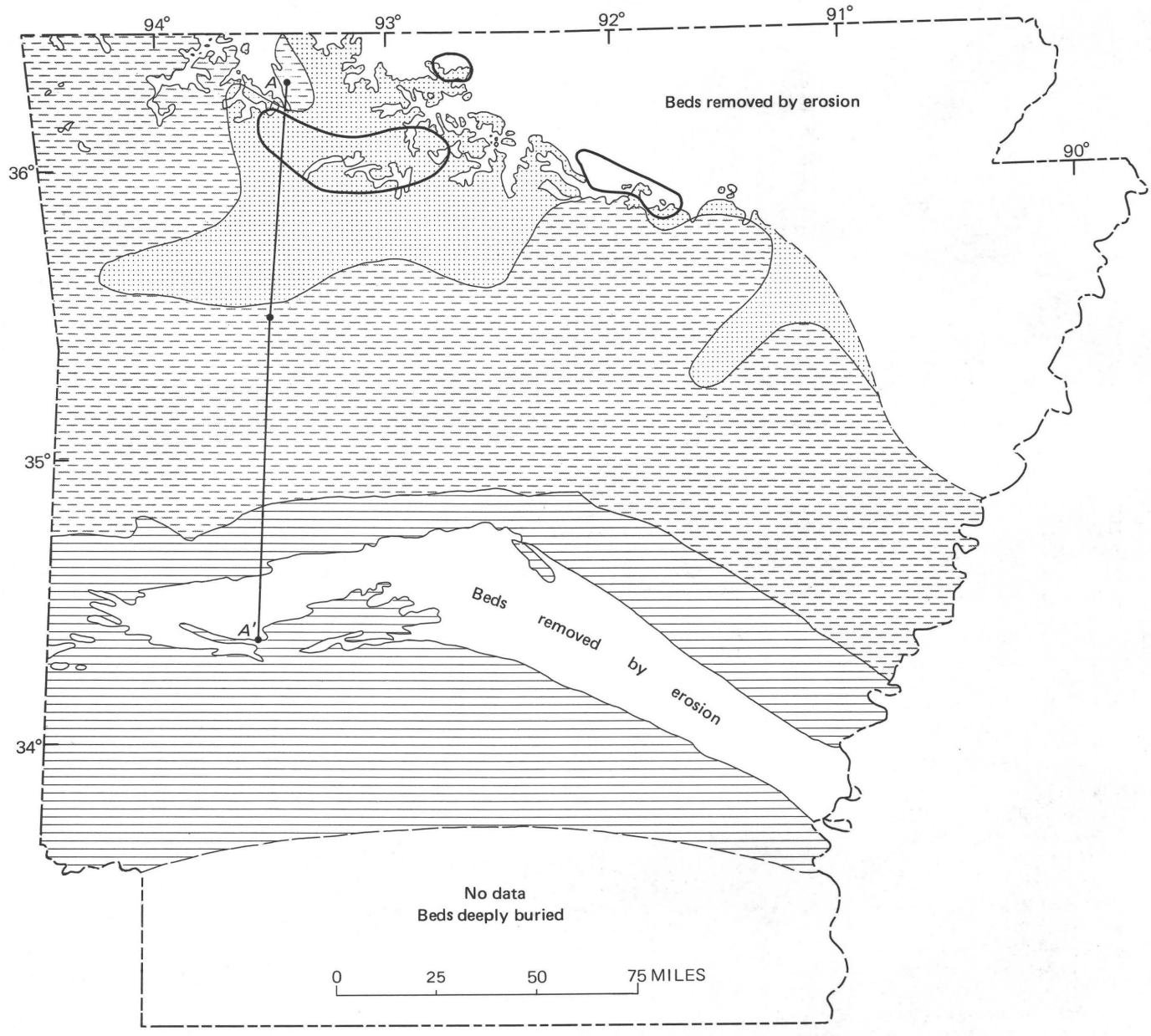


FIGURE 23. — Cross section showing relations of Lower Mississippian and pre-Mississippian formations from the northern to the southern part of Arkansas. Line of section shown in figure 24.

UNITS UNDERLYING MISSISSIPPIAN

In the Ouachita Mountains area the Mississippian System rests in apparent conformity on Upper Devonian beds of the lower part of the middle division of the Arkansas Novaculite. The entire middle division of the Arkansas Novaculite is excluded from this study. In the Arkansas Valley and most of the Ozark region, the systemic boundary lies within the Chattanooga Shale. For convenience, however, the shale facies of the Chat-



EXPLANATION

Area in which Chattanooga Shale contains both shale facies and basal sandstone facies — The sandstone (Devonian) pinches out basinward (southward) at about the edge of the shelf

Area in which sandstone forms basal bed of the Boone Formation beyond the limit of shale facies of the Chattanooga — Sandstone is of Devonian age or is of Devonian age in its lower few inches and of Mississippian age in its upper few inches or is all of Mississippian age

Generalized areas in which conodonts as young as Kinderhook in age occur at the base of the basal Boone sandstone

Area of silicified claystone and siltstone of the middle division of the Arkansas Novaculite

Line of section of figure 23

FIGURE 24. — Distribution of rock types in units transitional from Late Devonian to Early Mississippian age in Arkansas.

tanooga is excluded arbitrarily from this study, and on plate 2 the Mississippian rocks are shown as resting on Chattanooga Shale of Mississippian and Devonian age. North of the limit of the shale in the Chattanooga, a thin sandstone at the base of interval A rests on the Penters Chert of Early and Middle Devonian age and on formations of Silurian and Middle and Late Ordovician age.

DEPOSITION OF THE CHATTANOOGA SHALE

Whether the sea actually withdrew from northern Arkansas just prior to Chattanooga time and readvanced during Chattanooga time is debatable. Most, if not all, previous authors have reasoned that it did withdraw (McKnight, 1935, p. 72; Swanson and Landis, 1962; Freeman and Schumacher, 1969); the present author reasons that it did not. The shale facies of the Chattanooga and the thin sandstone unit at its base were deposited in a transgressing sea that probably remained in the area following early Middle Devonian deposition. The sand (derived locally by slow submarine erosion of the shelf area), along with a larger volume of clay and silt, was winnowed and transported across the shelf by gentle sea-bottom currents. The clay and silt were swept out of the shelf area while the sand was spread as a thin blanket across the shelf. Phosphatic nodules and glauconite grains were formed in this relatively quiet sea-bottom environment and became incorporated in the basal sandy unit. As the sea deepened or lost energy, bottom currents around the edge of the shelf became too weak to transport sand grains and, eventually, too weak to transport their entire load of clay and silt basinward across the outer sand deposits. Then, the deposition of organic-rich clay and silt that had been restricted to the basin to the south was slowly extended shelfward. Undoubtedly, the clay was deposited northeastward beyond its present limit and may have been deposited across nearly all of the Ozark shelf of Arkansas.

Early in the Mississippian Period, following the maximum northward onlapping of the shale facies of the Chattanooga, the sea became more energetic and removed the thin, probably discontinuous, layer of newly deposited clay from an area of at least 5,000 square miles in the north-central and northeastern part of the Ozark region of Arkansas and the adjacent part of the upper Mississippi embayment (fig. 24). The sandy unit again was exposed to sea-floor conditions but was disturbed only enough to mix into it the durable remains of an Early Mississippian fauna —part of which may have been exhumed as the layer of clay was removed and part of which may have been indigenous. Throughout much of that area, the upper few inches — and locally all — of the sandy unit of Devonian age was reworked in Mississippian time.

Farther south, beyond the outer limits of the shelf, dark-gray clay accumulated without interruption from late Middle Devonian through Mississippian time. In the central and southern Ouachita Mountains area, still deeper water occupied a broad trough. A thick layer of organic-rich clay and fine silt (middle division of the Arkansas Novaculite) accumulated and was in part silicified. The Devonian-Mississippian systemic boundary is marked in both areas largely by a change in the conodont fauna.

PALEOTECTONIC IMPLICATIONS

Throughout Late Ordovician to Mississippian time, the northwestern and north-central Ozark region of Arkansas tended to subside less, to receive a thinner layer of sediment, and to be beveled more deeply during periods of erosion than did the areas to the south and east. As much as 500 feet of Middle and Lower Ordovician rock was removed by erosion from the most positive part of the Ozark region of Arkansas, and no additional pre-Chattanooga sediments accumulated. In areas to the south and east, where subsidence was greater, a nearly complete Lower and Middle Ordovician sequence is present, as well as additional units of Ordovician, Silurian, and Devonian age that accumulated. A nearly complete but only moderately thick sequence of pre-Mississippian rock accumulated in the Ouachita Mountains area (fig. 23; Ham, 1959, fig. 1).

Tectonic movement, perhaps a result of adjustment to relative load, tended to keep the more positive shelf slightly convex, in spite of erosion, and to keep the adjacent basin floor slightly concave. The line of inflection between the two areas shifted irregularly with time but in general moved shelfward. Of the several upper Middle Ordovician to Middle Devonian formations deposited across the line of inflection, (1) most are slightly sandy at their base locally (St. Peter-like sand) and rest unconformably on older units and (2) most overlap the truncated edge of the subjacent unit locally. These relations suggest that the peneplain at the base of the Chattanooga Shale in northern Arkansas is the result not of a single period of erosion but of development at near sea level during 100 million years of pre-Chattanooga time.

Paleotectonic activity in Arkansas during earliest Mississippian time consisted only of slight subsidence of the sea floor. The slight deepening of the sea and a substantial increase in the circulation of water across the shelf permitted the start of deposition of a relatively thick sequence of carbonate. During the Mississippian,

relatively rapid subsidence took place in an area that had been at near sea level but tectonically only slightly negative since Middle Ordovician time (fig. 23).

INTERVAL A

FORMATIONS INCLUDED

Although several formations in Arkansas include rocks of Kinderhook age, only a lower part, approximately the lower one-third, of the St. Joe Member of the Boone Formation has been mapped as interval A in this report, and interval A is shown as restricted to the north-central and northwestern parts of the State (pls. 3-A, 3-B). This assignment of the St. Joe is made on the basis of conodont determinations (Freeman and Schumacher, 1969, p. 2329; Thompson and Fellows, 1969, p. 7–23 and appendix, p. 145–257).

Other rocks of Kinderhook age in Arkansas excluded from interval A, because they are not readily identifiable, include parts of the Chattanooga Shale and Arkansas Novaculite. The Walls Ferry Limestone (Gordon, 1964, p. 8–12) is mapped with the St. Joe Member.

STRATIGRAPHIC RELATIONS

In the western, southern, and eastern parts of the Ozark region of Arkansas, the St. Joe Member of the Boone Formation appears to rest conformably on the Chattanooga Shale but may rest unconformably on the Chattanooga locally. In the central Ozark region of Arkansas, the St. Joe rests on thin beds of sandstone that are in essence marine regoliths developed during two periods of submarine erosion and winnowing — Devonian time and pre-St. Joe Kinderhook time (pl. 15, cols. 69, 70). Locally, some of the sand grains are reworked into the lower few inches of the St. Joe.

The St. Joe Member thins southward to a zero line (pl. 3-A) and probably does not grade into or come into contact with basinal, relatively deepwater units of equivalent age to the south, being separated from them by a zone of nondeposition. The St. Joe Member apparently grades upward into the overlying siliceous facies of the Boone.

In the Ouachita Mountains area the middle and upper divisions of the Arkansas Novaculite are conformable and are not everywhere separable. The upper division is mapped arbitrarily as interval B and the middle division is mapped arbitrarily as pre-Mississippian, although both divisions almost certainly contain beds of interval A (Kinderhook) age. Probably, the upper division grades laterally and vertically into the Stanley Shale (B. R. Haley and C. G. Stone, oral commun., 1971). In the Arkansas Valley a thick sequence of dark-gray shale is considered to be the product of nearly continuous deposition from Middle Devonian through Mississippian time (Maher and Lantz, 1953b). Apparently, the upper division of the Arkansas Novaculite grades northward into part of an equivalent shale sequence — Stanley Shale on plate 15, column 71.

UPPER BOUNDARY OF INTERVAL A

Paleontologic data (Thompson and Fellows, 1969) provide a relatively definitive guide to the top of interval A, which in the Ozark region is the top of rocks of Kinderhook age, and suggest that the interval A–B boundary within the St. Joe Member of the Boone Formation is conformable. South of the probable zero line of the St. Joe Member interval A is regarded as absent. Farther south the upper boundary of interval A can be projected arbitrarily through a black shale sequence, locally through the lower part of the Stanley Shale, and into the upper division of the Arkansas Novaculite of the Ouachita Mountains area. No data are available to locate the position of the boundary within the upper division of the Arkansas Novaculite, but some of the lower part appears to be of Kinderhook age (Hass, 1951). No estimate of the thickness and distribution of that segment of the upper division, nor of interval A south of the limit of the St. Joe Member, is attempted.

THICKNESS TRENDS

The maximum thickness of rocks in northern Arkansas assigned to interval A is 29 feet of the lower part of the St. Joe Member of the Boone Formation in northern Boone County that, on the basis of conodonts, is of Kinderhook age (Thompson and Fellows, 1969, p. 185). The thickness of the Kinderhook decreases progressively to the southeast and south proportionally with the decrease in thickness of the St. Joe Member. Regardless of how thin the St. Joe Member becomes, only the lower part has been determined as of Kinderhook age.

In the Arkansas Valley and in the Ouachita Mountains areas, interval A is not mapped. However, rocks of Kindherook age — the upper part of the Chattanooga Shale, the upper part of the middle division of the Arkansas Novaculite or its shale equivalent, and perhaps some of the lower part of the upper division of the Arkansas Novaculite — attain a thickness of at least 29 feet (Hass, 1951).

LITHOFACIES TRENDS

Only a shallow-water facies of rocks of Kinderhook age is mapped in interval A, and it is mostly limestone (pl. 3-B). Some of the lower part of the St. Joe Member of the Boone Formation is calcareous greenish-gray

shale, but by far the major part is crinoidal limestone in well-defined beds that are generally parallel to the base of the unit and are about 4 inches thick. Both in the subsurface and on the surface, the St. Joe tends to be pink to reddish brown or greenish gray. Some beds are made up of small reddish-brown crinoid columnals in a finely crystalline greenish-gray clayey limestone matrix.

The rocks of Kinderhook age in the Arkansas Valley and Ouachita Mountains areas are dark-gray shale or dark-gray shale containing a few beds of novaculite or novaculite conglomerate. These rocks are arbitrarily excluded from interval A and are not shown on plate 3-B.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Almost all of the rock assigned to interval A in Arkansas is crinoidal limestone. The associated clay is detritus from distant, unidentified sources. Sand grains in the lower part of the sequence were derived locally, largely in pre-Mississippian time, from rocks of Ordovician age that lay exposed to erosion on the sea floor. The sand was further winnowed during early interval A time in a marine environment.

Shortly after the beginning of interval A time the sea covering the Ozark region changed from one in which deposition was extremely slow to one in which carbonate sediment started to accumulate rapidly. Crinoids began to flourish, probably first on sand-covered shoals and then outward across the adjacent clay-covered sea floor. Marine currents, probably moving northwestward across the shelf, brought in nutrients from deeper water areas to the south and dispersed the columnals and other organic debris, building an ever-expanding platform for the next generation of crinoids. Locally, other animals found their niche among the crinoid "forests," some helping to build small bioherms on the sea floor. Normal marine conditions continued to the close of interval A time (pl. 11, fig. 1). The debris from the crinoid colonies continued to be restricted to the shallow-water Ozark region; clay was accumulating in deeper water to the south. A zone around the outer edge of the shallow-water area probably received little or no sediment.

PALEOTECTONIC IMPLICATIONS AND PALEOGEOGRAPHIC INTERPRETATIONS

Probably all of Arkansas was covered by the sea throughout interval A time (pl. 12, fig. 1). No tectonic activity of any magnitude is reflected in the sediments of the region; however, slight subsidence probably was required in the central part of the Ozark region as well as in the area to the south (fig. 25*A*) to accommodate the relatively thin layers of sediment that were deposited during interval A time (pl. 10, fig. 1).

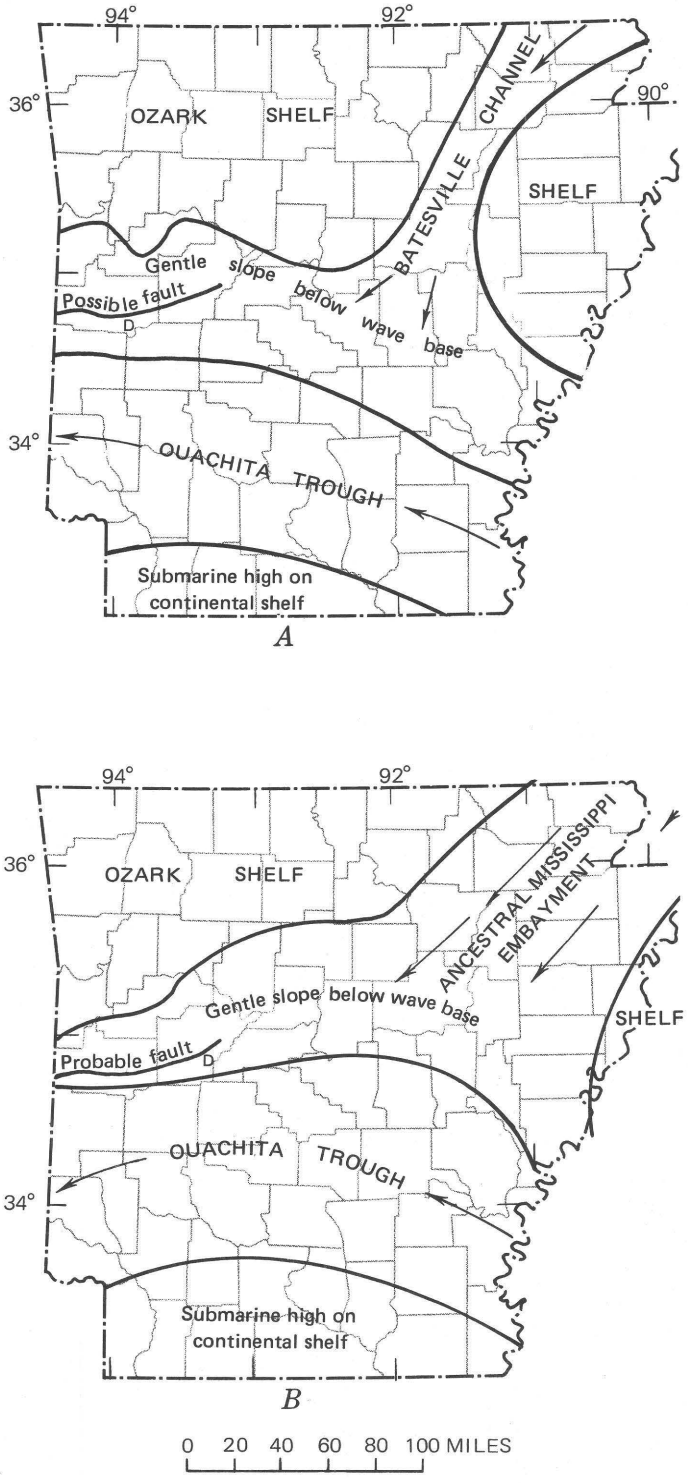


FIGURE 25. — Paleogeographic and tectonic features of Arkansas during the Mississippian Period: *A*, during intervals A, B, and C; *B*, during interval D.

INTERVAL B

FORMATIONS INCLUDED

In the Ozark region of northwestern and north-central Arkansas, interval B includes the part of the Boone Formation that is of Osage age (pl. 15, col. 69). This includes the upper two-thirds of the St. Joe Member and most of the overlying siliceous facies of the Boone up to but not including the beds of oolitic limestone or the uppermost part of the siliceous facies. The upper 10 percent of the Boone is of Meramec age and is assigned to interval C. Where the boundary between rocks of Kinderhook age and rocks of Osage age is indefinite in the St. Joe Member, intervals A and B are combined on plate 4-A. At these places, as much as 29 feet of rocks of Kinderhook age may be included with interval B.

The Boone Formation covers much of the Ozark region and the northern part of the Arkansas Valley, and it extends eastward beneath Mesozoic and Tertiary rocks of the Mississippi embayment. It thins southward, apparently by grading into rocks here considered to be the Moorefield Formation. Thus, some of the lower part of the Moorefield of the northeastern Ozark region and northern Arkansas Valley is included in interval B.

In the Ouachita Mountains area of Arkansas, interval B is composed of the upper division of the Arkansas Novaculite, which is of Kinderhook and Osage age (Hass, 1951). In the southern part of the Arkansas Valley and along some of the adjacent northern limb of the Broken Bow–Benton uplift (fig. 22), neither the Boone Formation nor the upper division of the Arkansas Novaculite has been identified. In that area a small part of the thick Stanley Shale of Mississippian age (pl. 9-E, secs. *F–F'*, *G–G'*; pl. 15, col. 71) is considered to be the equivalent of the Boone Formation and of the upper division of the Arkansas Novaculite and is arbitrarily assigned to interval B.

UPPER BOUNDARY OF INTERVAL B

The boundary between intervals B and C is selected locally on the basis of contrasting lithology, but over most of the region it is selected arbitrarily within a uniform lithologic sequence. Few paleontologic data are available as a guide.

In the central and northwestern Ozark region of Arkansas, layers of oolitic limestone commonly occur 40–60 feet below the top of complete sequences of the Boone Formation. These beds of limestone are herein correlated with the Short Creek Oolite Member of the Boone Formation of southeastern Kansas and northeastern Oklahoma (McKnight and Fischer, 1970, p. 37–40). The upper part of the Boone Formation — that above the oolite beds, and, locally, medium-crystalline limestone beds within the oolite beds — contains the *Marginirugus magnus* (Meek and Worthen) fauna, which is of Meramec age (Gordon, 1964, p. 12). Where no oolite is noted in a complete sequence of the Boone, all but the upper 40 feet of the formation is included in intervals A and B.

The oolite is not known to occur in outcrop or in the subsurface of the Ozark region east of Searcy County. In that area, the base of the zone of the *Marginirugus magnus* (Meek and Worthen) fauna is used locally to indicate the upper boundary of interval B, but the data are sparse. For the most part, the upper boundary of interval B is arbitrarily placed 40 feet below the top of the Boone Formation in the general area east of Searcy County and west of the Batesville channel (fig. 25*A*).

The upper part of the Boone that is considered to be of Meramec age and is assigned here to interval C, probably rests disconformably on interval B in the central and western Ozark region, as it does in northeastern Oklahoma (McKnight and Fischer, 1970, p. 39). This disconformity, resulting from either erosion or nondeposition, is extremely difficult to trace regionally. The formation as a whole shows no abrupt changes in thickness; where interval C thickens, the underlying part of the Boone assigned to interval B thins a compensating amount. The disconformity within the Boone Formation is considered to mark a lengthy period of nondeposition or erosion only in the Batesville channel area (fig. 25*A*). In west-central Independence County (fig. 22), the lower member of the Boone is 12–100 feet thick (Gordon and Kinney, 1944) and is overlain disconformably by an upper limestone member of Meramec age. A hiatus between the two members in this area appears to represent all of Keokuk (late Osage) time (Gordon and Kinney, 1944). In the Mississippi embayment of eastern Arkansas, the truncated edge of the Boone is overlapped by rocks of Late Cretaceous and Tertiary age.

In the vicinity of Hot Springs, Ark., the base of the Hot Springs Sandstone (interval C) provides a lithologic contact at the top of the upper division (interval B) of the Arkansas Novaculite. The Hot Springs Sandstone has been reported to rest unconformably on the upper division of the Arkansas Novaculite (Purdue and Miser, 1923). However, the regional relations suggest that both units are essentially conformable deepwater deposits. Elsewhere in central Arkansas, the Stanley Shale in interval C appears to rest conformably on the upper division of the Arkansas Novaculite (C. G. Stone, oral commun., 1971).

On the north side of the Ouachita Mountains west of Garland County, the upper division of the Arkansas

Novaculite apparently grades westward into the lower part of the Stanley Shale (B. R. Haley and C. G. Stone, oral commun., 1972).

In the Ouachita Mountains area, the upper boundary of interval B is placed at the top of the upper division of the Arkansas Novaculite as exposed near Hot Springs or at a projected time-plane equivalent to that horizon. The time plane is projected arbitrarily in the thick conformable sequence of shale of Mississippian age in the subsurface of the southern Arkansas Valley and in the thinner sequence of limy siltstone and shale farther north (Maher and Lantz, 1953a).

THICKNESS TRENDS

In Arkansas, interval B is thickest in the central and western Ozark region (pl. 4-A) where the thickness is consistently 300–400 feet. The interval thins westward into Oklahoma, southward into the Arkansas Valley, and eastward into the area of the Batesville channel (fig. 25*A*). East of the area of the Batesville channel in northeastern Arkansas the interval thickens slightly.

In the Ouachita Mountains, interval B ranges in thickness from a featheredge to 200 feet, and in most of this area it is slightly more than 100 feet thick. The upper division of the Arkansas Novaculite thins westward toward Oklahoma and has not been recognized in outcrops in the northwestern part of the Ouachita Mountains. In the Arkansas Valley the amount of the shale sequence arbitrarily assigned to interval B is less than 200 feet.

LITHOFACIES TRENDS

Limestone, siliceous limestone, limy chert, and chert compose interval B in the Ozark region of Arkansas (pl. 4-B). Probably about 50 percent of the total sequence is silica, and about 50 percent is calcium carbonate. Both the limestone and the chert are fossiliferous. The sequence apparently grades southward into limy siltstone and then into shale. The shale facies is present in the Ozark region in the Batesville channel.

In the Ouachita Mountains area, the upper division of the Arkansas Novaculite is novaculite, novaculite conglomerate, shale, and sandstone. Novaculite makes up the entire interval on the south flank of the mountains in the vicinity of Hot Springs. There, the interval is more than 90 percent silica, and some beds are 99 percent silica. Much of the novaculite is finely laminated parallel to the bedding. Some of it contains grains of sand, and some of it is interbedded with sandstone. North and west of Hot Springs the novaculite facies thins and probably grades into dark-gray shale in the lower part of the Stanley Shale.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Interval B in Arkansas was deposited under normal open-marine conditions (pl. 11, fig. 2). Calcium carbonate and intimately associated silica accumulated in shallow water in the Ozark region; novaculite, sand, and clay accumulated in deeper water in the Ouachita Mountains area; clay and silt were deposited on a gentle slope between the two areas and in water of moderate depth in the axial portion of the Batesville channel (fig. 25*A*).

Early in interval B time, crinoid debris that makes up the bulk of the St. Joe Member of the Boone Formation accumulated in a sea that covered the central and northwestern part of a shelf in the Ozark region of Arkansas (fig. 25*A*). Little of the detritus was swept either southward or eastward toward the edge of the shelf, and little of it was silicified. Probably, currents that moved northwestward across the shelf helped to nourish the crinoid "forests" and perhaps tended to inhibit silicification of the fine-grained calcite matrix that accumulated with the columnals. A change in the regimen of the sea, such as, perhaps, a slight restriction of the northwestward-moving currents, stopped the prolific growth of crinoids, and beginning with this change more than 50 percent of the subsequently deposited calcium carbonate was converted into chert or siliceous limestone.

By the end of St. Joe time, growth of crinoids had become much inhibited and most of the newly deposited fine-grained calcium carbonate was silicified, especially along the outer part of the shelf. Brachiopods, bryozoans, and crinoids continued to live on the shelf, presumably in shallow water, but generally only in moderate numbers. Fossiliferous beds of limestone were silicified, but the coarsely crystalline calcite in the fossils, especially in crinoid columnals, generally was not affected (McKnight, 1935, p. 65–66). Conditions favorable for silicification extended basinward a few miles beyond the zero edge of the St. Joe, but the thickest part of the siliceous facies accumulated in the central shelf area where the St. Joe Member is also thickest (pl. 9-E, secs. *F–F'*, *H–H'*).

Silt and clay in the Batesville channel (fig. 25*A*) had a northeastern source (pl. 10, fig. 2). This area evidently was along the seaward extension of a stream system that headed far to the north. Currents in this area carried terrigenous detritus southward. Minor amounts of silt and clay accumulated on a broad submarine slope that separated the shelf from the deeper water Ouachita Mountains area to the south where novaculite was being deposited.

The Arkansas Novaculite was deposited in a deep-

water environment in a submarine trough, called the Ouachita trough, that trended east-west across central Arkansas and extended an unknown distance southward, perhaps as far as northern Louisiana (Griswold, 1892; Miser and Purdue, 1929; Park and Croneis, 1969; McBride and Thomson, 1970; and C. G. Stone and B. R. Haley, oral commun., 1971). Material deposited in the trough was largely transported into the area by turbidity currents that probably flowed westward down the trough, but some was generated there biologically. Material brought into the area either was nearly pure silica or was capable of being converted, under the existing conditions, into nearly pure silica.

Most carbonate particles probably were carried by density currents from a distant shelf into the Ouachita trough and there were converted into novaculite. The fine carbonate particles may have been at least partly silicified as they were being transported, which could explain the textural difference between novaculite and chert. The distinctive beds of conodont-bearing novaculite conglomerate, consisting of angular and rounded granules of novaculite in a translucent silica matrix, may be made up of detritus from limestone flour that accumulated in thin layers and was silicified on the flank of the trough and later was detached, fragmented, and redeposited deeper in the trough.

One possible source area for carbonate particles may have been broad, wave- and current-swept areas of northern Arkansas and other States to the north and northeast. If the source of the carbonate that altered to novaculite was this area, it could have furnished little or no quartz sand; thus, detrital quartz grains associated within the Arkansas Novaculite probably came from a different area, perhaps a local source to the southeast.

PALEOTECTONIC IMPLICATIONS AND PALEOGRAPHIC INTERPRETATIONS

During early interval B time, a clear sea covered all Arkansas (pl. 12, fig. 2). The rate of deposition in the shelf area in the northern part of the State tended to be fastest where the water was shallow, and a shallow-water shelf facies became more sharply delineated, not by differential subsidence of a deepwater basin to the south but by upbuilding of a limestone platform in the shallow-water areas.

Tectonism was slight in Arkansas during interval B time; a small amount of subsidence is suggested by sediment thicknesses in both the trough and shelf areas (pl. 10, fig. 2). At about the end of interval B time, the central Ozark region of northern Arkansas was uplifted slightly, or sea level dropped slightly, and broad areas were probably exposed at low tide.

INTERVAL C

FORMATIONS INCLUDED

The upper part of the Boone Formation, about 40 feet of rock extending from the base of the oolite beds to the top of the formation, makes up interval C in northwestern Arkansas. The upper member of the Boone (Gordon and Kinney, 1944), the upper part of the Moorefield Formation (Adams, 1904, p. 26), and the overlying Ruddell Shale (Gordon, 1944) compose interval C in north-central Arkansas (pl. 15, cols. 69, 70). All three formations are present in sequence in Independence County west of the Batesville channel. In the Arkansas Valley, all these formations grade southward into the lower — but not lowest — part of a conformable unnamed sequence of dark-gray shale that probably is contiguous in part with the Stanley Shale. This unit, in addition to interval C, includes older and younger intervals (pl. 9-E, secs. *F–F'*, *G–G'*). In the central Ouachita Mountains area, some of the lower part of the Stanley Shale is of Meramec age (Hass, 1950), and an arbitrary amount, about 500 feet, is included in interval C, along with the Hot Springs Sandstone which is present locally beneath the Stanley. Where neither the Hot Springs Sandstone nor the upper division of the Arkansas Novaculite is present, a few feet of the lowermost Stanley is herein assigned to interval B.

STRATIGRAPHIC RELATIONS

The stratigraphic relation of the Boone Formation to the Moorefield and the Ruddell Shale is controversial. The Boone thins where the overlying Moorefield and Ruddell thicken (Ark. loc. 174; Lantz, 1950). Gordon and Kinney (1944) reported that an obscure unconformity separates the Boone from the Moorefield. Garner (1967, p. 1237) and Hopkins (1893) used a higher stratigraphic horizon for the top of the Boone than did Gordon and Kinney (1944). Garner (1967, p. 1237) stated that the Boone-Moorefield contact is "generally abupt but not obviously unconformable."

Gordon stated (1964, p. 84), "Exhaustive search has * * * [not disclosed] * * * a single cephalopod in the lower 2/3 of the Moorefield formation"; and (Gordon, 1964, p. 12), "With the exception of the St. Joe limestone member, no cephalopods are known in the Boone formation of Arkansas." Therefore, a hiatus between the Boone Formation and the overlying Moorefield Formation cannot be shown on the basis of cephalopods.

As the Boone is a clear-water deposit and the Moorefield is a muddy-water deposit, their faunas are quite different and are not easily assessed as to relative age. Certainly, part of the Moorefield is younger than at least part of the Boone because it fills in across thin se-

quences and around thicker sequences of the Boone. The assumption is made herein that the clear-water fauna of the Boone was killed by gradual but irreversible encroachment of the muddy Moorefield deposits, resulting in little if any interfingering of the two facies.

According to previous authors (Garner, 1967, fig. 1; Haley and Hendricks, 1971, p. *A8–A9*), and on the basis of fieldwork by the present author, the Moorefield is equivalent to the siliceous carbonate facies of the Boone, and the Ruddell is an offlap deposit that accumulated during the time that the surface of the Boone Formation of the central Ozark region was at about sea level. According to this interpretation, the unconformity at the top of the Boone Formation in the central Ozark region is more restricted than is generally believed; it dies out eastward toward the Batesville channel (Gordon and Kinney, 1944; Straczek and Kinney, 1950) and southward toward the Arkansas Valley.

UPPER BOUNDARY OF INTERVAL C

In the Ouachita Mountains the upper boundary of interval C is placed about 500 feet above the base of the Stanley Shale at an arbitrary horizon within a generally conformable sequence. In the southern part of the Arkansas Valley, the upper boundary is selected arbitrarily within an unnamed dark-gray shale sequence. In the eastern Ozark region the contact is conformable at the base of the Batesville Sandstone. In the north-central and northwestern Ozark region, the contact is a disconformity at the base of the Batesville Sandstone or its equivalent and is placed at the top of the Boone Formation.

The post-Boone unconformity that separates intervals C and D in the northern Ozark region apparently dies out southward in the subsurface in the southern part of the area and definitely dies out eastward along the outcrop. The easternmost reported occurrence of a basal interval D Batesville breccia and conglomerate of Boone origin is in northwestern Searcy County (McKnight, 1935, p. 77) and in western Searcy County (Glick and Frezon, 1965, p. 2).

In north-central and northwestern Arkansas (pl. 15, col. 69), the basal part (a few inches to as much as 50 ft) of the Hindsville Limestone Member of the Batesville Sandstone contains granules, pebbles, and cobbles of chert (fig. 26) that, in the opinion of some authors, were derived locally from the underlying Boone Formation (Garner, 1967, p. 1239–1241). These chert fragments are entirely decalcified — no limestone fragments are present (Garner, 1967, p. 1240). Yet, the uppermost beds of the Boone Formation in much of the area are unweathered limestone. Furthermore, the total thickness of the Boone Formation varies no more than about 40 feet throughout an area of at least 1,000 square miles in and around Boone and Newton Counties. Proposals that the post-Boone conglomerate developed locally by chemical weathering (Garner, 1967, p. 1241) and that a post-Boone "erosional relief approaching 100 feet" (Garner, 1967, p. 1238) developed in part of northwest Arkansas may be questioned. The post-Boone conglomerate does not necessarily mark a long time break in the central Ozark region.

THICKNESS TRENDS

Interval C is less than 100 feet thick throughout the central and western Ozark region (pl. 5-A). The thickness increases southward in the Arkansas Valley to as much as 200 feet and eastward into the Batesville channel region to more than 400 feet. From there it thins eastward, apparently in part depositionally and in part as a result of pre-Cretaceous erosion. An arbitrary thickness of about 500 feet is assigned to interval C in the Ouachita Mountains area.

LITHOFACIES TRENDS

In the central and western Ozark region, the lower few inches to few feet of interval C commonly is oolitic limestone that is nearly 100 percent calcium carbonate. The overlying 40 feet or so of the interval is fossiliferous siliceous limestone that locally may be no more than 50 percent calcium carbonate even where it is unweathered. The upper part of the siliceous facies of the Boone locally is oolitic, but the oolites are larger and much darker than the oolites of the *Short Creek Member* at the base of the interval. At the surface, the Boone weathers to form a regolith composed largely of porous silica rubble. The carbonate rocks thin eastward and southward to a featheredge beneath claystone and calcareous siltstone consisting of cryptocrystalline silica grains in a calcareous matrix along the northern border of the Arkansas Valley.

In the Arkansas Valley area and on the north side of the Ouachita Mountains, interval C is comprised largely of shale and limy siltstone. In the Batesville channel, interval C consists of a few beds of clayey limestone and green shale, along with much dark-gray shale; this sequence grades northwestward into bedded chert.

In the eastern part of the southern flank of the Ouachita Mountains, the Hot Springs Sandstone, locally containing beds of novaculite conglomerate, constitutes the largest single unit of coarse clastics in interval C. The part of the Stanley Shale assigned to interval C contains a larger total volume of quartz sand

than does the Hot Springs, but in the form of many thin beds of sandstone separated by shale. Important barite deposits are found on the south flank of the Ouachita Mountains in the part of the Stanley Shale assigned to interval C at or slightly above the top of the Hot Springs Sandstone. The barite deposits, however, are too small to affect the regional lithofacies pattern on plate 5-B. The lower part of the Stanley on the northern flank of the Ouachita Mountains also contains traces and thin beds of barite (C. G. Stone, oral commun., 1971).

In central and southwesern Polk County, sandy shale in the Stanley is interbedded with beds of tuff (Miser and Purdue, 1929, p. 62). The Hatton Tuff Lentil, the lowest and thickest tuff bed, is about 140 feet above the base of the Stanley Shale at Hatton and perhaps as much as 400 feet above the base within about 10 miles southwest of Hatton (Miser and Purdue, 1929, p. 63). This tuff lentil is as much as 90 feet thick locally.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Two sources (pl. 10, fig. 3) furnished detritus to Arkansas during interval C time. Fine detritus — clay at first but later a mixture of clay, silt, and a trace of fine sand — was carried from the Illinois basin into northeastern Arkansas by way of the Batesville channel (fig. 25*A*). Sand, clay, and novaculite granules and pebbles came into the Ouachita Mountains area of central Arkansas from an eastern or southeastern source early in interval C time. Later, clay, sand, and volcanic tuff came into the Ouachita Mountains area, probably from the same general direction.

At about the end of Osage (interval B) time, the sea that covered the Ozark region of north-central and northwestern Arkansas was very shallow. The sea bottom was uneven, cut by broad shallow channels that connected local depressions. Some of the area may have been above low-tide level. Deposition in parts of the area ceased for a long enough period of time to allow an appreciable change in the faunas by the time deposition resumed.

The sea became gradually deeper in early Meramec (interval C) time. Cool marine water — supplied by upwelling currents from the south — probably was gradually warmed in the shallow-water areas and became supersaturated with calcium carbonate. The excess calcium carbonate was deposited as oolites in higher energy zones and these oolites accumulated in a semicontinuous layer that was thickest where the oolites filled shallow depressions.

As soon as the water across the shelf attained a sufficient depth, non-oolitic siliceous limestone began to accumulate as it did during most of interval B time. Where the accumulation of carbonate debris was most rapid, little of the limestone was silicified. Thicker deposits of chert-free glauconitic carbonate — now quarried extensively for agricultural limestone and dimension stone — accumulated in Independence County along the right (west) bank of the Batesville channel area. A *Marginirugus magnus* (Meek and Worthen) fauna, including abundant crinoids, probably lived at most places in the shallow sea and flourished locally.

Away from the central part of the Ozark region where the water was deeper and the bottom currents were weak, the influx of clay was appreciable. In these areas cherty limestone of interval B grades upward into siliceous slightly clayey limestone of lower interval C which in turn grades upward into dark-gray limy shale of upper interval C. Silt-sized particles of cryptocrystalline silica, perhaps formed in situ but probably in part swept from shallow to deeper water areas, accumulated along the north side of the Arkansas Valley region to form silty limestone or limy siltstone incorporated in the Moorefield Formation.

By about middle Meramec (interval C) time the sea became shallower, and much of north-central and northwestern Arkansas became a land area that stood slightly above sea level (pl. 11, fig. 3). Only a minor drop in sea level was required to halt deposition and perhaps to expose a vast area. The upper few feet of the previously deposited cherty limestone may have been decalcified during this period of subaerial exposure, but the land area was not appreciably channeled. An embryo karst topography may have developed locally. The area where the Boone is thickest was probably the first to be exposed to weathering and the last to be covered by the readvancing early Chester sea. Truncation was slight, probably because the area was barely above sea level.

During the time that the central Ozark region of northern Arkansas was exposed to subaerial chemical weathering, a layer of dark-gray clay (Ruddell Shale) was deposited on the upper Boone and outward across the impure limestone of the Moorefield Formation in the surrounding part of the Ozark region that was still covered by the sea.

In the Ouachita Mountains area the sea bottom was much below wave base throughout interval C time. The Hot Springs Sandstone and its local basal conglomerate of angular to rounded novaculite granules and pebbles are considered to be deepwater deposits. The type of transport that delivered this gravel and sand to the trough has not yet been identified except that gravity rather than wave energy is thought to have supplied the impetus.

Clay and fine-grained sand of the lower part of the

tanley Shale probably was carried into the Ouachita Mountains area by turbidity currents from a outheastern source. Economically important beds of arite in the Stanley are here considered to be primary eposits that relate in some way to the sea-bottom opography of the time of deposition. Zimmermann 1965) also favored a primary origin; however, Scull 1958, 1959) considered the deposits to be secondary nd related the barite to igneous intrusions of Late Creaceous age. The Hatton Tuff Lentil and other beds of uff in interval C were derived from an unknown olcanic source. The tuff beds probably are turbidites nd a southeastern source is suspected. However, indblown volcanic debris from the south or southwest ould have accumulated on a submarine slope prior to liding or flowing laterally into the axial portion of the rough. In the Ouachita Mountains area, Meramec (interval C) time ended with no appreciable change in the nvironment or rate of deposition.

PALEOTECTONIC IMPLICATIONS AND PALEOGEOGRAPHIC INTERPRETATIONS

Most of Arkansas was covered by a sea during nearly ll of interval C time. The sea was shallow in the northrn half of the State and deeper in the southern half (pl. 2, fig. 3). The sum of tectonic movement in interval C me in the Ozark region must have been slight subsience, as shown on plate 10 (fig. 3), and this area reained a shelf.

In the deeper water of the Ouachita Mountains area, ne rate of deposition increased during interval C time ompared to interval B time, and subsidence probably ept pace. Along the northern boundary of the deeper ater area, subsidence of the Ouachita basin or trough elative to the shelf may have taken place along a norial fault that is thought to have become active during ne Mississippian Period.

An abrupt increase in the volume of sand delivered and deposited in the Ouachita trough early in leramec time and an influx of volcanic debris later in leramec time indicate tectonic activity in the source rea — perhaps in southern Alabama or southern lississippi.

INTERVAL D

FORMATIONS INCLUDED

In ascending order the Batesville Sandstone, Fayetville Shale, and Pitkin Limestone, all of Chester age, ake up interval D in the northern Ozark region (pl. 5, cols. 69, 70). Southward, in the southern Ozark gion and in the Arkansas Valley, all these units grade to a thick unnamed sequence of dark-gray shale.

In the Ouachita Mountains area, the upper 5,500 feet of the Stanley Shale is assigned to interval D (pl. 15, col. 71). In that area, faunal control for the base and top of rocks of Chester age has not been definitely established.

STRATIGRAPHIC RELATIONS

Interval D rests unconformably on interval C in the north-central and northwestern part of the Ozark region of Arkansas. Elsewhere in Arkansas, interval D is conformable on interval C.

The formations of interval D in the Ozark region — the Batesville, Fayetteville, and Pitkin — are conformable. In Washington and Madison Counties, however, conglomerates are present in a few outcrops at the base of the Wedington Sandstone Member of the Fayetteville Shale and at the base of a thin bed of limestone that lies between the Wedington and the Pitkin (Taylor, 1964). Across northwestern Arkansas, the Pitkin generally rests upon the upper shale of the Fayetteville but locally rests upon a thin finger of the Wedington Sandstone Member (Purdue and Miser, 1916, p. 14). These relations do not indicate regional stratigraphic breaks.

In western Arkansas, shale in the lower Fayetteville below the Wedington Sandstone Member thins northward from more than 100 feet to about 40 feet. The Batesville, represented largely by the Hindsville Limestone Member, thickens northward from less than 10 feet to more than 80 feet in the same area. An important change in the cephalopod fauna vertically "from the assemblage of the *Goniatites granosus* zone to that of the *Tumulites varians* zone occurs at the contact between the Batesville Sandstone and the Fayetteville Shale at Marshall, Searcy County," but, in contrast, "at Round Mountain, Independence County, fossils of the *G. granosus* zone persist over 55 feet stratigraphically upward into the Fayetteville Shale" (Gordon, 1964, p. 85). These relations suggest that the Batesville is partly a facies equivalent of the Fayetteville.

Facies changes also complicate the relations stratigraphically higher in interval D. In northern Washington County, the Wedington replaces all of the middle and upper shale and limestone of the Fayetteville, probably all of the lower limestone of the Pitkin, and possibly all of the upper limestone of the Pitkin. These units are all present farther to the east. In Washington County a shale facies everywhere separates the limestone facies from the gradually westward thickening Wedington Sandstone Member. Locally, along the narrow zone in which the shelf limestone grades basinward into shale, the limy shale is phosphatic, containing as much as 5 percent P_2O_5 in unpicked well cuttings.

The Fayetteville in north-central Arkansas is at

least as thick as the combined Fayetteville and Pitkin to the south (Purdue and Miser, 1916, p. 13). In northern Washington County, the Fayetteville Shale is at least as thick as the Fayetteville and Pitkin in adjacent areas to the east where the Pitkin is present. No Pitkin is known to overlie the Fayetteville where the Wedington Member is especially thick (Adams and Ulrich, 1905). These compensating thickness changes suggest that the Batesville, Fayetteville, and Pitkin in part are facies equivalents of one another.

In the central and eastern Ozark region, from Searcy County to Independence County, the uppermost unit of the Pitkin (upper interval D) is an eastward-thickening wedge of medium-gray limy fossiliferous shale that contains thin beds of crinoidal limestone (Gordon, 1964, p. 31, 32; and Glick and Frezon, 1965, p. 2). This unit, as much as 80 feet thick, probably has no correlative in northwestern Arkansas or in part of north-central Arkansas (pl. 15, col. 70).

UPPER BOUNDARY OF INTERVAL D

Interval D in Arkansas is overlain at most places by the Pennsylvanian System, with unconformable contact in the northern part of the State and a conformable contact in the central and southern parts. The truncated edge of the Mississippian System is overlapped by rocks of Cretaceous and Tertiary age in the Gulf Coastal Plain and in the Mississippi embayment (pl. 8; pl. 9-E, sec. *H–H'*).

The Pitkin Limestone is absent by nondeposition or was removed by pre-Pennsylvanian erosion from the northwestern and north-central Ozark region (pl. 9-E, sec. *F–F'*). Beyond the northern limit of the Pitkin, silty sandstone and claystone of the lower part of the Hale Formation of Pennsylvanian age rests unconformably on the Fayetteville Shale. Locally, as in northwestern Newton County (Ark. loc. 104A), a conglomerate of Early Pennsylvanian age containing pebbles of the Pitkin rests on a truncated sequence of Fayetteville Shale.

In north-central Arkansas, a distinctive basal sandstone of the Hale Formation rests on the upper shale member of the Pitkin, furnishing an easily identifiable upper boundary of interval D in most of that area. Locally, in Searcy and Van Buren Counties, channel-fill deposits of Pennsylvanian age, including coal, phosphorite, and carbonaceous shale, rest on the Pitkin. In southern Independence County, the upper boundary of interval D appears to be especially uneven; post-Pitkin erosion in the ancestral Mississippi embayment (fig. 25*B*) selectively removed several tens of feet of the Pitkin. The overlying Pennsylvanian rocks are claystone interbedded with thin layers of sandstone containing quartz granules and pebbles.

In the subsurface in the southern Ozark region an in the Arkansas Valley, limestone of the Pitkin (inter val D) and most of the sandstone of the overlying Hal (Pennsylvanian) grade southward into shale. The up per boundary of interval D becomes increasingl difficult to identify as it is traced southward.

Deposition was continuous from Mississippian int Pennsylvanian time in the Ouachita Mountains are and in the southern and eastern parts of the Arkansa Valley; it was nearly continuous in the northern part o the Arkansas Valley. In the Ouachita Mountains are the upper boundary of interval D is arbitrarily drawn a the top of the Stanley Shale and at the base of th Jackfork Sandstone, a boundary that is not necessaril at the exact top of the Chester Series and does no necessarily delimit rocks of the same age everywhere The upper part of the Stanley Shale is largely shale, an the overlying Jackfork Sandstone is largely sandstone Sandstone in the upper Stanley tends to have more cla and feldspar than does sandstone in the overlyin Jackfork, providing a basis for selecting the top of in terval D in areas where the upper Stanley is especiall sandy.

White (1937b) concluded, on the basis of fossil plant from the Stanley and from the overlying Jackfor Sandstone, that the systemic boundary is within th Stanley. However, some of his "Stanley" fossils ma have come from beds that are now mapped as Jackfor (C. G. Stone, oral commun., 1972). Other geologist have placed the systemic boundary above the Jackfor (Miser and Hendricks, 1960). In this report th systemic boundary is projected southward, on the basi of regional sedimentary patterns, from its known posi tion at the base of the type section of the Morrow Serie in northwestern Arkansas to the Stanley-Jackfork con tact in the Ouachita Mountains area.

THICKNESS TRENDS

Rocks of Chester age in Arkansas thicken southwar from the outcrop in the Ozark region and from th subcrop in the Mississippi embayment into th Ouachita Mountains area (pl. 6-A). Probably the thin nest depositional sequence of interval D in Arkansas i along the Oklahoma border north of Fort Smith wher less than 200 feet of rock of Chester age accumulate (Ark. loc. 166A). Interval D is less than 200 feet thic under Pennsylvanian cover elsewhere in northern Ar kansas (Ark. loc. 104A), but only as a result of pr Pennsylvanian erosion. In the Ouachita Mountains, th total thickness of the Mississippian System is thoug to be at least 6,000 feet (Ark. locs. 255B, 306B) of whic more than 5,000 feet is assigned to interval D.

Drilling indicates that interval D is more than 1,50 feet thick at two widely spaced places (Ark. locs. 22

291) in the Mississippi embayment of east-central Arkansas. In the embayment, the interval thickens generally southward toward the southern part of the Ouachita Mountains at a fairly uniform rate of 50–100 feet per mile.

The rate of southward thickening of interval D along the western border of Arkansas is quite low as far south as the southernmost control point near Fort Smith (Ark. loc. 246). From there southward to the Ouachita Mountains, such a high rate of thickening of the interval is required that some faulting during deposition is suspected.

LITHOFACIES TRENDS

The regional lithofacies pattern (pl. 6-B) shows an east-trending linear lens of mostly bioclastic carbonate rock flanked by a shale facies to the south and a limestone-shale facies to the north.

Measured by volume, shale is by far the most abundant lithic element in interval D, sandstone is second in abundance, and limestone is third. Calcium carbonate, in the form of fossil fragments, oolites, and fine-grained lime mud, was deposited in parts of the Ozark region. Sand was deposited in the Ozark and Ouachita Mountains regions. Clay was deposited throughout the area.

Two sandstone wedges of interval D, the Batesville and the Wedington, are present in northern Arkansas and pinch out southward. The Batesville is at least 75 feet thick along the outcrop in Independence County (Ark. locs. 357, 128; Gordon and Kinney, 1944) and 100 feet thick in northern Boone and Carroll Counties (Purdue and Miser, 1916, p. 12). The greatest reported thickness, "about 225 feet near Batesville" (Croneis, 1930, p. 59), probably includes parts of adjacent formations. The Wedington is as much as 150 feet thick along the outcrop in northwestern Washington County (Adams and Ulrich, 1905, p. 4). These sandy units affect the regional lithofacies pattern (pl. 6-B) largely in northern outliers of interval D where the wedges are thickest.

In the Ouachita Mountains area the upper 5,000 feet or so of the Stanley (interval D) is shale and interbedded limestone. The sandstone, at least 10 percent of the unit, makes up most of the sand deposited in Arkansas during interval D time. Sandstone of the Stanley is shown graphically on the cross sections of plate 9-E (secs. *F–F'*, *G–G'*).

In the Arkansas Valley and locally in the southern Ozark region, interval D is nearly all dark-gray shale. Farther north, the shale thins and about half of it — mostly the upper part of the interval — grades abruptly into limestone (Caplan, 1957, p. 4). The limestone and shale are fossiliferous throughout the Ozark region and in the northern part of the Arkansas Valley. Some of the limestone is oolitic, much of it probably is pelletal. In much of the Ozark region, shale in the lower part of the interval (lower Fayetteville) contains limy concretions.

The maximum development of limestone in interval D in Arkansas, both in thickness and number of discrete units, is in Stone County and southern Independence County where the Fayetteville locally contains a Pitkin-like middle limestone member (pl. 15, col. 69) as much as 60 feet thick, in addition to beds of black fine-grained limestone in both the lower and upper dark-gray to black shale members (Gordon and Kinney, 1944). The overlying Pitkin is fossiliferous limestone of which individual beds are fine to coarse grained and oolitic, conglomeratic, arenaceous, encrinal, or cherty (Gordon and Kinney, 1944).

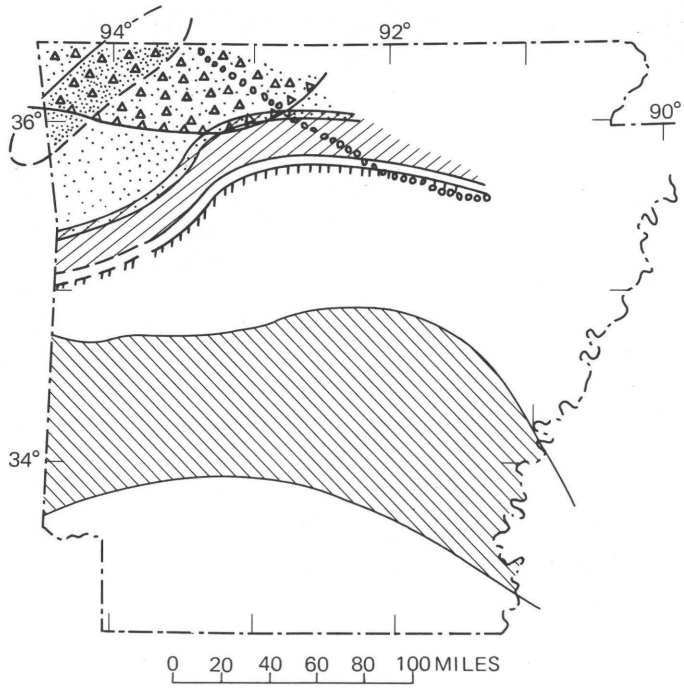
In the northwestern and northern Ozark region, the Pitkin, where present, is massive limestone resting on dark-gray shale of the Fayetteville or, locally, on the uppermost bed of the Wedington Sandstone Member (Purdue and Miser, 1916, p. 14; McKnight, 1935, p. 79; Easton, 1942, p. 38–39). In a 20-mile-wide east-trending zone at its southern limit, the Pitkin rests on the upper limestone unit of the Fayetteville (fig. 26). Along that zone, beneath the massive-bedded crystalline limestone of the Pitkin, the upper Fayetteville consists of beds of dark-gray clayey siliceous fine-grained limestone (or limy siliceous shale) spaced at about 1-foot intervals between slightly thicker layers of dark-gray limy shale (Easton, 1942, fig. 5). Spreng (1967, p. 806) reported a few chonetid and linguloid brachiopods in the shale and productid and orthotetid brachiopods in the limestone.

The Hindsville Member of the Batesville Sandstone affects the regional lithofacies pattern beyond the northwestern limit of the Fayetteville and Pitkin limestone units. In northeastern Benton County the Hindsville is 90 feet thick and consists of alternating beds of shale, sandstone, and limestone (Adams and Ulrich, 1905, p. 4).

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The lithofacies of interval D in Arkansas (pl. 6-B) reflect a northwest-to-southeast shelf-to-basin sequence (fig. 25*B*). Interval D in Arkansas consists of marine deposits derived from both organic and inorganic precipitation of minerals dissolved in sea water and from the erosion of detritus from distant source areas. The volume ratio was on the order of 1:200, about 100 cubic miles of limestone and other chemical precipitates to at least 20,000 cubic miles of detritus.

Southward-flowing currents (pl. 10, fig. 4) in the



EXPLANATION

Southeastern limit of the massive Pitkin deposits — A few thin beds of limestone extend southeast of this line

Upper limestone unit of the Fayetteville Shale

Wedington Sandstone Member of the Fayetteville Shale — Dots closely spaced where the thick Wedington sand barrier was deposited

Area in which decalcified chert pebbles are present in the Hindsville Member of the Batesville Sandstone

Southwestern limit of the Batesville delta terrace as determined by the thick part of the sandstone facies of the Batesville

Deepwater deposits in Ouachita trough

FIGURE 26. — Paleogeographic and sedimentary features of Arkansas that indicate relative depth of sea during Chester (interval D) time.

Mississippi embayment area (fig. 22) carried mature sand and clay into Arkansas from a positive area beyond the Illinois basin to the northeast. A southeastern source furnished an even larger volume of detritus, partly first-cycle clay and sand from a metamorphic terrain, to the deepwater Ouachita trough. Probably much of the detritus that reached the Ouachita trough was transported by turbidity currents and gravity-induced mass movement.

At the beginning of interval D time, the upper part of the Boone Formation (interval C) was at or near sea level in the Ozark region of northwestern Arkansas surrounded by a sea in which dark-gray clay now in the Ruddell Shale was the chief deposit. Deposition of the Chester Series began in northern Arkansas with the deepening and readvance of the sea northward. A sand delta, represented by the Batesville Sandstone, was built southwestward over the dark-gray clay in the area of the ancestral Mississippi embayment. In shallow-water areas to the west, the early Chester sea was more energetic than the late Meramec sea, and a blanket of dark-gray oolitic fossiliferous limestone (Hindsville Limestone Member of the Batesville Sandstone) was deposited at and above wave base. Cool nutrient-laden currents rose from deeper water areas to the south and spread across the shelf. Garner (1967) described reef and near-reef deposits that accumulated in the Hindsville largely southwest and west of the sand delta.

Decalcified chert debris similar to chert in the Boone is incorporated in the limestone of the Hindsville Member throughout a broad area of northwestern Arkansas (fig. 26). Part of the debris is in a thin layer of basal conglomerate, but much of it occurs as discrete pebbles in a matrix of massive-bedded limestone as much as 50 feet above the base of the Hindsville. Most of the chert gravel probably was transported into the area by westward- or southward-flowing marine currents and was spread across the sea floor at the time that the Hindsville Limestone Member was being deposited.

After at least 75 feet of deltaic sand had been deposited locally in the ancestral Mississippi embayment area, the depositional center migrated westward toward Boone County, marine currents reworked the deltaic deposits, carried clay into the basin, and spread sand westward into and eventually across the Hindsville reef deposits. About 50 feet of shale (lowermost Fayetteville) accumulated on the drowned delta in southwestern Independence County. Eventually, the Fayetteville Shale lapped onto progressively younger beds of the Hindsville Limestone Member, those in northern Benton County being the last to be covered.

The clay of the lower Fayetteville was deposited as a blanket slowly and uniformly across the shelf during the time when the sea in northern Arkansas reached its maximum depth. Apparently, wave energy was at a minimum. The water was increasingly turbid with depth, both from slight disturbance of the bottom and from new clay that was carried in suspension. A sparse and highly specialized fauna lived on the muddy sea bottom, and a free-swimming fauna, made up in part of cephalopods, lived in the relatively clear water above.

During middle Fayetteville time in a small area of Stone County and southwestern Independence County, a lens of bioclastic and oolitic limestone, the middle limestone unit of the Fayetteville, was deposited. The limestone probably indicates a localized area of higher energy water, wave action, and, perhaps, a brief period of lower sea level. The lens was later buried by clay deposits of the Fayetteville reflecting a possible rise of sea level.

At about the same time, well-sorted fine-grained sand now composing the Wedington Sandstone Member of the Fayetteville Shale was spread across the western side of the shallow-water Ozark shelf. The first layer of sand was the most extensive, reaching as far eastward as northwestern Searcy County and southeastward into northern Johnson County (fig. 26). A lowering of sea level or an increase in wave energy may have aided the transportation of the sand. The sand was carried into the area from the northeast by longshore currents. The bulk of the sand probably was first deposited as a regressive unit eastward of its present position. As the sea level rose, waves from the southeast probably transported much of the sand northwestward and deposited it in the form of a sand bar or island in shallower water. The final direction of movement was northward or northwestward, as indicated by cross-laminations (Croneis, 1930, p. 68). Some of the sand in the uppermost layers may have been windblown.

Marine animals lived in and on the sand, and plants grew on the sandbar — or plant debris floated onto it from some adjacent land area. White (1937a, p. 17) stated: "Though practically nothing is known of the internal structure of the plants, the inference may be drawn that the environment was not wholly favorable to the development of large size in growth of plants and trees and that the climate was probably marked by a seasonal deficiency of rainfall." An especially varied fauna, including the bryozoan *Archimedes*, lived in the area periodically during the accumulation of the clay and thin beds of limestone that were deposited above all but the thickest part of the Wedington.

Upwelling of cold water from the deep trough to the south was augmented by the presence of the Wedington barrier. Any surface water that the wind pushed southeastwardly away from the barrier tended to be replaced by a counterflow of cold water from the deep basin to the south. Phosphatic clay and beds of low-grade pelletal phosphorite accumulated in north-central Arkansas in an area parallel to but beyond the southeastern limit of the Wedington. Undoubtedly, the upwelling of the cold water not only increased the precipitation of phosphatic minerals but also made the area hospitable to a cool-water fauna.

Toward the end of Fayetteville time, part of the area accumulated alternating layers of dark-gray limy clay and siliceous clayey lime mud, producing the distinctive bedding in the upper limestone unit of the Fayetteville. A 20-mile-wide zone of lime mud, flanked on either side by dark-gray clay, extended across the State, barely lapping onto the outer edge of the Wedington Sandstone Member (fig. 26). Spiny productid brachiopods and a few other animals lived in this quiet marine environment.

The beginning of Pitkin time was marked across northern Arkansas by an abrupt change from the deposition of limy clay to the deposition of calcium carbonate that formed massive beds of limestone. That change resulted from a change in the circulation of sea water in the shelf area. As the area subsided, the Wedington barrier ceased to restrain the flow of surface water effectively and, therefore, upwelling decreased markedly. Relatively cool open-ocean water was circulated slowly across the broad shelf and was warmed enough to cause the release of some of its dissolved calcium carbonate. Oolites accumulated intermittently and were mixed into the debris generated by the prolific fauna that occupied the clear water of the open shelf. Reeflike masses of limestone accumulated in some areas during Pitkin time; generally, they are not readily mappable within the Pitkin biostrome and, therefore, are not well documented.

Along its shoreward edge in west-central Washington County, the Pitkin is in part crossbedded oolite; the crossbedding indicates currents from the south (Easton, 1942, p. 33). In central Newton County some of the oolitic Pitkin is saturated with hydrocarbons (Easton, 1942, p. 40).

During the latter part of interval D time, limestone deposition virtually ceased in the central Ozark shelf area and limy clay began to accumulate on the outer edge of the shelf. A thick blanket of clay filled the depressions around the shelf edge and lapped across the still-submerged eastern Ozark region to form as much as 80 feet of fossiliferous limy gray to greenish-gray claystone above a seemingly complete sequence of Pitkin Limestone. A few masses of fossiliferous limestone accumulated along the edge of the deepwater trough. Some of that limestone slid into the trough and was incorporated in the upper Stanley in such deposits as the boulder beds at Hollis in southern Perry County that consist of "slump-slide or erratic masses of limestone" (fig. 27) of Pitkin age (Stone, 1963, p. 1). Interval D time, as well as the Mississippian Period, ended with the regression of the strandline southward across the Ozark shelf. The basal sandstone of the Cane Hill Member of the Hale Formation (Pennsylvanian) was deposited across the slightly eroded top of the Pitkin.

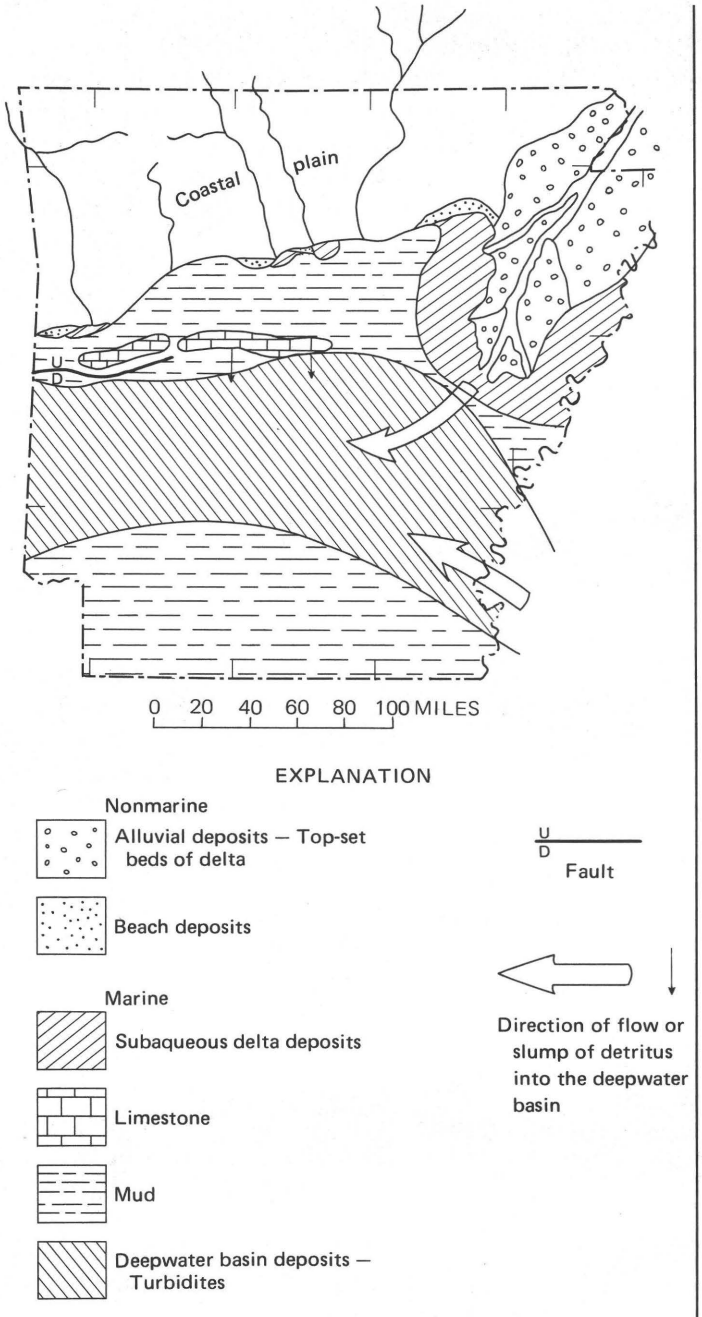


FIGURE 27. — Distribution of environments and sediment types in Arkansas at the time of the maximum withdrawal of the sea at the end of Chester time.

PALEOTECTONIC IMPLICATIONS AND PALEOGRAPHIC INTERPRETATIONS

During all of interval D time, the Ouachita trough was a well-defined subsiding structural trough that occupied the south-central part of Arkansas (pl. 10, fig. 4). The southern flank of the trough may have extended into northern Louisiana; at least, interval D appears to continue thickening southward as far as the margin of the Gulf Coastal Plain. Some volcanic debris, similar to that of interval C, was deposited in the Ouachita trough indicating the continuation of volcanic activity. To the northeast in Arkansas, the Batesville channel expanded laterally to become the ancestral Mississippi embayment, and the Ouachita trough of interval D time joined the lower end of the embayment, probably along a continuous and relatively gentle submarine slope. To the northwest, the Ouachita trough abutted the southwestern edge of the Ozark shelf at an east-trending normal fault or sharp flexure (shown as flexure on pl. 10, fig. 4) which was a forerunner of Early Pennsylvanian (Morrow) large-scale faulting.

Increased subsidence was the important structural change in Arkansas during interval D time (pl. 10, fig. 4) and more than 20,000 cubic miles of sediment accumulated in the region. Sea-level changes were a result of either epeiric or eustatic movements. The more positive part of the shelf area in north-central and northwestern Arkansas was at or slightly above sea level in early interval D time, in middle Fayetteville time, and in latest interval D time. Otherwise, all the region remained below sea level (pl. 12, fig. 4), the Ozark shelf was generally near wave base, and the Ouachita trough was continuously well below wave base.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

The depositional thickness of rocks of Mississippian age in Arkansas (pl. 7) ranges from about 400 feet in the subsurface just north of Fort Smith to at least 6,000 feet in the Ouachita Mountains. Pre-Cretaceous erosion removed the entire Mississippian System from large areas in north-central and northeastern Arkansas (Caplan, 1954, pl. 4) and from the central Ouachita Mountains and their eastward extension beneath younger rocks in the Mississippi embayment.

The Mississippian rocks thicken eastward across the Ozark region into the ancestral structural Mississippi embayment. Although thickness data are sparse, the system apparently thickens rapidly enough southward within the embayment to merge smoothly with the thick Mississippian rocks of the Ouachita Mountains area.

In western Arkansas, rocks of Mississippian age gradually thicken southward to the area about 15 miles south of Fort Smith. The sequence becomes abruptly thicker farther south along a presumed east-trending sharp flexure or normal fault that must have been active during at least the latter part of Mississippian time.

PALEOTECTONIC IMPLICATIONS

The Mississippian Period was a time of subsidence in Arkansas. Every part of the State received sediment,

only a trace of which was derived from older rocks in the State.

Although the evidence is inconclusive, probably no area in Arkansas was uplifted during the Mississippian Period: The Ozark region emerged briefly during the latter part of Meramec time and again during latest Mississippian time, perhaps as the result of regional upwarping, but more likely as the result of a minor eustatic lowering of the sea level. The sea soon returned to about its former level and deposition and subsidence continued after only slight local erosion. The northern part of the State was a structural shelf that had relatively uniform subsidence during Mississippian time, increasing somewhat irregularly in Chester time.

The Ouachita trough was a deepwater basin. The relatively small accumulation of detritus during Early Mississippian time indicates that the trough was a starved basin that probably did not subside much more than did areas farther to the north. Both subsidence and deposition increased markedly in this area in Chester time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

In Arkansas the Mississippian System is overlain by rocks of Morrow age (Early Pennsylvanian), rocks of Cretaceous age, and rocks of Tertiary age (pl. 8). In both the Ozark region and the Ouachita Mountains, Mississippian rocks are at the surface over large areas.

The Jackfork Sandstone (Pennsylvanian) rests conformably on the Stanley Shale (Mississippian) in the Ouachita Mountains. In southern Arkansas the contact is conformable, and the stratigraphic position of the systemic boundary is controversial.

In east-central Madison County, rocks of Morrow age rest unconformably on truncated Mississippian rocks as old as the middle part of Fayetteville Shale (middle Chester); the youngest Pennsylvanian rocks directly above the systemic contact are the Prairie Grove Member of the Hale Formation (middle Morrow). Thus, the unconformity at the systemic boundary in the Ozark region locally is a hiatus represented elsewhere by the upper part of the Fayetteville Shale (Chester), the overlying Pitkin Limestone (Chester), and the Cane Hill Member of the Hale Formation (earliest Pennsylvanian).

The Hale Formation, which is the oldest post-Pitkin unit in northern Arkansas, was assigned to the Pennsylvanian System by Ulrich (in Adams, 1904, p. 109–113). Conodont data indicate that "the uppermost part of the Pitkin in north-central Arkansas can be no older than the Grove Church Shale" (Lane, 1967, p. 923), the uppermost rocks of the type Chester in Illinois. Therefore, use of the unconformity at the base of the Cane Hill Member of the Hale Formation as the base of the Pennsylvanian in northern Arkansas probably is compatible with the placement of the systemic boundary in adjacent regions, and such use provides a distinctive mapping surface for the purpose of this study. However, in Newton and Searcy Counties (Snowball quadrangle), the Cane Hill is of formational rank and has been assigned a Mississippian and Pennsylvanian age (Glick and others, 1964, fig. 3). That assignment was based on the definition of the Mississippian-Pennsylvanian systemic boundary as the base of the British R_1 goniatite zone.

In the Gulf Coastal Plain of southern Arkansas and the Mississippi embayment of eastern Arkansas (fig. 22), rocks of Cretaceous age and overlapping rocks of Tertiary age rest on truncated edges of the Mississippian sequence where it extends beyond the Pennsylvanian cover. The distribution of the various units that overlie the Mississippian of southern and southeastern Arkansas, shown on plate 8, is generalized inasmuch as data are sparse in those areas.

PALEOTECTONIC IMPLICATIONS

Either gentle regional tilting or an eustatic lowering of sea level caused the withdrawal of the sea at the end of Mississippian time (fig. 27). Local uplift in the Ozark region also may have occurred. Perhaps the entire Mississippi Valley was tilted southward slightly.

The area sank slightly in earliest Pennsylvanian time sufficiently for the sea to cover most of the Ozark region. In middle Morrow time (post-Cane Hill), a drop in sea level of about 300 feet allowed streams to channel deeply into and locally through lowest Pennsylvanian deposits (Glick and others, 1964, fig. 6). The Prairie Grove Member of the Hale Formation, which fills the channels and overlaps the truncated earliest Pennsylvanian beds, locally rests on Mississippian rocks. It is the youngest Paleozoic rock deposited directly on the Mississippian of Arkansas.

Uplift, folding, and faulting took place in central Arkansas during the Late(?) Pennsylvanian Ouachita orogeny. These events were followed by a long period of erosion — about 200 million years in duration — during which Mississippian rocks became widely exposed in both the Ozark region and in the Ouachita Mountains. Subsidence during the Mesozoic allowed transgression of the sea northward. Rocks of Cretaceous and Tertiary age that were deposited overlapped the folded, faulted, and truncated Paleozoic sequence in the Gulf Coastal Plain and the Mississippi embayment.

SUMMARY[1]

Two major cycles of deposition are recorded in the Mississippian System of Arkansas. The cycles are most evident in the shallow-water deposits to the north. There, the dark-gray Chattanooga Shale and its basal sandstone (Devonian and Mississippian) were deposited in a low-energy transgressive environment and are conformably overlain by biogenic and locally oolitic limestone of the Boone Formation (intervals A, B, and C), which was deposited in a medium- to high-energy environment. Each bed of limestone that was not buried quickly was silicified. Partial withdrawal of the sea and slight local erosion ended this cycle.

The dark-gray Fayetteville Shale and its basal sandy unit (interval D), deposited in a medium-energy transgressive environment, are in turn conformably overlain by the biogenic and oolitic Pitkin Limestone (interval D), deposited in a medium- to high-energy environment. Partial withdrawal of the sea and slight local erosion ended this cycle. The rate of supply of detritus from distant sources increased many fold after the first cycle; apparently, the supply of silica in solution decreased.

An unbroken sequence of deepwater sediments was left in the Ouachita trough. Early in the Mississippian Period, clay accumulated slowly (upper part of the middle division of the Arkansas Novaculite), interrupted rarely by rapid accumulating turbidity deposits. During late Kinderhook and Osage time, shallow-water lime mud (upper division of the Arkansas Novaculite), derived from an unknown source, was spread across the flat bottom of the trough and was silicifed. Later, a large supply of clay and sand and a few layers of volcanic ash (Hot Springs Sandstone and Stanley Shale) was supplied from a southeastern source (Cline, 1966). Subsidence kept the water deep.

The Mississippian Period ended as quietly as it began in Arkansas. Deposition continued in the Ouachita trough. A transgressive beach sand of Early Pennsylvanian age (Cane Hill Member of the Hale Formation) was deposited on slightly truncated beds of Mississippian limestone and shale in the Ozark shelf area.

[1] Data from recent investigations in the subsurface of the Mississippi embayment region, including data from deep oil and gas test-holes, have modified earlier concepts of the structure of that area and have refined earlier estimates of the subcrop distribution of Mississippian units. The Batesville trough and its later and wider counterpart, the ancestral Mississippi embayment, now appear to be situated within an ancient rift zone that is outlined by geophysical anomalies as interpreted by T. G. Hildenbrand and others of the U.S. Geological Survey (written commun., 1977). The structure has been designated the Mississippi rift. As the geophysicists postulate that the rift may have been active intermittently since Precambrian time, it could have played a major role in the development of structural and depositional trends in eastern Arkansas during Mississippian time, especially in the zone along the western edge of the Mississippi embayment.

East of that postulated rift zone the shallow-water shelf facies of the Mississippian System appears to have extended farther south, in a similar pattern to that of the west side of the rift zone. However, beneath the central part of the Mississippi embayment much of the Mississippian sequence has been removed by erosion across major anticlines that parallel the rift zone and plunge gently toward the south-southwest. The two major anticlines of the area, an unnamed one in eastern Arkansas and the Senatobia anticline of northern Mississippi show block-faulting and appear to be separated by block-faulted syncline that trends roughly north-northeast through the vicinity of Memphis, Tenn.

The central Arkansas Valley syncline or graben extends eastward beneath the Mississippi embayment sediments. In that area it now appears to be sharply defined by faults along which depositional units are displaced several thousand feet, the northern boundary fault abruptly terminating the anticlines discussed above. The southern boundary fault is in the eastern extension of the faulted zone that marks the northern limit of the Ouachita Mountains. Drilling reports suggest that the strata at depth between those two fault zones are tilted toward the north in the Mississippi embayment region.

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Oklahoma

By SHERWOOD E. FREZON *and* LOUISE JORDAN

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–I

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

OKLAHOMA

By SHERWOOD E. FREZON AND LOUISE JORDAN[1]

ABSTRACT

In most of the Oklahoma region the Mississippian rocks of this report are underlain by the Chattanooga Shale or the Woodford Chert (or Shale) of Devonian and Mississippian age. Where these rocks are locally absent beneath the Mississippian, rock units as young as Silurian and Devonian age and as old as Precambrian age are present beneath the Mississippian.

The Mississippian System in this region includes rocks assigned to four intervals that correspond approximately to the four provincial series of the system in the midcontinent region. The series and their approximate interval equivalents are Kinderhook — interval A, Osage — interval B, Meramec — interval C, and Chester — interval D.

Tectonic activity in the region was probably at a minimum at the start of Mississippian time. During this time the upper part of the Chattanooga-Woodford sequence and equivalent rocks were being deposited in the region, and evidence of stratigraphic breaks indicating extensive uplift are not found in the sequence.

At the end of Chattanooga time minor uplift probably occurred over the region, but the subsequent erosion was not extensive enough to remove the Chattanooga. Carbonates were deposited in the new depositional environments established at the start of interval A time, and carbonate deposition continued, probably without appreciable interruption, to the end of interval B.

At the start of interval C time subsidence started along the axis of the Anadarko-Ouachita trough, an arcuate structure extending across the southern part of the region. Downwarping continued for the remainder of Mississippian time and well into Pennsylvanian time. During intervals C and D time great thicknesses of detrital rocks were deposited in the trough. On the shelf area north of the trough a sequence of mostly marine carbonates and detrital rocks was deposited.

The greatest amount of erosion at the end of interval D time probably occurred on the shelf areas in the northern part of the region. In the trough areas deposition was virtually continuous from Mississippian into Pennsylvanian time.

REGION DEFINED

The region reported in this chapter consists of an area of about 64,000 square miles that includes most of the State of Oklahoma (fig. 28). Two areas of the State are not included in this report; the three counties that compose the westward-extending panhandle in the northwestern part of the State, and parts of four counties in the southwest part of the State, south of the Wichita uplift, are included in the region described by Mapel and others (chap. J).

The Mississippian System in this region includes rocks assigned to four intervals that correspond approximately to the four provincial series of the system in the midcontinent region. The series and their approximate interval equivalents are: Kinderhook — interval A, Osage — interval B, Meramec — interval C, and Chester — interval D.

AUTHOR RESPONSIBILITY

During the data-gathering stage for this chapter the late Dr. Louise Jordan of the Oklahoma Geological Survey was supervising graduate students at the University of Oklahoma in subsurface studies of rocks of Mississippian age in Oklahoma. She had much original material resulting from these studies as well as copies of the various theses about subsurface geology in Oklahoma which contain lithic descriptions and lines of sections in the study areas. In addition to these data, Jordan had a personal log file. Jordan had no plans for publishing these data, and she readily made it available to Frezon for use in this report.

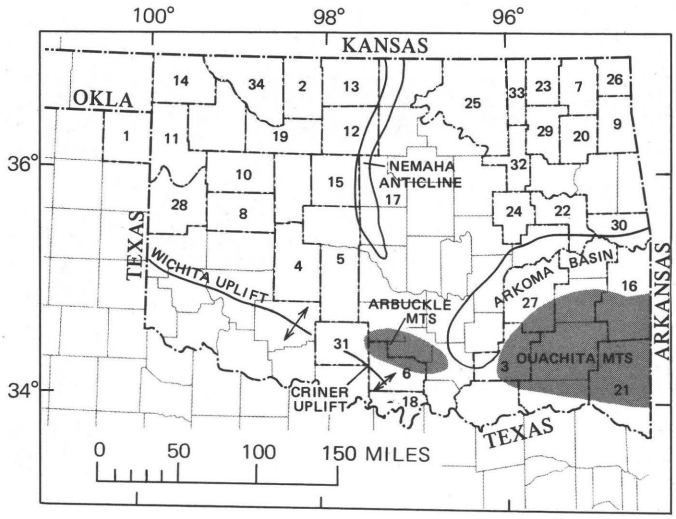
Opinions and interpretations expressed in this paper are those of Frezon and may or may not reflect opinions that Jordan may have held.

STRUCTURAL FEATURES

The mountain systems and structures shown in figure 28 are the result of Pennsylvanian tectonism that has altered the thickness and distribution of Mississippian rocks. These structures are used in this chapter as geographic references, and, unless specifically stated, they were not present or tectonically active in Mississippian time.

Surface structural and topographic features are the Arbuckle Mountains, the Ouachita Mountains, and on the Wichita and Criner uplifts, the Criner Hills and Wichita Mountains. Subsurface structures are the Nemaha anticline and the Arkoma basin. The Arkoma basin is the northern side of the earlier Ouachita basin. The Nemaha anticline is a linear positive feature of

[1] Oklahoma Geological Survey. Deceased, November 22, 1966.



IDENTIFICATION OF COUNTIES

TEXAS	8. Custer	17. Logan	26. Ottawa
1. Hemphill	9. Delaware	18. Love	27. Pittsburg
OKLAHOMA	10. Dewey	19. Major	28. Roger Mills
2. Alfalfa	11. Ellis	20. Mayes	29. Rogers
3. Atoka	12. Garfield	21. McCurtain	30. Sequoyah
4. Caddo	13. Grant	22. Muskogee	31. Stephens
5. Canadian	14. Harper	23. Nowata	32. Tulsa
6. Carter	15. Kingfisher	24. Okmulgee	33. Washington
7. Craig	16. Le Flore	25. Osage	34. Woods

FIGURE 28. — Index map of Oklahoma region showing structural and geographic features mentioned in text.

post-Mississippian age that extends from Oklahoma northward across Kansas.

Structural features that were active in Mississippian time are shown in figure 29. Tectonism was not strong during the first half of the period and structural differentiation of the region did not come into existence until the last half of Mississippian time. The structures consisted of an arcuate trough across the southern part of the region, within which two deep basin areas developed — in the west, the Anadarko basin, and in the east, the Ouachita basin. The Anadarko basin is the structural feature called the Southern Oklahoma geosyncline by Ham, Denison, and Merritt (1964, p. 150). North of the trough a broad shelf extended northward into Kansas. In the northeastern part of the State the southwestern part of the Ozark uplift was apparently active in Late Mississippian time.

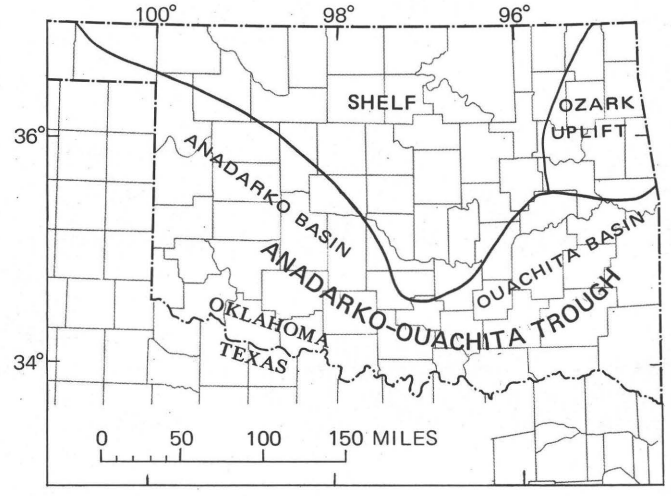


FIGURE 29. — Structural features in the Oklahoma region during Mississippian time.

PALEOGEOLOGY

LOWER BOUNDARY OF THE MISSISSIPPIAN

In Oklahoma, as in many other areas of the country, the lower boundary of the Mississippian System falls within a rock-stratigraphic unit. In the Arbuckle Mountains, as shown by studies of the conodonts in the Woodford Shale, the formation is Mississippian and Devonian in age (Hass and Huddle, 1965). In northeastern Oklahoma the equivalent of the Woodford, the Chattanooga Shale, is generally Devonian in age, but in at least one locality the upper part of the formation has a conodont fauna that is Mississippian in age (Hass, 1956, p. 29).

In the Ouachita Mountains in both Oklahoma and Arkansas, the middle division of the Arkansas Novaculite is in part the equivalent of the Chattanooga and Woodford of other areas, and the Devonian-Mississippian boundary occurs within this division (Hass, 1951, 1956).

The stratigraphic relations at the systemic boundary in Oklahoma are apparently similar to those at the boundary in Arkansas, where a basal sandstone of the Chattanooga, the Sylamore Sandstone Member, is a time-transgressive unit deposited in seas that encroached northward onto the Ozark uplift in Late Devonian and earliest Mississippian time. The Chattanooga in Arkansas is apparently both Mississippian and Devonian in age (Freeman and Schumacher, 1969).

In this chapter the base of the lower interval of the Mississippian (interval A) is placed arbitrarily at the top of the Chattanooga and equivalent units. This is an easily recognizable rock-stratigraphic boundary that is more indicative of paleotectonic change than the time-stratigraphic boundary in that it marks a distict change in sediment type and environment of deposition.

UNITS UNDERLYING THE MISSISSIPPIAN

Rocks assigned to the Mississippian in the Oklahoma region overlie rocks as young as Mississippian-Devonian age and rocks as old as Precambrian age (pl. 2). In most of the region the widespread Chattanooga Shale and its lithostratigraphic equivalents underlie the oldest rocks of Mississippian age of this report. These equivalents are called in various areas: the Woodford Shale, the Woodford Chert, and, in the Ouachita Moun-

tains, the middle division of the Arkansas Novaculite (pl. 15). All these units are believed to be earliest Mississippian and Late Devonian in age.

In five areas rocks of Mississippian age overlie rocks older than the Chattanooga and its equivalents. In northwestern Harper County, in the extreme northwest part of Oklahoma, the Woodford is absent, and Mississippian rocks overlie the Hunton Group of Devonian-Silurian age.

Four other areas where Mississippian rocks overlie rocks older than the Chattanooga occur in northeastern Oklahoma. The largest of these areas is almost entirely in Osage County (fig. 28). In this area rocks of the Simpson Group (Ordovician) are present in the southwest half of the area and older rocks of the Arbuckle Group (Cambrian-Ordovician) are present in the northeast half of the area. This is the area which Buchanan (1927) called the "Osage Island," and Leatherock and Bass (1936, p. 92) outlined in some detail.

The second largest area is an elongate, northwest-trending area centered in Rogers County. In this area rocks of the Simpson Group (Ordovician) are present beneath Mississippian rocks. This area includes the isolated localities in Tulsa and Rogers Counties where Leatherock and Bass (1936) indicated that the Chattanooga was absent.

Two isolated localities in which the Chattanooga is absent occur in northeastern Oklahoma: (1) in central Washington County, where rocks of Arbuckle age are present beneath the Mississippian, and (2) in southern Osage County, where basement rocks of probable Precambrian age underlie the Mississippian rocks.

In the southeastern part of the Ouachita Mountains, the middle division of the Arkansas Novaculite, the equivalent of the Chattanooga and Woodford, is presumed to be present beneath the upper division of the novaculite. In other areas of the Ouachita Mountains only the Devonian part or the lower division is believed to be present. The lower division of the novaculite and the Chattanooga to the north do not join, and the relations between the two formations are unknown.

PALEOTECTONIC IMPLICATIONS

The Chattanooga is thin or absent along a linear belt, here referred to as the Osage Island trend, that extends from Osage County southeastward to the Arkansas border. The two large areas where the Chattanooga is absent are along the northwest end of the trend. To the southeast along the trend the formation is present, but it is thinner on the trend axis than it is to the northeast and southwest. This trend is probably the locus of the greatest uplift at the end of Chattanooga time.

The Chattanooga thickens northeastward from the Osage Island trend. Most authors concur that in this area the Chattanooga is unconformably overlain by rocks of Kinerhook age (St. Joe Limestone Member of the Boone Formation), but there are many localities where the physical evidence of unconformity is obscure. In these localities the contact between the Chattanooga and the overlying St. Joe appears gradational, and the unconformity, if present, does not represent an extended period of uplift and erosion. A similar situation exists in northwestern Arkansas where the Chattanooga probably is conformably overlain by the St. Joe (Frezon and Glick, 1959, p. 178).

Outside northeastern Oklahoma the top of the Chattanooga is an unconformity representing a considerable length of time. In all except a few scattered localities interval C overlies the Chattanooga. This unconformity is the combined result of pre-interval A and pre-interval C erosion. The extent of pre-interval A erosion is uncertain, but probably uplift was slight. Interval A is believed to have been deposited over much of the area underlain by the Chattanooga.

Thus, some differential uplift occurred at the end of Chattanooga time. Only locally along the Osage Island trend was the uplift sufficient or prolonged enough to allow complete removal of the Chattanooga and to indicate a definite structural trend.

INTERVAL A

FORMATIONS INCLUDED

Some of the formations included in interval A are shown on plate 15. The most extensive outcrops of interval A occur in the Ozark area in northeastern Oklahoma where the interval consists of rocks included in the St. Joe Limestone Member of the Boone Formation. This member is the same as the *St. Joe Group* of Huffman and is equivalent to, in ascending order: the Compton, Sedalia, Northview, and *Pierson* Formations (Huffman, 1958) of southwestern Missouri (pl. 15, col. 56). The Compton and Northview are Kinderhook in age, and thus belong in interval A, but the *Pierson* is late Kinderhook and early Osage in age (Thompson and Fellows, 1969). The formations of the *St. Joe Group* (Huffman, 1958) have not been mapped and have not been differentiated for this report. For this reason interval A includes the *Pierson* equivalent wherever it is present in the region.

In the Arbuckle Mountains area of south-central Oklahoma interval A consists of the Welden Limestone. Its equivalents are the *pre-Sycamore beds* of Harlton (1956) and the *Cornell Ranch Member* of the *Sycamore Formation* of Prestridge (1959).

UPPER BOUNDARY OF INTERVAL A

In outcrops in northeastern Oklahoma, rocks of interval A are overlain unconformably by the cherty limestones in the Boone Formation. Some authors assign these limestones in this area to the Reeds Spring Member of the Boone Formation, and this member of the Boone is described by Huffman (1958, p. 43) as dark-gray fine-grained limestone and interbedded dark-gray and blue-gray chert. These rocks are readily separable from the light-colored and generally green clayey underlying rocks of the St. Joe Limestone Member assigned to interval A.

The Boone and equivalent interval B rocks are absent in the northern Arbuckle Mountains and in the subsurface north of the mountains, where the Welden Limestone of interval A is unconformably overlain by the Caney Shale of interval C. The base of the Caney is dark-gray, locally glauconitic, calcareous siltstone. In the southern part of the uplift and in the subsurface south and west of the uplift, rocks equivalent to the Welden in interval A are unconformably overlain by the Sycamore Limestone of interval C.

THICKNESS TRENDS

Rocks of interval A in the Oklahoma region are the thinnest and areally the most restricted of all the Mississippian intervals (pl. 3-A). In the northeastern part of the region rocks of interval A form a large continuous body, which has a maximum thickness of about 100 feet. Thicknesses greater than 50 feet occur in Craig, northern Mayes, and western Nowata Counties. South and west from this area of maximum thickness the interval thins to a zero edge.

Outliers of the interval occur in a band extending from the area of Okmulgee County southwestward as far as Love County in the south-central part of the region. Within this band, but not shown on the map, are occurrences of Welden Limestone 3–5 feet thick in the northeasternmost part of the Arbuckle Mountains (Barker, 1951). The maximum thickness (37 feet) in these outliers occurs in Love County. One isolated outlier in northern Kingfisher County has 20 feet of limestone assigned to the interval.

The thickness differences of the numerous outliers of interval A are apparently the result of post-interval A erosion. The interval may have been considerably thicker than indicated on the restored thickness maps (pl. 10, fig. 1).

LITHOFACIES TRENDS

Interval A consists mainly of limestone and clayey limestone, and of lesser amounts of calcareous mudstone (pl. 3-B).

Sand, where present in the interval, is restricted to the basal beds and occurs most commonly at places where the interval overlies rocks of pre-Chattanooga age. It rarely is present in northeastern Oklahoma in sufficient quantity to form a discrete bed of sandstone similar to the *Bachelor Formation* in Missouri (Thompson and Fellows, 1969). Generally, sandy beds in the lowest part of the interval consist of sandy or muddy limestone. The thickness of sandy beds is nowhere great enough to reflect in the lithofacies pattern.

A narrow, sinuous belt of limestone containing less than 20 percent mudstone extends from northeast Oklahoma southwestward to the Arbuckle Mountains area (pl. 3-B). Outside this belt, interval A contains greater than 20 percent mudstone. Other lithofacies trends for the interval in the region are not apparent.

DEPOSITIONAL HISTORY

Deposition of the Chattanooga Shale was followed by a withdrawing of the sea, after which the region was again flooded and a new and a different marine depositional environment was established. The widespread euxenic environment that characterized Chattanooga time was replaced during interval A by a more normal marine environment in which the water was shallow and well oxygenated. An abundant marine fauna (Huffman, 1958, p. 42–43) established itself early in this sea and persisted to the end of the interval. Apparently, relief on the sea floor was extremely low.

Sands in interval A were probably derived from a sandy regolith that had developed on pre-Chattanooga rocks, and the sands of the regolith were cleaned and distributed by current action.

The source of clays in the interval is not known. If the clays were derived from the underlying Chattanooga, they have been extensively altered.

PALEOTECTONIC IMPLICATIONS

Minimal tectonic activity occurred in and near the region during interval A time. Tectonic stability is indicated by the fact that the water depth remained consistently shallow enough to support the marine fauna that persisted through interval A time.

Both thickness and lithofacies trends cut across the axes of structural features that were active in Devonian time.

INTERVAL B

FORMATIONS INCLUDED

In northeastern Oklahoma rocks of interval B comprise the lower part of the Boone Formation below the Short Creek Oolite Member (pl. 15, col. 72). In the Ar-

buckle Mountains area the interval is not represented in the stratigraphic section. In the Ouachita Mountains in Oklahoma and Arkansas the upper division of the Arkansas Novaculite makes up interval B. The upper division apparently is late Kinderhook and Osage in age (Hass, 1951) and probably equivalent to part of the Boone. The upper division of the novaculite apparently is not present in Oklahoma northwest of McCurtain County (Hass, 1951, p. 2540, citation of Miser). In the subsurface, rocks that are stratigraphically equivalent and lithically similar to the Boone are included in the interval.

UPPER BOUNDARY OF INTERVAL B

In Ottawa County, in northeasternmost Oklahoma, the top of interval B is placed within the Boone Formation at a conformable contact at the base of the Short Creek Oolite Member of the Boone. The Short Creek has been mapped as far south as the southern border of Ottawa County (Reed and others, 1955). At most places south of Ottawa County, the Moorefield Formation unconformably overlies a part of the Boone Formation older than the Short Creek (fig. 30). In this area the base of the Moorefield is the upper boundary of interval B. At a few scattered places where the Short Creek is found south of Ottawa County, it has been included in interval B.

At most places in northeastern Oklahoma the top of interval B is a readily recognizable unconformity. The lithic difference between interval B rocks and overlying rocks is not everywhere the same, but the break everywhere is believed to separate rocks of Osage age from overlying rocks of Meramec age. Limestone in interval C characteristically is darker colored and less cherty than limestone in interval B. Shale and siltstone are present in interval C but absent in interval B. Where the Moorefield Formation and the equivalent rocks overlie interval B, the lower part of the Moorefield is glauconitic, and at most places the lowermost bed of the formation contains abundant glauconite. The upper part of interval B at some places contains scattered glauconite, but glauconite is nowhere as abundant as in overlying interval C.

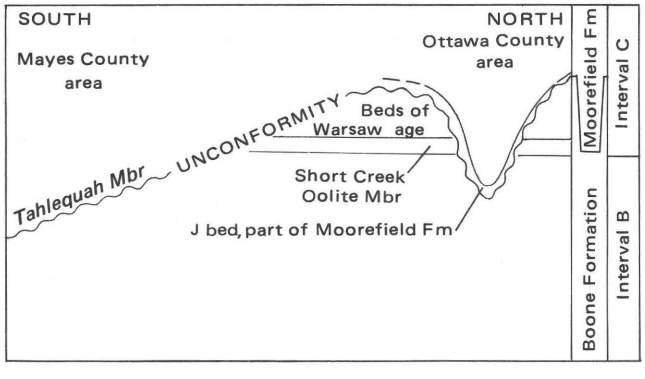


FIGURE 30. — Diagrammatic section showing relation of stratigraphic units in Ottawa County to units in Mayes County and areas to the south, all in Oklahoma.

The contact between rocks of Osage and Meramec ages in northern Oklahoma is drawn somewhat lower in this report than it was by Jordon and Rowland (1959, fig. 2); consequently, the zero line for the Osage (interval B) is drawn farther north in this report than it was by them.

THICKNESS TRENDS

Rocks of interval B are present in the northeast, north, and west half of the Oklahoma region (pl. 4-A). The maximum thickness, probably greater than 700 feet, occurs in western Roger Mills and Ellis Counties in the western part of the Anadarko basin. No thicknesses greater than 700 feet are reported in western Oklahoma, but a thickness of 760 feet occurs in northeastern Hemphill County, Tex., about 6 miles west of the western boundary of Roger Mills County, Okla., and the 700-foot isopach probably extends into western Oklahoma. The interval is thicker than 600 feet in the western Oklahoma counties and in western Dewey and Custer Counties, Okla.

From the area of maximum thickness in the western part of the Anadarko basin, interval B thins northward and westward to zero edges in adjoining areas, and interval B thins eastward to a northeast-trending zero edge in Caddo and Canadian Counties. In western Logan and eastern Garfield Counties, the east limits of the interval are faults of post-interval D age along the trend of the Nemaha anticline.

Along the southern edge of the Anadarko basin the limits of the interval are a combination of a post-interval D fault and depositional margins that mark the northern part of the Wichita uplift.

The maximum thickness of the interval is greater than 360 feet in Ottawa County in northeastern Oklahoma. The interval thins southward and southwestward to a zero edge in the subsurface (pl. 9-F, secs. *I–I'*, *K–K'*). Thinning from Ottawa County southeastward around the edge of the Ozark uplift has been demonstrated by surface mapping (Huffman, 1958, fig. 4). At three surface localities, one in northeastern Muskogee County and two in northwestern Sequoyah County, interval B (Boone Formation) is absent and rocks of interval C age rest on the Chattanooga Shale.

Mapping studies around the edge of the Ozark uplift in northeastern Oklahoma (Huffman, 1958) indicate that two parts of the Boone Formation (*Keokuk Formation* at the top and Reeds Spring Formation at the base)

can be distinguished over the whole area. The absence of the *Keokuk* where the Boone is thinnest is convincing evidence of southward truncation of the Boone by erosion. This evidence had led Frezon to conclude that no shale facies of the Boone Formation exists in eastern Oklahoma, such as has been proposed by Glick (chap. H) for the formation in western Arkansas.

The zero edge of interval B southeast of eastern Le Flore County, Okla., cannot be located accurately. The west-trending zero edge of the interval in McCurtain County is drawn to show the presence of the upper division of the Arkansas Novaculite in this area. The distribution of the upper division of the novaculite in the De Queen and Caddo Gap quadrangles of Arkansas and easternmost Oklahoma (Miser and Purdue, 1929) indicates that the zero line probably swings eastward from McCurtain County, Okla., for some undetermined distance into western Arkansas and then turns northwestward into eastern Le Flore County, Okla.

LITHOFACIES TRENDS

In most of the Oklahoma region, interval B consists predominantly of limestone or cherty limestone that contains less than 20 percent terrigineous material. In five small areas (pl. 4-B) in Grant, Garfield, Major, Kingfisher, and Harper Counties, the interval has more than 20 percent terrigineous material; however, no trend is apparent in the distribution of these areas of more clastic rocks.

Rocks of interval B are characteristically cherty. On plate 4-B an overlay pattern has been added to distinguish areas where chert constitutes more than 10 percent of the rock. Data used in compiling this chapter do not permit an accurate determination of the chert percentage, and the boundaries are approximate.

Chert in interval B resulted from the replacement of carbonates by silica. The carbonate-chert relations, the lateral and vertical variations in the amount of chert, and possible origins of the chert in the Boone Formation of interval B were discussed by Giles (1935).

DEPOSITIONAL HISTORY

The carbonates that were the original sediments of interval B were deposited in a low-energy marine environment. Within the area of present occurrence of the interval, the waters were clear and well oxygenated, and they supported an abundant marine invertebrate fauna. The extensive and abundant carbonate that must have been present in the area prior to silicification suggests a warm climate. There is no evidence of a shoreward, coarse detrital facies in this region; the rocks present were probably deposited in an open sea, remote from land (pl. 11, fig. 2; pl. 12, fig. 2).

The detrital sediments in the interval consisted almost entirely of mud that generally was incorporated in the limestone rather than deposited as separate beds. In the northeast part of the region, Giles (1935) stated that detrital heavy minerals present in limestones in the interval include apatite, hematite, magnetite, limonite, ilmenite, pyrite, zircon, and tourmaline. Fragmental crystalline and milky quartz are also present in the limestones (Giles, 1935). The amounts and distribution of the detrital heavy minerals are unreported. The fine size of the detrital materials and the character of the previously described suite of heavy minerals (Giles, 1935) do not require a nearby source.

PALEOTECTONIC IMPLICATIONS

The thickness variations of the isopach pattern (pl. 10, fig. 2) are the result of erosion prior to the deposition of interval C, rather than variations in depositional thickness. The uniform facies pattern (pl. 4-B) is characteristic of carbonates deposited on a marine shelf. An optimum, probably shallow water depth for the carbonate-secreting organisms was established early in the interval, and subsidence and deposition maintained that depth throughout the interval.

INTERVAL C

FORMATIONS INCLUDED

Rocks assigned to interval C crop out in the areas of the Ozark uplift and the Arbuckle and Ouachita Mountains. In all areas rocks of the interval are of the Meramec age (pl. 15, cols. 72–74).

In most of the Ozark area the rocks of the interval have been mapped as the Moorefield Formation (Huffman, 1958). This unit is divided into four members which are, in ascending order: the *Tahlequah*, *Bayou Manard*, *Lindsey Bridge*, and *Ordnance Plant*. In northeasternmost Oklahoma the interval includes, in addition to the Moorefield Formation equivalents, the part of the Boone Formation above the base of the Short Creek Oolite Member.

The fauna of the Short Creek includes fossils of Warsaw (Meramec) age in association with other species that ranged upward from rocks of Osage age below the oolite, according to Mackenzie Gordon Jr. (*in* McKnight and Fischer, 1970, p. 39). *Marginirugus magnus* (Meek and Worthen), one of the brachiopods believed to indicate Warsaw age, is found in nonoolitic limestone associated with the oolite.

The distribution of the Short Creek Oolite Member south of Ottawa County is discontinuous largely because of truncation beneath the J bed (of Fowler and

Lyden, 1932) at the base of the Moorefield (fig. 30). There are no published maps of Delaware County to the south so the distribution of the member in this area is unknown. The member is not discussed or indicated as being present in Craig County to the west (Branson and others, 1965). From northern Mayes County southward in the area of outcrop, the distribution of the member is very restricted (Huffman, 1958, p. 45), and an oolite believed to be the Short Creek occurs at the top of the Boone only at a few places where the formation is thicker than 200 feet.

Rocks of Meramec age are not well defined in the Ouachita Mountains. Interval C includes some undetermined thickness of the Caney Shale in the northern and western parts of the Ouachita Mountains (the so-called frontal Ouachitas), and the interval includes some undetermined part of the Stanley Shale in the main part of the Ouachita Mountains to the south.

In the Arbuckle Mountains the interval includes, on the north side of the mountains, the *Ahloso Member* of Elias (1956) of the Caney Shale and, on the south side of the mountains, the Sycamore Limestone. The terms "Mayes" and "Ada Mayes" are used to designate equivalent rocks in the subsurface in the eastern part of the region. In the western part equivalent rocks in the subsurface are designated by the series name.

UPPER BOUNDARY OF INTERVAL C

In northeastern Oklahoma the Moorefield Formation is overlain by the Hindsville Limestone. The Hindsville is of early Chester age and is assigned to interval D. In the Arbuckle Mountains interval C is overlain by the *Delaware Creek Member* of Elias (1956) of the Caney Shale. The *Delaware Creek* is a noncalcareous shale that can be readily differentiated from the calcareous shales of the interval C age.

In the subsurface throughout the region the top of interval C rocks is the top of a calcareous shale and siltstone sequence correlated with the *Ahloso Member* of the Caney Shale. These rocks are overlain by noncalcareous shale in which some limestone beds are locally present.

The upper boundary of interval C in the Ouachita Mountains in Oklahoma cannot be determined consistently. In this area the lower part of the Stanley and the underlying Caney occupy the stratigraphic position of rocks of intervals C and D in adjoining areas. The base of the Caney is probably Meramec in age (Hass, 1950), and the upper part of the Stanley is Chester in age (Gordon and Stone, 1969); the Meramec-Chester boundary within this sequence has not been established.

THICKNESS TRENDS

Areas of maximum accumulation of rocks of interval C age occur in the west, the southwest, and the southeast parts of the Oklahoma region (pl. 5-A). The greatest measured thickness in the region, more than 900 feet, occurs in the northwest in Harper County. This area appears to be a northeast-extending arm of the Anadarko basin. In the southwest, north of the Wichita and Criner uplifts, the interval locally is thicker than 700 feet in southern Caddo County, Okla. In the southeast, the interval thickens southward into the Arkoma basin, and the known thickness of the interval is greater than 400 feet in eastern Pittsburg County, Okla.

Thickness trends of interval C rocks are at places parallel to and at other places normal to thickness trends in the older intervals. At no place is there any evidence that folding during deposition of the older rocks continued and controlled depositional thickening and thinning in interval C rocks.

In the Ouachita Mountains the thickness of the interval is unknown because of the inability to determine an upper boundary for the interval. It is assumed, however, that rocks of Meramec age are thicker than 500 feet in most of the area, even though thicknesses are not shown on the isopach map. In northern McCurtain County the thickness is probably greater than 1,000 feet and may be as great as 6,000 feet.

The distribution pattern of interval C rocks in northern Oklahoma differs from the pattern presented by Jordan and Rowland (1959, fig. 4) because Frezon has included rocks in interval C which those authors assigned to interval B.

LITHOFACIES TRENDS

Rocks of interval C consist of muddy limestone and calcareous mudstone (pl. 5-B). Muddy limestone (mud less than 20 percent), the most prevalent rock type, occupies the entire northern one-third of the region and has spotty distribution in the southern part of the region. Mudstone that is variably calcareous is dominant to the south. The Caney Shale in the northern Ouachita Mountains is a southern extension of the calcareous shale. Farther south, in the area where the Stanley is differentiated, the shale is not calcareous. Sandstone is present in the Stanley (Cline and Shelburne, 1959), and, if adequate data were available, facies patterns indicating the presence of sand probably would be shown on the map.

In the outcrop area in northeastern Oklahoma, the interval locally contains detrital chert in clasts up to pebble size, but the percentage of sandstone and conglomerate is not great enough to influence lithofacies

patterns. The chert was derived from the underlying Boone Formation and is found at the base of the Moorefield in the *Tahlequah Member*. This relation shows that silicification of the Boone was completed prior to interval C time (Huffman, 1958, p. 50).

The distribution of muddy limestone or calcareous mudstone is not related to the thickness trends.

In most of the area where rocks of the interval are composed of more than 80 percent carbonate, they have become silicified in various degrees. The percentage of chert in these rocks cannot be determined accurately from available lithic descriptions. It has been assumed to be generally 10–15 percent, but nowhere does it approach the percentage found in interval B. As in interval B, silicification appears to have been the result of diagenetic alteration of carbonate rocks.

DEPOSITIONAL HISTORY

Interval C was the first time in the Mississippian that all the Oklahoma region become a depositional area (pl. 10, fig. 3).

During interval C time relatively thick sediments accumulated in a deep trough in the Ouachita Mountains area in southern Arkansas and southeastern Oklahoma, and a thinner sequence accumulated in shallower water in the remainder of the region.

Mud, silt, and sand from north and south of the trough were moved into deeper water by turbidity currents, then were swept westward along the axis of a trough by currents, then were deposited in the area of the present-day Arkoma Basin and Ouachita Mountains (Johnson, 1968). Apparently, the closest source for these sediments were positive areas south of the trough in Texas and Louisiana, although eastern source areas were also available. The nearest possible source area of detrital material to the west would have been in western New Mexico (pl. 10, fig. 30), but apparently, nothing was contributed from this direction.

Some differentiation of sediments occurred in the process of westward transport by axial currents. East of an area in the vicinity of Atoka County, Okla., interval C contains sand in the lower part of the Stanley Shale. In Atoka County and in the area to the west, little or no sand-sized material is present in the interval. Apparently, shallow-water shoals were present in and near Atkoa County, and current energy was not great enough to move sand-sized material westward onto and across the shoals.

Currents entering the Anadarko basin region west of the shoal flowed into a large shallow-water inland sea with no outlet to the west, north, or south. Bottom currents were forced to spread outward over a relatively broad area and dissipate their energy. The absence of mud indicates that transporting ability of the currents decreased in the western part of the sea. Apparently, normal circulation could not generate currents of the magnitude and scale necessary to move detrital sediments into the eastern part of the sea.

In the Anadarko basin region differentiation of sediments is rather striking. In the eastern part of the area, presumably adjacent to the shoal, the sediments are muddy. The shoal was probably in a higher energy environment than the deeper water area to the west and deposition of terriginous material took place in the lower energy area. A comparison of the restored thickness and lithofacies maps (pl. 10, fig. 3; pl. 5-B) shows that the western part of the Anadarko region was an area of carbonate deposition in which thickness trends of the carbonates are normal to the lithofacies boundaries. This region apparently was beyond the extent of mud-bearing currents.

Two characteristics in the sediments suggest deep water in the Ouachita Mountains area (pl. 12, fig. 3). These are the absence of calcium carbonate and the presence of dark shale, which suggest reducing conditions and, hence, low current or wave activity (Johnson, 1968).

The Anadarko basin area, where carbonates are present, was probably shallower than the Ouachita basin area, where carbonates are absent. Carbonates deposited north of the Anadarko basin suggest that shallow normal marine conditions also prevailed (pl. 11, fig. 3) in that area. In the oxygenated, shallow water north and northwest of the Ouachita basin, an abundant invertebrate fauna was able to grow and flourish (Huffman, 1958, p. 53–54). Some muddy waters from the Ouachita basins area to the south were introduced westward but muds that were brought in were never sufficient to obliterate the carbonate-producing faunas.

Shoals adjacent to the deeper water Ouachita basin to the south were a marginal area in which mud influx was great enough to prevent the extensive growth of marine faunas in an otherwise favorable environment.

In the Ozark area chert conglomerate in various parts of the Moorefield Formation indicate a land area of appreciable relief in this vicinity during a considerable span of the interval (Huffman, 1958). The maximum thickness of the Moorefield in this area is about 100 feet. The land was inundated by the end of the interval.

PALEOTECTONIC IMPLICATIONS

During interval C, the axis of a trough extended in an arc from northern McCurtain County, Okla., westward to Carter County and then northwestward to the Texas border in the vicinity of northern Ellis County, Okla. (pl. 10, fig. 3). The trough was divided into two more rapidly sinking parts separated by an intervening saddle. The two parts, the Ouachita and Anadarko

basins, were in the extreme east and in the extreme northwest ends of the trough, respectively; the intervening saddle apparently was in the vicinity of Atakoa County, Okla.

The rate of subsidence was slow in a shelf region to the north of the Anadarko and Ouachita basins and varied from fast to slow in the Ouachita basin in the southern part of the region.

The Anadarko-Ouachita trough resulted from a rejuvenation of a negative tectonic feature that had existed intermittently as far back as late late Precambrian or Early Cambrian time (Ham and others, 1964, p. 149–150). The axis of the trough in Mississippian time is generally along the axis of thicker sediments of Cambrian, Ordovician, Silurian, and Devonian ages (Sloss and others, 1960). Subsidence that started in interval C time continued until Des Moines (Middle Pennsylvanian) time when the Ouachita part of the trough was elevated and folded during the Ouachita orogeny.

Subsidence in the Anadarko area was greater than in the Ouachita area (pl. 10, fig. 3). Sediment-bearing currents flowing along the axis of the trough were probably tectonically induced by the subsidence. Subsidence of 1 inch per square mile would increase the volume of the basin by about 3 million cubic feet, and the subsidence probably occurred over an area of several thousand square miles. Regardless of whether subsidence was continuous or whether it occurred in distinct episodes, water and sediment would have flowed westward into the subsiding area. Sedimentation in the basin must have been fairly rapid to maintain a shallow-water environment suitable for the formation of bioclastic carbonate that makes up the interval.

INTERVAL D

FORMATIONS INCLUDED

Rocks of interval D age crop out in three areas in the region. In the Ozark area formations included in the interval are, in ascending order: the Hindsville Limestone, Fayetteville Shale, and Pitkin Limestone. The Hindsville in Oklahoma is regarded as a formation; whereas in the type locality of the unit in northwestern Arkansas it is considered a member of the Batesville Sandstone. In the northern Arbuckle Mountains area the upper part of the *Delaware Creek Member* and the overlying *Sand Branch Member* of the Caney Shale and the *Rhoda Creek Formation* are of Chester age and are assigned to interval D age. All the *Delaware Creek* is mapped in the interval even though its lower part is of Meramec age. On the south side of the Arbuckle Mountains the Goddard Shale is the equivalent of the *Sand Branch Member* of the Caney and assigned to interval D.

In the Ouachita Mountains area the upper part of the Stanley Shale is the interval D equivalent. Because fossils are extremely scarce and the formation cannot be divided on a lithologic basis, thicknesses in the Ouachita area are not shown on the map (pl. 6-A). This practice is the same as that used in interval C.

UPPER BOUNDARY OF INTERVAL D

Interval D is overlain by rocks ranging in age from late Chester (Springer) or early Morrow to Des Moines.

At most places the upper boundary of interval D is a readily recognizable lithic boundary; the overlying rocks generally are sandy, whereas interval D rocks are shale or limestone. Where shales of Morrow or late Springer age overlie interval D, they are non-calcereous and commonly have siderite beds in contrast to the calcareous shale or limestone of interval D. In northeastern Oklahoma the top of the interval is placed at the top of the Pitkin Limestone. In other areas where the Pitkin is absent, the top of the calcareous Caney Shale is the top of the interval.

THICKNESS TRENDS

Rocks of interval D are present in the eastern and western parts of the Oklahoma region and are absent in an area that broadens northward in the central part of the region (pl. 6-A; pl. 9-F, sec. *I–I'*).

Interval D thickens gradually southeastward from a northeast-trending edge in central Oklahoma. Thicknesses of several hundred feet probably occur in southeastern Oklahoma in the Arkoma basin, but the absence of wells precludes determination of these thicknesses. The thickening trend probably continues southeastward into the Ouachita Mountains area where a depositional trough was present during interval D time. In an area of maximum deposition, in northern McCurtain County, the maximum thickness of the interval probably exceeds 1,000 feet.

In the western part of Oklahoma the interval thickens southwestward from a north- or northwest-trending edge in central Oklahoma. The greatest thickness, more than 1,700 feet, occurs in Caddo County in the area of the Anadarko basin.

LITHOFACIES TRENDS

Interval D rocks are mainly mudstone and carbonate (pl. 6-B). In northeastern Oklahoma, sandstone is locally present in the Fayetteville Shale (Wedington Sandstone Member), but neither the thickness nor the distribution of the member is great enough to effect the lithofacies pattern in the area. In the Ozark area the Hindsville Limestone locally has chert conglomerate at its base. This conglomerate occurs where the formation rests on the Boone which was the source of the chert.

In the Ouachita Mountains area the interval consists of shale, siltstone, and sandstone, but a trend toward increasingly larger proportions of shale southward indicates that shale probably composes more than 80 percent of the interval in most of southeastern Oklahoma.

In the area where control is available in eastern Oklahoma, increasing thickness is accompanied by a general change from limestone to mudstone. In the Anadarko basin region in the western part of the State muddy limestone grades eastward to mudstone which is normal to the thickness trends.

DEPOSITIONAL HISTORY

The depositional history of interval D is similar to that of interval C. Axes of maximum deposition locally shifted, but the environments of deposition were essentially the same (pl. 11, figs. 3, 4).

Terrigenous sediments apparently came from the same southern and eastern sources as interval C sediments. Sand was largely confined to the Stanley Shale in the southeastern part of the region in the area of the Ouachita trough. The distribution of carbonate and terriginous mud in the western part of the region probably resulted from distributive currents similar to those that existed during interval C time.

Undetermined but probably minor amounts of sediment came from the land areas in northern Kansas and in eastern Colorado and New Mexico (pl. 12, fig. 4). Because rocks exposed in these land areas were in part calcareous mudstone and carbonate of underlying interval C, the sediments of interval D from these sources were exclusively fine grained.

In northeastern Oklahoma, some alternation of shale and limestone suggests cyclic sedimentation. The interval consists of, in ascending order: limestone (Hindsville), shale (Fayetteville), and limestone (Pitkin), which probably were deposited in open-marine, nearshore, and open-marine environments, respectively. Water depth during deposition of the Hindsville was probably shallow as a result of slight uplift on a land area that may have existed in the north-central part of the region, or of a slight uplift of the Ozark dome.

PALEOTECTONIC IMPLICATIONS

Thickness relations indicate that the Anadarko-Ouachita trough (fig. 29) was active and was a major structure controlling sedimentation. The trough became broader during interval D than during interval C, and the Anadarko and Ouachita basins within the trough became better differentiated (pl. 10, fig. 4). This differentation largely came about as the result of a change in the trend of the main axis of the Ouachita basin. In the preceding interval C time the major axis of the Ouachita part of the trough trended westward through southeastern Oklahoma as far as western Atoka County. In interval D time the main axis of the Ouachita part of the trough trended southwestward across the corner of Oklahoma into Texas (pl. 10, fig. 4). The Ouachita and Anadarko basins were joined along a subsidiary axis extending westward from a west-trending salient in the Ouachita Mountains part of the trough.

At the end of interval C time or the beginning of interval D time uplift occurred in the northeastern part of the region, but it does not appear that the region was emergent for a long period or that erosion was extensive. Because the Hindsville locally rests on the Boone Formation in northeastern Oklahoma, any rocks of interval C that were deposited were removed. In the Anadarko-Ouachita trough to the south subsidence continued slowly from interval C (Meramec) time into interval D (Chester) time; no interruption in sedimentation can be detected at the interval boundary.

Probably, the major new tectonic event during the interval was a realinement of the Ouachita basin as a distinct arcuate southwest-trending negative feature distinct from the previously west-trending trough. The change of the major axis of the trough to a southwest direction was a change of about 45° from the previous axial trend. Restored thickness lines (pl. 10, fig. 4) indicate that a minor axis branched off the major axis. This minor axis of downwarping trended westward and then northwestward to a saddle between the Ouachita and Anadarko basins in Stephens County, Okla.

The strong west-trending axis of deposition linking the Ouachita and Anadarko basins redeveloped in Early Pennsylvanian (Morrow) time.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

The pattern of thickness of Mississippian rocks (pl. 7) shows three regions of different structural activity. In the west downwarping permitted accumulation of a thick sequence of Mississippian rocks in the Anadarko basin; in the northeast relatively thin Mississippian rocks indicate a more stable shelf north of the Anadarko and Ouachita basins; and in the southeast the southward thickening of the Mississippian rocks shows the influence of downwarping along the Ouachita trough. A relatively narrow, north-trending area where rocks of the system are either absent or are thin is the Nemaha anticline (fig. 28), a post-Mississippian structural feature.

In the Anadarko basin, rocks of Mississippian age are thickest in Ellis County, Okla., and in the Texas Panhandle immediately to the west. In this area, which

was a depositional site during most of Mississippian time, the rocks of the system are more than 3,000 feet thick.

The thinning that occurs in all directions from the area of maximum thickness shows a rather broad basin. The area of maximum total thickness coincides with the area of maximum thickness of intervals B and C; the area of maximum thickness of interval D is farther east. The restored thickness map of interval D (pl. 10, fig. 4) best defines the configuration of the basin during all Mississippian time because the marginal parts of the basin have been altered by post-Mississippian erosion.

In the Anadarko-Ouachita trough, where rocks of interval D are overlain by rocks of Springer or Morrow age, thicknesses have not been extensively altered by post-Mississippian erosion. In the north-central part of the region, where Mississippian rocks are overlain by younger rocks of Pennsylvanian (Des Moines) age, irregular amounts of Mississippian rocks have been removed by postdepositional erosion.

The southeasternmost part of the nose of the Central Kansas uplift, a post-Mississippian structural feature, is indicated by thinning in northern Woods and Alfalfa Counties.

The shelf north of the Ouachita basin is characterized by thickness of less than 500 feet and by widely spaced and irregularly spaced isopachs. The trends in this area are largely the result of post-Mississippian erosion; most of the area was subjected to repeated and extensive erosion in Early Pennsylvanian time until it was finally covered in Middle Pennsylvanian (Des Moines) time.

The greatest thicknesses of Mississippian rocks in the Oklahoma region occurs in the Ouachita Mountains where the Stanley Shale reportedly exceeds 10,000 feet. (See Hart, 1963, for summary.) In Oklahoma the Stanley includes beds of Pennsylvanian age, but even if 50 percent of the thickness is Pennsylvanian, which is improbable, total thickness of Mississippian rocks in the area would be on the order of 5,500 to 6,000 feet. Overthrusting in the Ouachita Mountains has displaced rocks of Pennsylvanian and older age northward an unknown distance. Thus, the locus of maximum depositional thickness in the area cannot be determined.

Rocks of Mississippian age were deposited across the Wichita uplift, but they were removed from this structure by uplift and erosion that started in Early Pennsylvanian time.

A poorly defined north-trending area in central Oklahoma where Mississippian rocks are thin or absent was originally covered by rocks of Chester (interval D) age as well as some older Mississippian rocks. Absence of Mississippian rocks is due to post-Mississippian uplift and erosion on most of this structure.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM AND PALEOTECTONIC IMPLICATIONS

Pennsylvanian sedimentary rocks overlie rocks of Mississippian age in all parts of the region (pl. 8). At the end of Mississippian time general uplift resulted in withdrawal of seas into the arcuate Anadarko-Ouachita trough in the southern part of the region where deposition was continuous from Chester to Morrow time. The axis of this trough extended westward from northern McCurtain County to Carter County and then northwestward to Roger Mills County.

The lower part of the Springer Formation was deposited in the western (Anadarko) part of the trough. Although data are not available for the eastern (Ouachita) part of the trough, it is probable that a similar Upper Mississippian sequence, equivalent to the Mississippian part of the Mississippian and Pennsylvanian Cane Hill Member of the Hale Formation in northern Arkansas, was deposited in this area and is now represented in the uppermost Stanley Shale and lowermost part of the overlying Jackfork Sandstone.

Pennsylvanian rocks were deposited elsewhere during a slow encroachment of Pennsylvanian seas onto land areas where Mississippian rocks were exposed. By Morrow time this encroachment was well under way in parts of Arkansas where the Pennsylvanian Cane Hill Member of the Hale Formation was deposited on Mississippian rocks. Rocks of late Morrow age are the basal Pennsylvanian deposit farther north, and rocks of Atoka age lap onto Mississippian rocks over a more extensive region still farther north.

Mississippian rocks in the north-central part of the region were finally covered in Middle Pennsylvanian (early Des Moines) time.

During the Ouachita orogeny, which started in late Des Moines time, rocks of early Des Moines and older ages were uplifted, folded, and eroded. Mississippian rocks were exposed in several folds prior to Early Cretaceous time, and in the southeasternmost part of the region rocks of Early Cretaceous age were deposited on the Mississippian.

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Southern Midcontinent and Southern Rocky Mountain Region

By WILLIAM J. MAPEL, ROSS B. JOHNSON, GEORGE O. BACHMAN, *and* KATHARINE L. VARNES

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–J

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

SOUTHERN MIDCONTINENT AND SOUTHERN ROCKY MOUNTAINS REGION

By WILLIAM J. MAPEL, ROSS B. JOHNSON, GEORGE O. BACHMAN, and KATHARINE L. VARNES

ABSTRACT

Initial Mississippian deposits at most places in the Southern Midcontinent and Southern Rocky Mountains region are Kinderhook, Osage, or Meramec in age and are unconformable on broadly folded older rocks ranging in age from Late Devonian to Precambrian. These Mississippian rocks were laid down during transgression of the sea across a Late Devonian or Early Mississippian surface of low relief.

Interval A, which is as much as 200 feet thick, has an irregular thickness and distribution in New Mexico and central Texas that may have resulted from deposition in shallow drowned valleys on the slowly sinking pre-Mississippian land surface.

Intervals B and C are mostly limestone deposited in a shallow epicontinental sea. During this time, local downwarping occurred in the Texas Panhandle at the west end of the Anadarko basin where nearly 2,000 feet of limestone accumulated together with some interbedded sandy limestone and calcareous shale. A second depositional center in west Texas and adjacent southeast New Mexico sank sufficiently for nearly 1,000 feet of cherty and oolitic limestone to accumulate in a broad basinal area roughly coincident with the later Delaware basin of Pennsylvanian and Permian time. Low-lying land may have persisted throughout Early Mississippian time in west-central New Mexico and in an area east or southeast of the Llano region in central Texas.

At the end of interval C (late Meramec time), the region was slightly raised, and the sea withdrew to the southwest and northeast. Erosion stripped away most or all previously deposited Mississippian rocks in parts of southwest and central Texas along the trend of the Pecos arch, in the Llano region, and in the eastern part of the Fort Worth basin. Elsewhere, dissection was slight.

During interval D (beginning probably in latest Meramec time), the Ouachita geosyncline formed in a broad northwest-facing arc across east and south Texas, and thereafter, until sometime before the close of Early Permian, this structural element sank rapidly and continuously. Seas readvanced quickly northwestward across Texas, northward into southern New Mexico, and westward into western Oklahoma and southeastern Colorado. Silt and mud were deposited several thousand feet thick in the geosyncline. The same kinds of sediment spread across the bordering craton early in the interval; later, limestone was deposited on the craton. The west end of the Anadarko basin and the ancestral Delaware basin both continued to sink moderately. As much as 2,000 feet of black shale accumulated in the Delaware basin, west Texas, compared to an estimated 100–300 feet of shale and limestone in a belt that extended from the Llano region northwestward across Texas into east-central New Mexico, and about 1,300 feet of shale and limestone in the Anadarko basin in the Texas Panhandle.

Near the end of the Mississippian Period, the cratonic area became slightly positive and the sea withdrew from the interior region. The Ouachita geosyncline remained negative, and there deposition continued into Pennsylvanian time.

REGION DEFINED

The Southern Midcontinent and Southern Rocky Mountains region includes all of Texas and New Mexico, the Panhandle and extreme southwestern corner of Oklahoma, and southeastern Colorado generally east of long 105°30'W. (fig. 31). The frontal ranges of the southern Rocky Mountains lie along the western boundary of the region in Colorado. The Rocky Mountain system terminates with the Sangre de Cristo Mountains in north-central New Mexico. A chain of generally north trending mountain ranges extends from there southward in a broad belt across New Mexico into extreme west Texas (fig. 32).

Mississippian rocks crop out (pl. 1) on the flanks of the Llano region in central Texas and in some of the mountain ranges of west Texas, central New Mexico, and central Colorado. These include the Glass, Van Horn, Hueco, and Franklin Mountains in west Texas; the Big Hatchet, Peloncillo, Caballo, San Andres, and Sacramento Mountains and Black Range in southern New Mexico; the Sierra Nacimiento and Sandia and Sangre de Cristo Mountains in northern New Mexico; and the Wet Mountains and Rampart Range in Colorado (fig. 32). Elsewhere in the region Mississippian rocks are deeply buried and are known only from drilling.

In the subsurface, Mississippian rocks extend continuously in a west-trending very irregular belt across

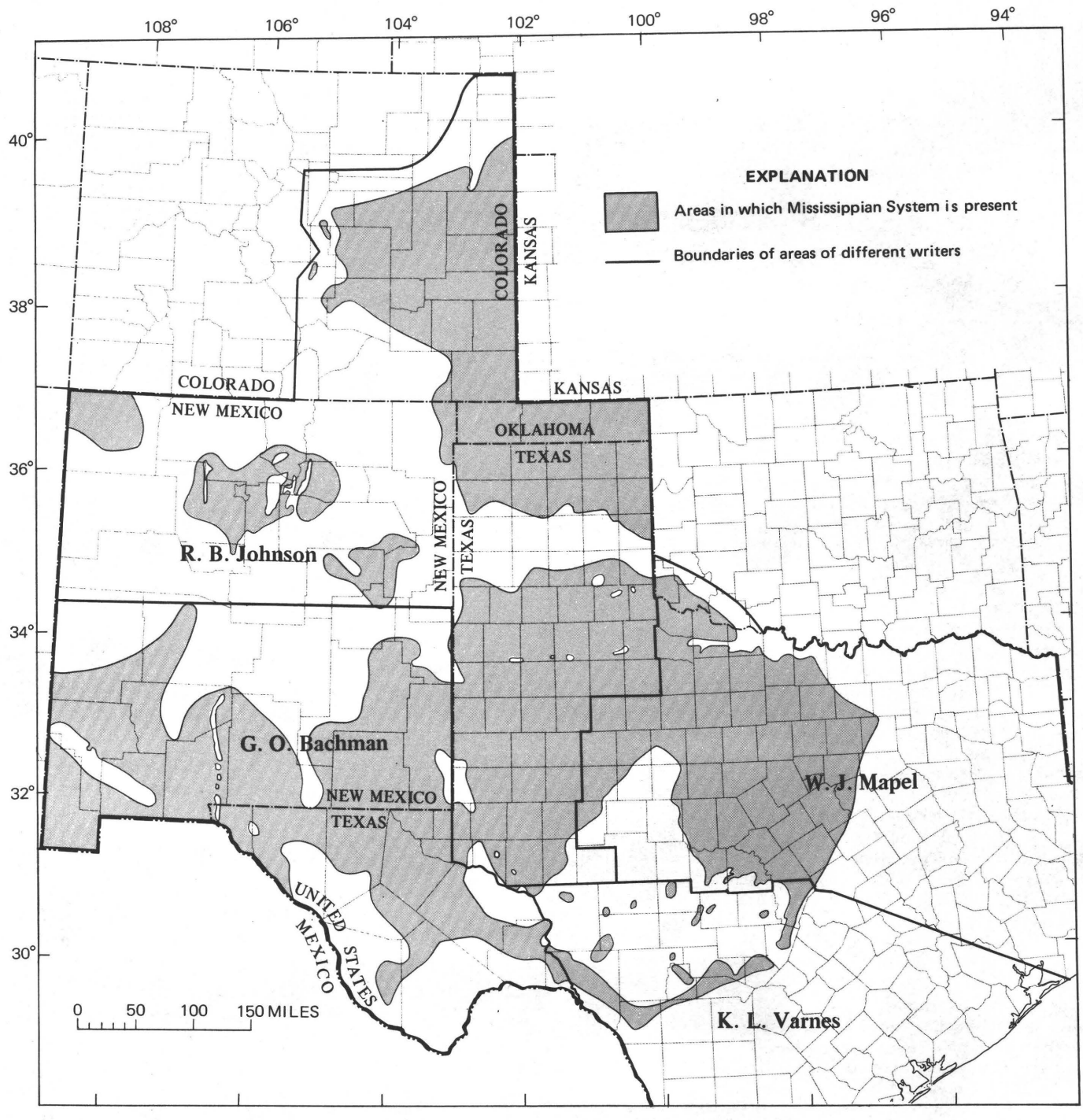

FIGURE 31. — Distribution of Mississippian rocks in the Southern Midcontinent and Southern Rocky Mountains region and areas studied by different writers.

central Texas and the southern half of New Mexico (fig. 31). Three outlying areas that are underlain by Mississippian are scattered across northern New Mexico. Mississippian rocks occur also in the subsurface in the panhandles of Texas and Oklahoma and in an irregular area in eastern Colorado. The rocks in these last named areas compose part of a much more extensive sheet of Mississippian rocks that continues northward and eastward across neighboring Kansas and Oklahoma.

The region includes several large well-known post-Mississippian depositional basins separated by diversely oriented platforms, arches, domes, and other uplifts as shown by figure 33. Almost all these features owe their

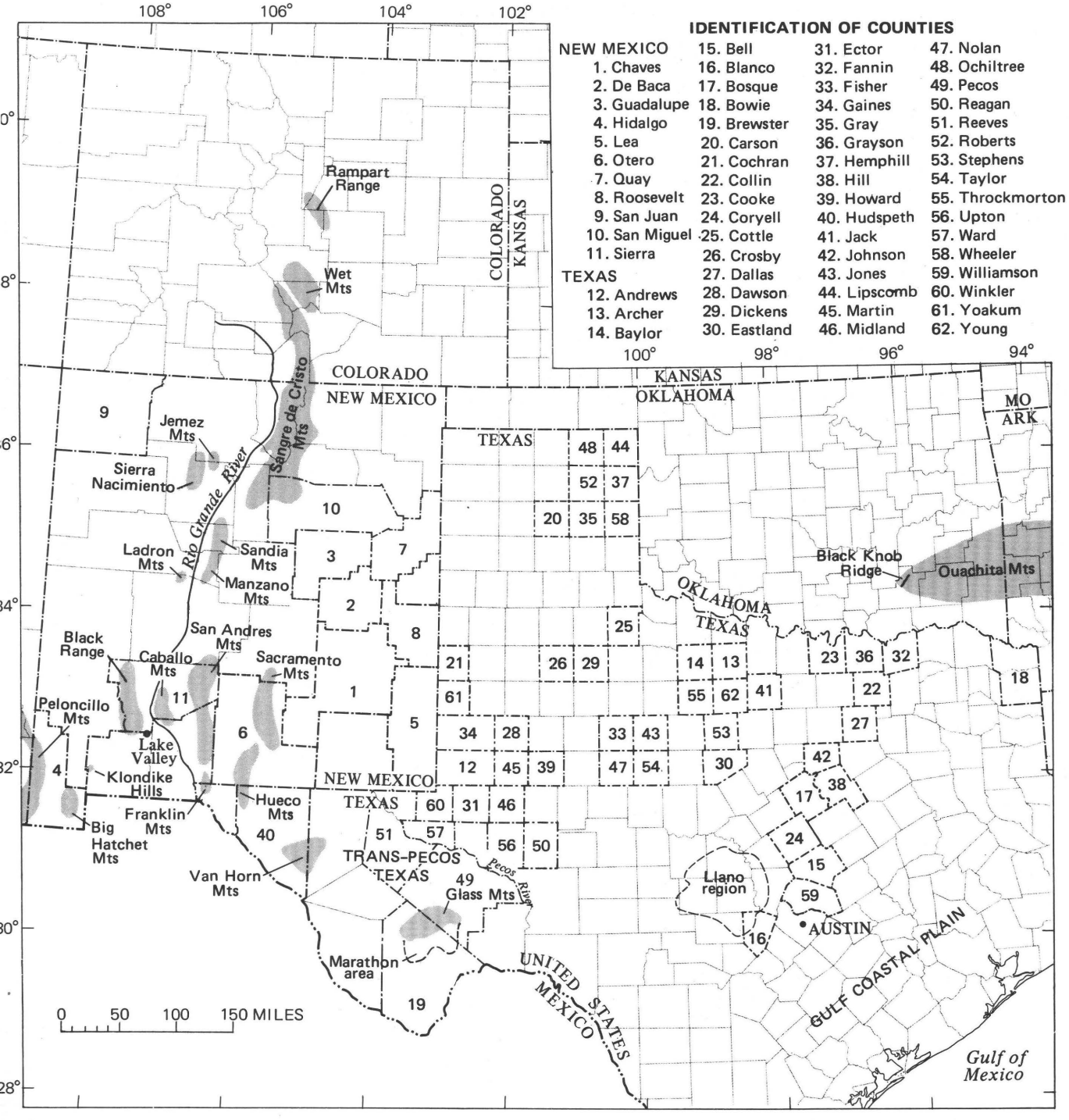

FIGURE 32. — Localities and geographic features in Southern Midcontinent and Southern Rocky Mountains region referred to in text.

resent configuration to structural movements that occurred after Mississippian time; they therefore had no influence on Mississippian sedimentation. Several of them, however, affect the present distribution of the system because of widespread post-Mississippian erosion. Among the major uplifts that interrupt the continuity of the Mississippian in the subsurface are the Amarillo and Wichita uplifts and Muenster arch in northern Texas and southern Oklahoma; the Matador and Red River arches in the Texas Panhandle and along the Oklahoma-Texas State line; and the Central Basin platform, Pecos arch, and Llano uplift in southwestern and south-central Texas. Large sedimentary basins included or partly included in the region are the

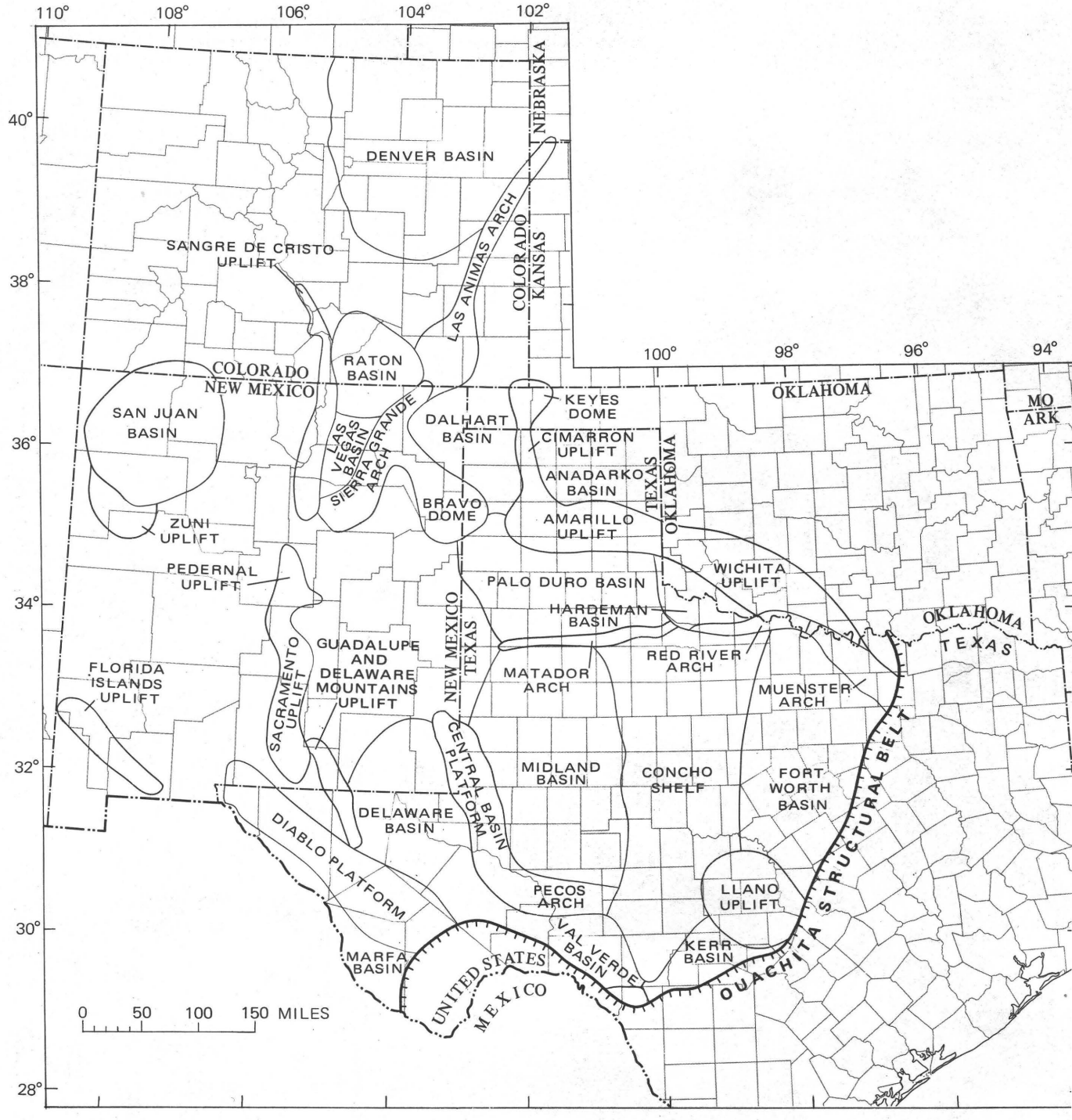

FIGURE 33. — Some post-Mississippian structural elements in the Southern Midcontinent and Southern Rocky Mountains region.

Denver, Dalhart, and Anadarko basins in the north and northeastern part of the region, the Fort Worth, Palo Duro, Midland, and Delaware basins in the central part, and the San Juan basin in northwestern New Mexico.

The structural pattern shown by figure 33 was superimposed on a much simpler Mississippian structural pattern, shown by figure 34. The principal Mississippian structural features were the Ouachita geosyncline, which formed in Late Mississippian time in the area of the present-day Ouachita structural belt and counterparts of the post-Mississippian Delaware and Anadarko basins in the south-central and northeastern parts of the region, respectively. At the beginning of the Mississippian a low broad arch, the Texas arch, not recognizable in post-Mississippian time, extended from central Texas northwestward into north

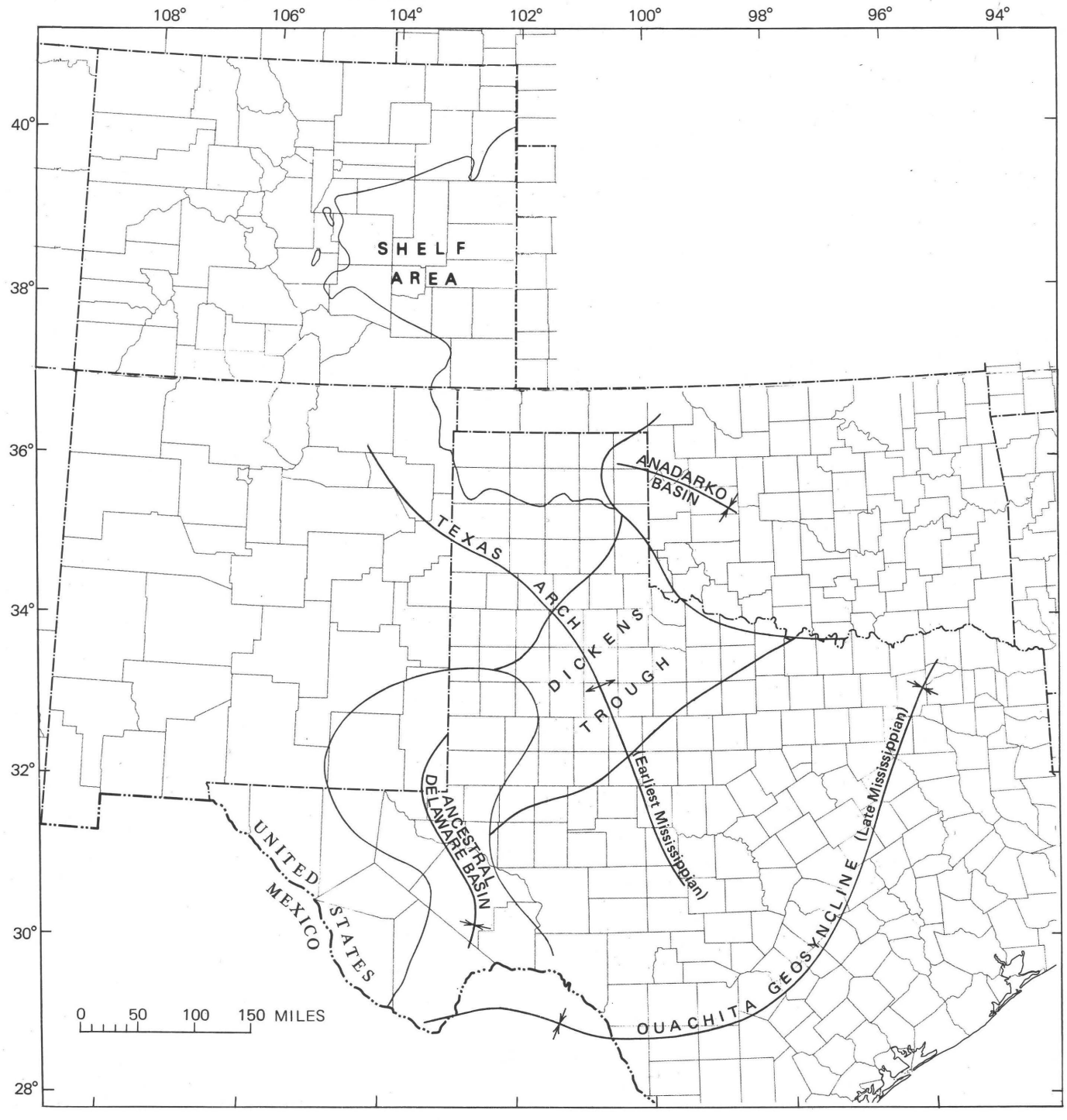

FIGURE 34. — Structural elements of Mississippian age in the Southern Midcontinent and Southern Rocky Mountains region.

eastern New Mexico. During most of the Mississippian, shifting, indefinitely located areas at the ends of the Texas arch were areas of retarded subsidence or were slightly positive. Early in the Mississippian, however, the Anadarko and ancestral Delaware basins were connected by a shallow, intermittently active structural depression that developed as a saddle across the central part of the Texas arch. This depression is referred to in figure 34 as the Dickens trough, named for Dickens County, Texas, which is on the axis.

The front of the buried Ouachita structural belt determines the eastern and southern extent of detailed mapping of Mississippian rocks for this investigation. However, Mississippian as well as Devonian rocks are known from a few scattered boreholes to the east and south of this belt and may well occur over a large area

beneath the thick younger deposits of the Gulf Coastal Plain. The Ouachita structural belt extends across eastern and southern Texas from Grayson County southwestward to eastern Blanco County and from there generally westward to the Marathon area in Brewster County (fig. 33; Flawn and others, 1961, pl. 2). Scattered drill holes and outcrops in the Marathon area in west Texas show Mississippian rocks telescoped in folds and fault slices along the front of the structural belt; within the belt, however, information adequate for any regional stratigraphic analysis is lacking.

The boundaries of intervals and formations referred to in describing the Mississippian rocks in the Southern Midcontinent and Southern Rocky Mountains are shown on plate 15.

PALEOGEOLOGY

UNITS UNDERLYING THE MISSISSIPPIAN

Precambrian igneous and metamorphic rocks underlie the Mississippian in parts of northern and east-central New Mexico and in a broad lobe that narrows southeastward in the Texas Panhandle (pl. 2). The *Arbuckle Limestone* of Late Cambrian and Early Ordovician age or the Ellenburger Group of Early Ordovician age is found directly beneath the Mississippian in a broad area in central Texas, eastern Colorado, and central eastern New Mexico. Undifferentiated Silurian rocks, partly correlative with the Fusselman Dolomite of south-central New Mexico, occur directly below Mississippian rocks north of the Central Basin platform in Texas and New Mexico in an arcuate area concave to the south. Subcrops of the Ordovician *Simpson Group* and *Viola Limestone* and the Silurian and Devonian *Hunton Group* form parallel northwest-trending bands that broaden northward in north-central Texas and in the Texas and Oklahoma Panhandles.

At the isolated localities of Mississippian outcrop in south-central Colorado, the units underlying the Mississippian System are Precambrian granite and gneiss, the Upper Cambrian Peerless Formation, the Lower Ordovician Manitou Dolomite, the Middle Ordovician Harding Sandstone, and the Middle and Upper Ordovician Fremont Limestone.

Devonian rocks underlie the Mississippian in much of the southern part of the region. These are referred to the *Woodford Formation* of Late Devonian and Early Mississippian age (Ellison, 1950, p. 14) in south-central Texas and southeastern New Mexico and to Devonian rocks, undifferentiated on plate 2, that include from base to top the *Onate, Sly Gap,* and Percha Formations in extreme west Texas and southwestern New Mexico.

In northern New Mexico, Paleozoic rocks locally found directly beneath the Mississippian are assigned to the Ouray Limestone of Devonian age (Cooper, 1955) in the San Juan basin and to the Espiritu Santo Formation of Devonian(?) age (Baltz and Read, 1960) in the Sangre de Cristo Mountains.[1]

In the Llano uplift, cracks and collapse structures on the eroded surface of the Ellenburger Group locally contain thin remnants of phosphatic rocks transitional from Devonian to Mississippian and assigned by Cloud, Barnes, and Hass (1957) to the Houy Formation. These rocks are the central Texas equivalent of an Upper Devonian–Lower Mississippian phosphatic black shale sequence that is widespread in the Central and Eastern United States (Barnes and others, 1947; Cloud and others, 1957).

Rocks immediately below the Mississippian System in the eastern part of the Ouachita structural belt, where the contact is not a thrust fault, belong to the Arkansas Novaculite of Devonian age; or, in the Marathon area, to the approximately correlative Caballos Novaculite (Flawn and others, 1961, p. 50–51).

LOWER BOUNDARY OF THE MISSISSIPPIAN

Fossils of Early Mississippian (early Kinderhook) age occur in the topmost part of the Woodford Formation in south-central Oklahoma (Hass and Huddle, 1966). Likewise, at some places in the Llano region, the topmost 3–4 feet of phosphorite and black shale in the Houy Formation contains conodonts that are considered early Kinderhook in age (Cloud and others, 1957, p. 813). Both formations are mainly Late Devonian in age, however, and, in the subsurface in the Midland basin and nearby parts of west Texas, the *Woodford* is reported to be entirely Devonian in age (Ellison, 1950, p. 14). Mississippian parts of these formations, where present, could not be differentiated on a lithologic basis from underlying Devonian parts, and for this reason all the *Woodford* and Houy Formations were excluded from the Mississippian System in this report. Similarly, possible Lower Mississippian (Kinderhook) rocks that may be present at the top of the Caballos Novaculite in west Texas and the *Arkansas Novaculite* in east Texas are not differentiated from underlying Devonian parts of the formations, and the Caballos and *Arkansas* Novaculites are treated as though they were entirely Devonian. Regional stratigraphic relations indicate that the contact of the Houy, *Woodford,* Caballos, and *Arkansas* Formations with overlying Mississippian rocks is an unconformity representing at least some part of the Kinderhook Series.

[1] Armstrong (1967, figs. 4, 5) and Armstrong and Mamet (1974, fig. 2), on the basis of endothyrid identifications assigned the Espiritu Santo to the late Osage.

In the San Juan basin, northwestern New Mexico, the Ouray Limestone, which is mostly of Devonian age (Cooper, 1955), may be transitional into Early Mississippian (Knight and Baars, 1957). Any stratigraphic break at the contact of the Ouray with overlying Mississippian rocks is obscure, but many geologists regard the contact as marking a hiatus of regional importance (Parker and Roberts, 1966, p. 2416). The Ouray Limestone has been excluded arbitrarily from this study because of the difficulty of separating any thin Mississippian part of the formation.

Except where lowermost Mississippian rocks may locally occur at the top of the *Woodford*, Houy, Ouray, Caballos, and *Arkansas* Formations, an unconformity everywhere separates Mississippian from older rocks. Not only are rocks of widely different ages directly under the Mississippian (pl. 2), but the age of the basal Mississippian deposit varies from place to place (fig. 35). The duration of the hiatus represented by the unconformity is greatest in the subsurface in east-central New Mexico and locally in the Rampart Range in Colorado where Mississippian rocks of Meramec age (interval C) unconformably overlie the Precambrian.

In the vicinity of Black Knob Ridge, which lies at the west end of the Ouachita Mountains of south-central Oklahoma (fig. 32), the predominately Mississippian Stanley Shale may rest unconformably on a part of the Arkansas Novaculite that is correlative with the Chattanooga Shale or Woodford Formation of central and northern Oklahoma (Hendricks and others, 1937, p. 11–12; Hendricks and others, 1947). Presumably, the two formations are also unconformable in the subsurface to the southwest in the Ouachita structural belt in Texas. Conodonts reported by Hass (1950; 1956, p. 29–32) from near the base of the Stanley Shale in the Black Knob Ridge area and elsewhere in the Ouachita Mountains of Oklahoma and Arkansas are forms found in rocks of Chester age (interval D) in the Illinois basin (Rexroad and Clarke, 1960, p. 1205; Rexroad and Collinson, 1961; Collinson and others, 1962), which seems to indicate that the unconformity at the base of the Stanley in southwestern Oklahoma and nearby parts of Texas represent most or all of pre-Chester (pre-interval D) Mississippian time (pl. 15, col. 79).

Data supporting the conclusions just summarized regarding the age of the Stanley Shale and the stratigraphic relations with the Arkansas Novaculite are discussed in a later part of this chapter; it should be noted, however, that some geologists (Elias, 1956, 1959, 1966; Maher, 1960, p. 114; Goldstein, 1961; Cline, in Cline and Shelburne, 1959, p. 180; Branson, 1959b, p. 119) have considered the base of the Stanley to be early Meramec (interval C) in age, or older. Frezon and Glick (chaps. I, H, present rept.) discuss other interpretations of the relations at the base of the Stanley Shale in Oklahoma and Arkansas.

In the Marathon area, the Mississippian Tesnus Formation contains conodonts of Late Mississippian (Chester) age within about 90 feet of its base (S. P. Ellison, Jr., written commun., 1967). The contact of the Tesnus and underlying Caballos Novaculite presumably is a hiatus of about the same duration as the one represented at the Stanley Shale–Arkansas Novaculite contact elsewhere along the Ouachita structural belt to the northeast.

PALEOTECTONIC IMPLICATIONS

A broad, generally northwest trending arch, called by Adams (1954; 1962, fig. 1) the Texas peninsula or Texas arch, is the dominant structural feature affecting the pre-Mississippian paleogeologic map (pl. 2; fig. 34). The axis of this uplift crosses the region with a curving trace from northeastern New Mexico to the Llano region in central Texas. A north-trending syncline in pre-Devonian rocks, open to the south, flanks the Texas arch on the southwest in west Texas and adjacent parts of New Mexico. The northeast flank of the Texas arch dips gently northeastward toward a broad basinal area in Oklahoma and Kansas.

Barnes, Cloud, and Warren (1947, p. 140) pointed out that the presence of Devonian rocks on different parts of the Ellenburger Group in the Llano region dates the main truncation of the Ellenburger in central Texas as pre-Early Devonian. Silurian rocks are involved in the folding on the flanks of the Texas arch. The age of the folding that produced the Texas arch and the flanking structural features, therefore, seems rather closely bracketed as latest Silurian or earliest Devonian.

In Late Devonian time the region was submerged and the *Woodford Formation* and equivalent rocks were deposited in a sea that transgressed the region from the south and southeast. Scattered remnants of the Houy Formation in the Llano region, far removed from the main body of the equivalent parts of the *Woodford Formation*, show that Upper Devonian rocks were once considerably more widespread in central Texas than at present. It seems likely that they connected directly with similar deposits of the Woodford in Oklahoma. If so, a shallow epicontinental sea — in which the Woodford, Houy, Percha, Ouray, and correlative black shale and limestone were deposited — occupied most of Texas and New Mexico (Ellison, 1950, fig. b; Conant and Swanson, 1961, p. 51–52).

Deposition of black shale or limestone continued in some areas after the end of Devonian time, but, in general, the Mississippian Period was initiated by epeirogenic uplift and a general withdrawal of the sea.

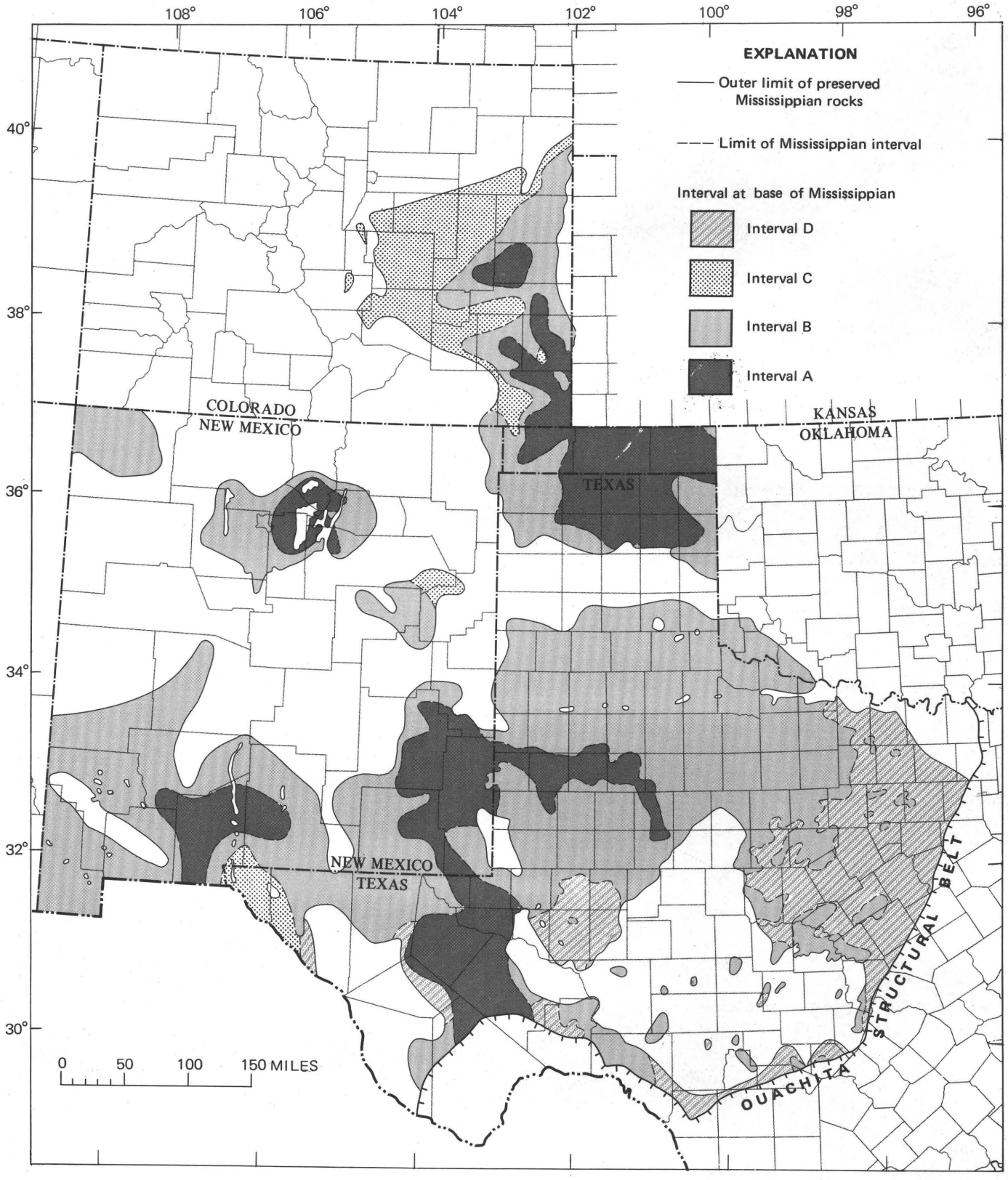

FIGURE 35. — Distribution of intervals at the base of the Mississippian System in the Southern Midcontinent and Southern Rocky Mountains region.

Subsequent dissection was not very deep, although erosion was sufficient to strip away Upper Devonian rocks from the crest and from part of the north flank of the Texas arch and from parts of northern New Mexico. Within a relatively short interval, represented by a limited part of early or middle Kinderhook time, the sea readvanced, and the deposition of the main body of Mississippian rocks began.

INTERVAL A

FORMATIONS INCLUDED

Named stratigraphic units included in interval A (pl. 15) are the *Caballero Formation* in the San Andres and Sacramento Mountains of southern New Mexico, the lower part of the Keating Formation in southwestern New Mexico, and the Macho Member of the Tererro Formation[2] in the Sangre de Cristo Mountains of northern New Mexico. Elsewhere, interval C consists of unnamed rocks presumed to be Kinderhook in age. Fairly large areas of such rocks are present (pl. 3-A) in the Oklahoma and Texas Panhandles and along the Colorado-Kansas State line north of the Amarillo uplift, in the northern parts of the Midland and Delaware basins, New Mexico and Texas, and in trans-Pecos Texas. Smaller areas are found in north-central and south-central New Mexico. Fossils useful for dating have not been reported from rocks in interval A in most of these areas.

Unnamed rocks assigned to the upper part of interval A north of the Amarillo uplift are continuous with the Gilmore City Limestone and Sedalia(?) Dolomite of Kinderhook age in southwestern Kansas (Goebel, 1966), and they occupy the stratigraphic position of the Welden Limestone in Oklahoma.

The rocks assigned to interval A south of the Amarillo uplift in Texas and New Mexico have the same lithology and stratigraphic position as those north of the arch and are considered Kinderhook on this basis.

The Leadville Limestone of Kinderhook and Osage age occurs in northwestern New Mexico. For convenience in mapping, we have included all of this formation in interval B.

Interval A is not shown within the Ouachita structural belt on the isopach and facies maps (pls. 3-A, 3-B). Rocks of Kinderhook age make up the upper 30 feet of the middle division of the Arkansas Novaculite in Montgomery County, western Arkansas (Hass, 1956, p. 29). About 100 miles to the west, however, in the western part of the Ouachita Mountains, southern Oklahoma, the Arkansas Novaculite is about a third as thick as it is in Montgomery County, Ark. Beds of Devonian age are present, but the upper division of Arkansas Novaculite of Montgomery County cannot be recognized (Hendricks and others, 1947; Ham, 1959, p. 75). It is not certain, therefore, that the Arkansas Novaculite contains any Mississippian rocks in the western part of the Ouachita Mountains or in the frontal zone of the Ouachita structural belt in Texas to the south. Although the uppermost part of the Caballos Novaculite is assigned a Mississippian age (Flawn and others, 1961, p. 51) rocks of Mississippian age cannot be recognized as a unit separable from the underlying Devonian rocks in the upper part of the formation where it is exposed in the Marathon area; hence, interval A is not mapped there.

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A in Texas and southeastern New Mexico west of the Ouachita structural belt is a gradational contact between rocks that are mostly green mudstone and siltstone and overlying rocks that are primarily limestone but that locally contain some thin beds of mudstone. This lithologic change is a convenient boundary for mapping; however, there is no paleontologic information to show that it corresponds to any series boundary. Some geologists (for example. R. V. Hollingsworth, Paleontological Laboratory, Midland, Tex., unpub. reports for Andrews, Martin, and Howard Counties, Tex.) have placed the Kinderhook-Osage boundary somewhat higher within the limestone sequence, but a boundary within the limestone is not a mappable interval boundary on the sample logs available for this investigation. North of the Amarillo uplift in Texas, Oklahoma, and eastern Colorado, the upper contact of interval A appears to be gradational from sandy, silty, or muddy limestone and dolomitic limestone in interval A to crinoidal, locally argillaceous, limestone or dolomitic limestone that commonly contains abundant white and gray chert in interval B.

In north-central New Mexico the Manuelitas Member of the Tererro Formation of interval B lies unconformably on the Macho Member of interval A. At places the lowest beds of the Manuelitas have collapsed into sinkholes in the Macho (Baltz and Read, 1960, p. 1760).[3] In outcrops in the San Andres and Sacramento Mountains in south-central New Mexico, the *Caballero Formation* is conformably overlain by the Lake Valley Limestone of interval B.

[2] Armstrong (1967, figs. 4, 5), and Armstrong and Mamet (1974, fig. 2, p. 149) assign an early Meramec age to the Macho Member of the Tererro Formation based on endothyrid identifications.

[3] For alternative explanations of these features, see Sutherland and Land (1959), Sutherland (1963), Armstrong (1967), and chapter K of this report.

In extreme southwestern New Mexico, member A of the Keating Formation contains a fauna of very latest Kinderhook or earliest Osage age and is transitional upward into member B (Armstrong, 1970, p. 62).

THICKNESS TRENDS

Strata of interval A are thin; they reach a thickness of 200 feet or more at only a few localities: three in Chaves and Roosevelt Counties, southeastern New Mexico; one in the Big Hatchet Mountains, Hidalgo County, N. Mex.; one in northern Roberts and southern Ochiltree Counties, Tex.; and one in Ward County, Tex. Shaly limestone or green shale compose the base of Mississippian rocks in the Midland and Palo Duro basins in a few wells beyond the outlines shown for interval A on plate 3-A. These occurrences are arbitrarily included with interval B as being too thin and discontinuous to map separately. Interval A in north-central New Mexico is interpreted here as an isolated, erosional remnant that probably was once somewhat thicker and much more extensive. No regional pattern of thickness change is apparent.

LITHOFACIES TRENDS

Rocks assigned to interval A in southeastern Colorado and the Oklahoma Panhandle are predominantly dolomite interbedded with thin beds of oolitic limestone, some sandy limestone, and some thin mudstone. The basal part of interval A in this area is composed of dolomitic sandstone that contains fragments of weathered chert (Maher and Collins, 1949).

In New Mexico interval A is mainly limestone (pl. 9-F, secs. *I–I'*, *J–J'*) interbedded with subordinate green and gray siltstone and mudstone. In the Sangre de Cristo Mountains the Macho Member of the Tererro Formation is assigned to interval A following the interpretation of Baltz and Read (1960). An alternate interpretation based on foraminiferal determinations (Armstrong, 1967, p. 5–8; Armstrong and Mamet, 1974, fig. 2) assigns these beds along with the Espiritu Santo Formation entirely to Osage and Meramec age. (On plate 15, col. 88 presents the interpretation of Baltz and Read; table 3 (in chap. K) presents the interpretation of Armstrong.) The carbonate rocks of the Macho Member are largely limestone breccia, formed by collapse of parts of the underlying beds into a cavernous karst surface (Baltz and Read, 1960, p. 1768), or formed as a basal comglomerate by erosion in an advancing sea (Sutherland and Land, 1959), or formed as a collapse breccia resulting from removal of gypsum (Armstrong, 1967, p. 21). In south-central New Mexico the *Caballero Formation* of the San Andres and Sacramento Mountains and the Black Range is composed of gray nodular argillaceous limestone and subordinate gray calcareous mudstone (Laudon and Bowsher, 1949; Kottlowski and others, 1956). The nodules are generally less than 3 inches long and are elongated parallel to the bedding. This lithology is uniform and distinctively different from that of adjacent beds. Some sandy mudstone is present in the *Caballero Formation* in the Black Range. In the Big Hatchet Mountains of extreme southwestern New Mexico, interval A is made up entirely of limestone (Zeller, 1965).

Whereas carbonates dominate interval A rocks in southeastern Colorado, the Oklahoma Panhandle, and all New Mexico, sandstone, mudstone, and siltstone become predominant over the limestone in the Texas Panhandle and locally in west-central Texas (pl. 9, secs. *I–I'*, *J–J'*). In trans-Pecos Texas the carbonates are dominant over the detrital rocks.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Silt and clay in interval A in Texas and New Mexico are probably reworked insoluble residues and other clastic material derived from weathering of Paleozoic limestone and dolomite or Precambrian igneous and metamorphic rocks exposed in low-lying land areas (pl. 11, fig. 1). The general lack of sand or coarser material and the presence of interbedded limestone indicates either absence of coarse material in the source areas, remoteness from land areas having any great relief, quiet water lacking currents of any great force, or some combination of all three.

The paleogeographic map of interval A for the Southern Midcontinent and Southern Rocky Mountains region (pl. 12, fig. 1) shows two narrow shallow arms of the sea reaching from Mexico into Texas and New Mexico. Shallow marine waters that covered the central part of the continent gradually extended themselves into the region from the north.

The Texas arch was slightly above sea level and provided some detrital materials to the ancestral Anadarko basin in Oklahoma and the northern part of the Texas Panhandle. The Texas arch and most of New Mexico were flooded by the sea late in interval A or early in interval B.

PALEOTECTONIC IMPLICATIONS

The rocks of interval A in southeastern Colorado, the Oklahoma Panhandle, Texas, and New Mexico were deposited on a tectonically stable area of low relief near sea level. The outlines of local basins of deposition during interval A may have been determined more by topography inherited from Late Devonian time than by local downwarping during the deposition of interval A rocks.

As interpreted here, the controversial breccias of the Macho Member of the Tererro Formation in the Sangre

de Cristo Mountains probably originated as limestone in a Devonian(?) Espiritu Santo sea. Late in the Devonian or early in the Mississippian the seas withdrew as a result of a slight elevation of the area of deposition within and south of the present Sangre de Cristo Mountains. This uplift was sufficient for the formation of sinkholes and caverns into which the limestone now forming the breccias of the Macho Member locally collapsed during Kinderhook time (pl. 10, fig. 1; pl. 11, fig. 1).

The Texas arch was a slightly positive element in interval A. The interval closed with an overall sinking of the Texas region relative to sea level, including the area of the Texas arch.

INTERVAL B

FORMATIONS INCLUDED

Rocks of interval B are mostly Osage in age, although older and younger Mississippian rocks may be included locally where contacts are gradational and the age relations are not clearly worked out (pl. 15). Named stratigraphic units included in interval B in New Mexico are the Leadville Limestone in San Juan County (fig. 32), Manuelitas Member of the Tererro Formation (Baltz and Read, 1960) in the Sangre de Cristo Mountains; Arroyo Penasco Group in the Sierra Nacimiento and northern Sandia Mountains; *Caloso Formation* in the Ladron Mountains and nearby areas (Kelley and Silver, 1952, p. 86–87); Lake Valley Limestone and locally the Kelly Limestone in the San Andres Mountains and nearby areas; and the upper part of the Keating Formation and lower part of the Hachita Formation of the Escabrosa Group in southwestern New Mexico. The Arroyo Penasco Group (Armstrong, 1955; Fitzsimmons and others, 1956, p. 1940–1942; Armstrong, 1967, p. 22–24; Armstrong and Mamet, 1974, p. 149) contain beds of Meramec and Chester age as well as late Osage age, but the entire unit is here included in interval B. The age relations of formations in southern New Mexico were discussed by Laudon and Bowsher (1949) and Armstrong (1958b; 1962). Rocks included in interval B in southern New Mexico are fairly well dated as Osage in age.

In the Oklahoma Panhandle, the northern part of the Texas Panhandle, and in eastern Colorado, rocks of interval B include undifferentiated equivalents of the Keokuk and Burlington Limestones and perhaps rocks the age of the *Fern Glen Limestone* of Kansas (Maher and Collins, 1949).

Interval B consists of the Chappel Limestone, as described by Cloud and Barnes (1946, p. 49–50) in outcrops in the Llano region (fig. 32), central Texas, and most of the rocks that are continuous or virtually continuous with the Chappel in the Fort Worth basin (fig. 33) and in adjoining areas to the west. In the northwestern part of the Fort Worth basin, on the northern part of the Concho shelf, and in the Palo Duro and Hardeman basins, the Chappel is divided into two units, and only the lower or cherty unit is assigned to interval B. Both units have been called in local subsurface terminology the Chappel Limestone, the Mississippian lime, or the Osage-Meramec limestone. The part of the limestone assigned to interval B corresponds closely in thickness and appears to be continuous with limestone that Culp (1961, p. 10–11) assigned to the Osage Series in the Oklahoma part of the Hardeman basin (or Hollis basin of Culp's terminology). Culp's (1961, p. 11) correlation was based on lithologic similarity to rocks in the Anadarko basin which he called pre-Sycamore limestone.

Roundy and Girty (in Roundy and others, 1926, p. 3, 6–8) considered the Chappel Limestone of the Llano region Osage in age. Weller and others (1948, p. 143–144) considered it partly Kinderhook and partly Osage. Cloud and Barnes (1946, p. 50–51) and Hass (1959, p. 367) assigned the major part of the formation in outcrops to the late Kinderhook, and the uppermost part locally to the early Osage. Branson (1959a) regarded the Chappel in the Llano region as Kinderhook in age.

Paleontological information is insufficient to show the age of the Chappel Limestone or its lithologic equivalents in the subsurface north of the Llano region. The Chappel thins southward in the southern part of the Fort Worth basin partly because of erosion and truncation beneath overlying Mississippian rocks. The Chappel of the outcrop belt accordingly is equivalent only to some lower part of the Chappel in areas to the north, and the formation is presumed to be largely, if not entirely, younger than Kinderhook where it is fairly thick in the Fort Worth basin, on the Concho shelf, and in the Palo Duro and Hardeman basins. Parts of the Chappel that might be Kinderhook in age, and might properly be included in interval A, could not be distinguished from younger parts. For this reason, the base of interval B is placed at the base of the Chappel Limestone.

The limestone sequence referred to in many wells as the Osage-Meramec limestone is mapped entirely in interval B in southeastern New Mexico and most of trans-Pecos Texas. No lithologic or other criteria were developed to distinguish consistently rocks of Osage from those of Meramec age in this sequence, or for otherwise subdividing it. Thickness trends and lateral continuity with Mississippian rocks in adjacent areas suggest that these rocks are Osage in age at most places, and they are assigned to interval B on this basis. Rocks of possible Meramec age that in the Midland basin were assigned to interval C include the top

100–200 feet of limestone in the Osage-Meramec limestone where it is thickest in northern Lea County, N. Mex. Rocks of Meramec age may be present, also, where the Osage-Meramec limestone is locally thick in central Pecos County, Tex.

Available information indicates that rocks referable to interval B are absent in the Ouachita structural belt.

UPPER BOUNDARY OF INTERVAL B

A contact between locally crinoidal limestone or dolomite which contains much white and grayish-white chert, and overlying limestone, which is much less cherty and locally oolitic, is the boundary between intervals B and C in eastern Colorado and in the Dalhart, Anadarko, and Palo Duro basins in the Texas and Oklahoma Panhandles. Sample logs examined in these areas generally show this lithologic boundary as the top of the Osage Series. Maher and Collins (1949) stated that the contact is an unconformity. Cross sections later given by Maher (1950), however, show the contact as conformable in the subsurface in eastern Colorado. Interval B and the lower part of interval C are described as an onlapping sequence deposited in a sea that transgressed from east to west across the western part of the Dalhart basin and on the Sierra Grande and Las Animas arches in New Mexico and Colorado (Maher, 1950).

In the Hardeman basin and adjacent northern part of the Concho shelf, the top of interval B lies within a conformable limestone sequence characterized in the lower part by chert and in the upper part by oolites. The upper boundary of interval B in these areas is chosen to exclude the main body of oolitic limestone. Paleontologic confirmation that the boundary in any of these areas is the Osage-Meramec contact is lacking, but this boundary is convenient for mapping, and it appears to be a fairly consistent stratigraphic horizon in the area studied.

In the Sangre de Cristo Mountains, northern New Mexico, the contact of the Manuelitas Member of the Tererro Formation with the overlying Cowles Member of interval C is marked by an erosional unconformity (Baltz and Read, 1960) and by limestone-pebble conglomerate, marly shale, and cross-laminated calcarenite in the beds immediately above the contact. The Cowles Member of the Tererro Formation was not involved in the collapse into sinkholes which distinguishes the underlying members.

At exposures in south-central New Mexico the upper boundary of interval B is a sharp contact with overlying rocks of interval C. Laudon and Bowsher (1949, p. 19) described this contact in the southern San Andres and Sacramento Mountains as a northward overlap of rocks of Meramec age on a surface eroded on the Lake Valley Limestone.

In southwestern New Mexico the boundary at the top of interval B is a gradational contact within the Hachita Formation of the Escabrosa Group. This horizon corresponds to the Osage-Meramec boundary as shown by Armstrong (1962, figs. 2, 3, p. 10–13).

In southeastern New Mexico and trans-Pecos Texas where the so-called Osage-Meramec limestone (assigned in this area entirely to interval B) is overlain by younger Mississippian rocks, the contact is a disconformity marked by an abrupt change from limestone to dark-gray shale of interval D.

The upper boundary of interval B in the Llano region, in much of the Fort Worth basin, in the western part of the Midland basin, and on the Pecos arch, is the contact of limestone in the Chappel Limestone with overlying dark-gray to black shale of the Barnett Formation. An unconformity at this horizon in the Llano region (Cloud and Barnes, 1946, p. 110–111; Plummer, 1950, p. 21) presumably represents some part of Osage and most or all of Meramec time.

As shown in figure 36, interval B is at the top of the Mississippian System in several large and irregularly shaped areas. In these areas, it is generally overlain unconformably by rocks of Pennsylvanian age, and locally by rocks of Permian age. Shale, locally arkosic sandstone, or limestone of Pennsylvanian age unconformably overlie cherty limestone of interval B on much of the Concho shelf, in narrow bands on both flanks of the Amarillo uplift, along the south flank of the Red River arch, and in parts of west Texas and central and northwestern New Mexico. Sandstone and shale of Permian (Wolfcamp) age may overlie interval B locally in Hudspeth County, Tex., and adjacent parts of Otero County, N. Mex., and Permian rocks lie on rocks of interval B locally in Winkler, Ector, and Ward Counties, Tex. (fig. 32).

THICKNESS TRENDS

Interval B is present in its originally deposited thickness only where it is conformably overlain by interval C, as in eastern Colorado, western Oklahoma and nearby parts of north Texas, and in southeastern New Mexico. Elsewhere, thicknesses are considerably modified by postdepositional erosion.

Interval B attains a thickness of 760 feet in the Anadarko basin (pl. 4-A) in northeastern Hemphill County, Tex., and thins generally westward from this area across the Panhandles of Oklahoma and Texas to an irregular zero line near the Texas–New Mexico State line. Farther south the interval thins irregularly westward or northward from slightly more than 400

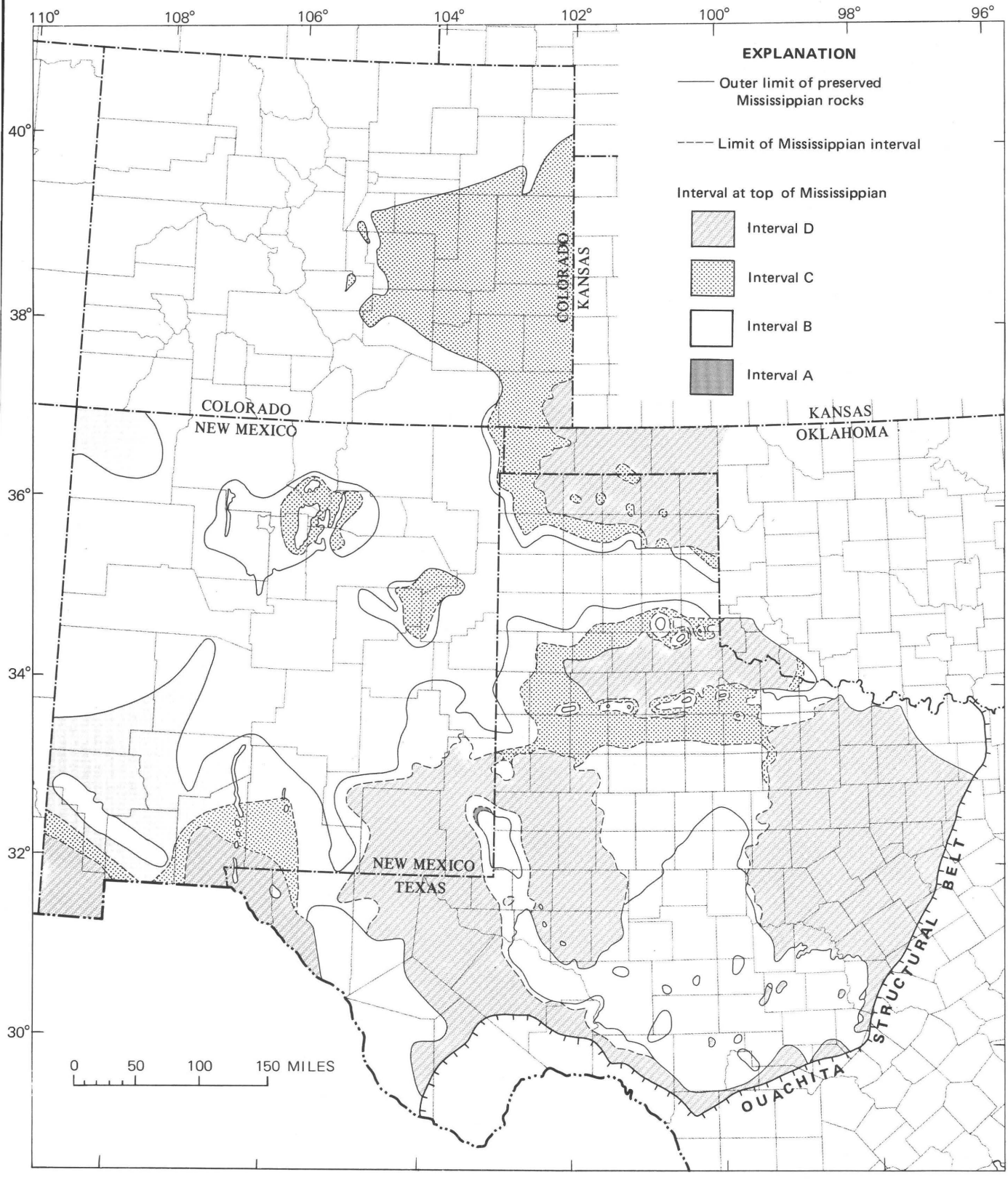

FIGURE 36. — Distribution of intervals at the top of the Mississippian System in the Southern Midcontinent and Southern Rocky Mountains region.

feet in the Hardeman basin and from 600–700 feet in parts of Cochran and Yoakum Counties in the Midland basin to less than 200 feet on the west side of the Palo Duro basin. Interval B is more than 800 feet thick in northern Lea County, N. Mex., and slightly less than 900 feet thick in a well in central Pecos County, Tex., but at these places no division was made between intervals B and C as mapped elsewhere in Texas.

More than 900 feet of strata are assigned to interval B in the Klondike Hills in southwestern New Mexico. Rocks of the interval thin depositionally from this area generally northeastward to about 200 feet in Sierra County. Farther north the interval is truncated beneath Pennsylvanian rocks.

Thickness trends in interval B in southeastern New Mexico and adjacent parts of Texas illustrate the effects of three periods of erosion and truncation. Originally deposited thicknesses in south-central and north-central New Mexico are at least slightly modified by erosion preceding deposition of interval C. In north-central New Mexico areas where interval C still overlies interval B are small, and no well-defined thickness trends resulting from this erosional episode are evident. In the Delaware and Midland basins interval B thins southward or southwestward to zero edges in northern trans-Pecos Texas and at the western end of the Pecos arch (fig. 33). Thickness changes in these areas resulted from truncation preceding deposition of Upper Mississippian rocks of interval D. Limestone in interval B thins southeastward across the Fort Worth basin to an extremely irregular digitated zero edge complicated by several disconnected masses of limestone well east of the main body. As in the Delaware and Midland basins, thickness changes in the Fort Worth basin resulted, at least partly, from planation beneath younger Mississippian rocks of interval D.

Erosion at the end of Mississippian or beginning of Pennsylvanian time accounts for rapid northward thinning of interval B at the northern margin of its subcrop in southern New Mexico, and for a gradual southward thinning of the interval on the Concho shelf in central Texas. Likewise, rapid thinning along narrow beveled edges of the interval on the flanks of the Amarillo uplift, the Central Basin platform, and the Red River and Muenster arches, as well as isolated areas of thin or locally absent interval B rocks on the trend of the Matador arch resulted from post-Mississippian uplift and erosion.

In the mountains of north-central New Mexico, the rocks assigned here to interval B range in thickness from a few inches to about 125 feet. In San Juan County, in the northwest corner of New Mexico, interval B ranges in thickness from a featheredge in the southeast part of the county to a little more than 250 feet in the northwest part.

In outcrops in the Llano region, the Chappel Limestone, which composes interval B, ranges in thickness from less than 1 foot to 50 feet; it is thickest where preserved in sinkholes or other depressions on the surface of the underlying Ellenburger Group (Cloud and Barnes, 1946, p. 50–51; Plummer, 1950, p. 21, 27).

Isolated subsurface occurrences of interval B in the vicinity of the Pecos arch and on the southern part of the Concho shelf are erosional outliers preserved in grabens that formed during a period of normal faulting at about the beginning of Pennsylvanian time.

LITHOFACIES TRENDS

At most places in the subsurface in Texas and parts of western Oklahoma interval B consists of light-gray, brownish-gray, pinkish-gray, or slightly greenish-gray crinoidal limestone containing light-gray, blue-white, or light-brown chert (pl. 4-B; pl. 9-F, secs. *I–I'*, *J–J'*, *L–L'*). In eastern Colorado and locally in Oklahoma, dolomite rather than limestone commonly makes up much of the interval. At a few localities the interval contains thin beds of gray or greenish-gray shale or siltstone. In eastern Colorado and adjacent parts of Oklahoma, Maher and Collins (1949) described a western shoreward facies consisting mostly of finely granular glauconitic very cherty dolomite, and an eastern or southeastern basinward facies composed of cherty coarsely crystalline crinoidal limestone, interbedded with some dolomitic limestone, fine-grained sandstone, and gray, green, and maroon shale. No systematic changes in proportion of shale, sandstone, and limestone or dolomite were noted (pl. 4-B).

Along much of the east side of the Midland basin and on nearby parts of the Concho shelf where interval B directly underlies Pennsylvanian rocks, the uppermost few feet to as much as 80 feet of the interval is chert. However, beyond the limit of Mississippian rocks in Nolan, Taylor, and parts of Jones and Fisher Counties, the Ordovician Ellenburger Group lies just below Pennsylvanian rocks and its uppermost part also commonly is chert. This stratigraphic relation indicates that chert in both the Mississippian and Ordovician rocks was concentrated during Early Pennsylvanian time in the carbonate rocks that then were exposed to weathering and that the massive chert concentration in this area, although making up a locally thick and fairly extensive lithic unit, is largely a secondary deposit that has no significance in interpreting the depositional history of interval B.

At outcrops in southern New Mexico, interval B consists of crinoidal limestone and subordinate marl (pl. 9-F, secs. *I–I'*, *J–J'*). Beds of limestone commonly are thick, biohermal, and ledge and cliff forming. In the Sacramento Mountains, reefs or bioherms form locally

conspicuous knoblike masses (Laudon and Bowsher, 1949). The Nunn Member of the Lake Valley Limestone, the upper member in the southern part of the Black Range, has a prolific fauna of crinoids that has been described by Wachsmuth and Springer (1897). Chert is not a major constituent of the interval in New Mexico, but it forms fairly abundant nodules at some horizons.

The Manuelitas Member of the Tererro Formation in the Sangre de Cristo Mountains, north-central New Mexico, is characterized by calcarenite that occurs in cyclic repetitions of coarse- and fine-grained units. Silt- to granule-sized quartz grains are abundant, and gray or pink oolitic or pisolitic chert is a distinguishing minor constituent (Baltz and Read, 1960, p. 1762–1763).

The Arroyo Penasco Group of Armstrong and Mamet (1974) of north-central New Mexico crops out in the Sierra Nacimiento and in the San Pedro, Jemez, Sandia, Manzano, and Sangre de Cristo Mountains, and includes rocks of late Osage (Zone 9), Meramec (Zones 12–14), and Chester (Zone 16_i). The late Osage-age beds rest unconformably on Precambrian crystalline rocks, and disconformities exist between the Osage, Meramec, and Chester strata. For this chapter, we have followed the terminology and interpretation of Baltz and Read (1960) for the Sangre de Cristo Mountains by excluding the Arroyo Penasco of Armstrong (1955, 1967) from that area, although recognizing that formation in the other areas mentioned. (See pl. 15, cols. 87, 88.) The entire Arroyo Penasco Group has been placed arbitrarily in interval B for convenience in mapping, although it includes both upper Osage, Meramec, and lower Chester rocks (Armstrong and Mamet, 1974).

The Leadville Limestone of northwestern New Mexico is largely light-brown cherty limestone and is recognized only in the subsurface.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Most of the Southern Midcontinent and Southern Rocky Mountains region was flooded by the sea beginning in late Kinderhook time, and marine conditions prevailed thereafter throughout much of Osage time (pl. 11, fig. 2). Crinoidal debris and the cementing limestone matrix, which make up much of interval B, originated at about the place they now are found. Crinoidal bioherms are common in the interval in New Mexico. Local, greatly thickened masses of Chappel Limestone have been described in Eastland, Stephens, Young, Jack, and Archer Counties and are explained by some writers as crinoidal bioherms or reefs (Turner, 1957, p. 61; Harmon, 1957; Vickers, 1957; North Texas Geological Society, 1954, well 18).

Probably much of the crinoidal limestone of interval B was deposited in moderately shallow water that was agitated, aerated, and warm. The delivery of food to flourishing colonies of crinoids or other stemmed echinoderms and the scattering of the skeletal remains of the animals would seem to require the stirring action of currents or waves on the sea floor.

The extent of the sea beyond the present margins of the interval is unknown. The general lack of terrigenous sediment in interval B, except perhaps for the rocks exposed in the Sangre de Cristo Mountains, implies either that land was remote or that it was so low that erosion was slight and the volume of material shed into ths sea was meager. No increase in terrigenous materials can be seen at the present margins of the interval to suggest nearness to land (pl. 12, fig. 2).

PALEOTECTONIC IMPLICATIONS

Much of the region discussed here was affected by only slight epeirogenic movements during interval B (pl. 10, fig. 2). Downwarping of the amplitude of a few hundred feet in distances of several hundred miles would have sufficed to flood the region and to accommodate the blanket of limestone deposited during the interval. Thickness trends in areas where interval B is represented by its originally deposited thickness suggest that the greatest and most continuous downwarping occurred in a basin that extended from Pecos County, Tex., northward into southeastern New Mexico, in part coincident with the Delaware basin of post-Mississippian time. The northern part of a second, roughly parallel trough extended a short distance into extreme southwestern New Mexico. The west end of a shallow ancestral southeast-trending Anadarko basin is indicated by isopachs in Hemphill and Wheeler Counties in the Texas Panhandle (pl. 4-A). An ill-defined northeast-trending depression at about right angles to the pre-Mississippian Texas arch is suggested by slightly thickened rocks of interval B in Cottle, Dickens, and Crosby Counties, Tex. A synclinal connection between the Anadarko and ancestral Delaware basins, called here the Dickens trough, is inferred for this area.

A slightly positive area in the vicinity of eastern San Miguel County in northern New Mexico is suggested by westward thinning of interval B in north Texas toward this area and by possible overlap of interval B by interval C locally in east-central New Mexico. A large part of eastern Colorado probably was also a stable or slightly positive area during interval B as indicated by depositional thinning in the eastern part of Colorado and overlap westward of interval B by interval C.

At about the end of Osage time, the Sangre de Cristo Mountains area of north-central New Mexico was

briefly emergent as shown by the unconformity or disconformity at the top of interval B in that region (Baltz and Read, 1960). Rocks of interval B are not deeply dissected, and the amount of uplift, therefore, probably was slight. According to the Armstrong (1967) interpretation for this area, there was no uplift at this time.

The eastern part of the Fort Worth basin, in central Texas, may have stood somewhat higher than nearby areas to the west during most or all of interval B (pl. 10, fig. 2), inasmuch as rocks of the interval thin generally eastward in east-central Texas, but at least some of the thinning is due to erosion before deposition of overlying interval D.

Laudon and Bowsher (1949, p. 19) dated the basal Mississippian rocks in the Franklin and Hueco Mountains as Meramec in age. If this assignment is correct, that area stood above sea level for at least part of Osage time. A thin shale unit of Late Devonian age underlies the Mississippian with a fairly uniform thickness everywhere in the Franklin and Hueco Mountains (King and Knight, 1944); any post-Devonian pre-Mississippian erosion, therefore, must have been slight.

INTERVAL C

FORMATIONS INCLUDED

Interval C includes rocks thought to be all or mostly Meramec in age (pl. 15). It contains equivalents of the *Ste. Genevieve, St. Louis, Salem,* and *Warsaw Limestones* in the Anadarko and Dalhart basins in western Oklahoma, northern Texas, and eastern Colorado (fig. 33), as traced into those areas by several geologists (Maher and Collins, 1949; Goebel, 1966; Texas Panhandle Sample Log Service, unpub. sample logs). The interval comprises the upper unit of the Chappel Limestone in central-northern Texas and the Mississippian lime or Osage-Meramec limestone, both of which are local subsurface terminology farther south on the northern part of the Concho shelf and in the adjacent Hardeman, Palo Duro, and Midland basins. The upper unit of the Chappel Limestone of southwestern Oklahoma was correlated by Culp (1961, p. 11) with the Sycamore Limestone of central Oklahoma. Fossils useful for dating have not been reported from any part of the Chappel Limestone or its equivalents in the subsurface in Texas or western Oklahoma. Assignments of the upper part of the Chappel to the Meramec and to interval C are based on stratigraphic position and lithologic similarity with rocks classified as Meramec in central Oklahoma and Kansas.

For convenience in mapping, rocks of known or presumed Meramec age were locally included in interval B rather than interval C. This is true in southwestern Texas (pl. 15, col. 81), locally in trans-Pecos Texas (pl. 15, col. 82), and in a small area in southeastern New Mexico (pl. 15, col. 82). The Arroyo Penasco Group of north-central New Mexico (pl. 15, col. 87) is also entirely within interval B.

In northern New Mexico, interval C includes the Cowles Member of the Terrero Formation in the Sangre de Cristo Mountains and an unnamed limestone in the subsurface in parts of San Miguel, Guadalupe, Quay, and De Baca Counties to the southeast. The Cowles Member of the Tererro Formation was classified as Meramec by Baltz and Read (1960, p. 1767) mainly because of stratigraphic position. Endothyrids in the Cowles are common to those reported from the upper part of the Arroyo Penasco Formation in the Sierra Nacimiento, which was considered Meramec in age (Fitzsimmons and others, 1956, p. 1942; Armstrong, 1967, p. 4). More recent study (Armstrong and Mamet, 1974, p. 150) indicates that the Cowles Member in the Sangre de Cristo Mountains contains a microfauna of early Chester age (Zone 16_i).

In south-central New Mexico and extreme western Texas, the *Las Cruces* and *Rancheria Formations* contain fossils of Meramec age, according to Laudon and Bowsher (1949, p. 19), and for this reason they are included in interval C, although in part they resemble rocks of interval B in other areas of New Mexico. In southwestern New Mexico rocks of Meramec age are present in the upper part of the Hachita Formation of the Escabrosa Group (Armstrong, 1970, 1962; Zeller, 1965). Rocks of Meramec age may be present in southeastern New Mexico; but proof is lacking, and any such rocks have been included with underlying interval B (pls. 4-A, 15).

Rocks included in interval C in the Wet Mountains of east-central Colorado are the Williams Canyon, Hardscrabble, and Beulah Limestones, all of which are separated by unconformites at top and base (Maher, 1950).

No rocks of interval C are recognized in the Ouachita structural belt, the Marathon area, or the Llano region.

UPPER BOUNDARY OF INTERVAL C

The upper boundary of interval C is an unconformity at the base of younger Mississippian rocks, except perhaps in the Peloncillo Mountains and nearby parts of southwestern New Mexico and in the trough of the Anadarko basin in Lipscomb and Hemphill Counties, Tex. The contact is made conspicuous by an abrupt lithologic change, commonly from oolitic or sandy light-gray limestone at the top of interval C to dark-gray, greenish-gray, or red shale and shaly limestone at the base of interval D. Southward or southeastward beveling of interval C in the northern part of the Midland basin in Cochran, Yoakum, and Gaines Counties and in the northeastern part of the Fort Worth basin in

Throckmorton, Young, and Baylor Counties clearly demonstrates the unconformable relations in these parts of Texas.

The upper boundary of interval C is exposed in the Hueco and Franklin Mountains (pl. 15, col. 83), where the contact with overlying rocks of interval D is probably a disconformity. Thin-bedded silty limestone at the top of the *Rancheria Formation* in this area is abruptly overlain by interlaminated green shale and brown sandstone in the Helms Formation in interval D.

The upper part of the Hachita Formation of the Escabrosa Group, which makes up interval C in the Peloncillo Mountains, southwestern New Mexico, was described as conformable with younger Mississippian rocks of the Paradise Formation (Gillerman, 1958, p. 30). The contact there is marked by a change from massive ledge-forming beds of cherty calcarenite at the top of the Hachita to mudstone or argillaceous limestone at the base of the overlying Paradise.

Pennsylvanian rocks unconformably overlie interval C where the interval extends beyond the limits of overlying interval D as shown on plate 9-F (secs. *I–I'*, *J–J'*, *L–L'*). Such areas include a wide band nearly surrounding the Palo Duro basin in northern Texas; a similar band on the south and west sides of the Dalhart and Anadarko basins farther north in Texas, Oklahoma, and eastern Colorado; and a generally east trending area in the southernmost row of counties in New Mexico. Pennsylvanian rocks directly overlie interval C, also, in northern New Mexico.

THICKNESS TRENDS

Interval C, because of unconformities at its top, probably nowhere is present in its originally deposited thickness except perhaps in the vicinity of the Peloncillo Mountains in extreme southwestern New Mexico and along the axis of the Anadarko basin in Texas. The depth of erosion preceding deposition of interval D probably was slight at most places in the Panhandles of Texas and Oklahoma. The top of the interval in this area is finely sandy limestone correlated with the Ste. Genevieve Limestone of Kansas (Maher and Collins, 1949) and the Sycamore Limestone of Oklahoma (Culp, 1961). This unit is widespread and fairly uniform in thickness and displays little evidence of deep dissection.

Interval C attains a thickness of 1,340 feet in central Lipscomb County, Tex., in the trough of the east-trending Anadarko basin (pl. 5-A). It thins rapidly from this area to the north and south, and it thins much more gradually to the west.

Interval C is as much as 400 feet thick in the Hardeman basin in southwestern Oklahoma, and its thickness ranges from 200 to 300 feet farther west in much of the Palo Duro basin. Post-Mississippian faults abruptly cut off the interval on the northeast side of the Hardeman basin, and faults terminate the interval on the north flank of the Amarillo arch in Carson and Gray Counties, Tex. Areas in which the interval is absent on the Matador arch and at the north edge of the Palo Duro basin resulted from post-Mississippian deformation and erosion.

Interval C attains a thickness of 210 feet in the subsurface in San Miguel County and parts of nearby counties, northeastern New Mexico. In the Sangre de Cristo Mountains the thickness of the Cowles Member of the Tererro Formation is irregular because of erosion at the top; its maximum is 50 feet.

In the Wet Mountains, south-central Colorado, a thickness of slightly over 200 feet is mapped for the combined thicknesses of the three formations assigned to interval C. The interval thickens to more than 600 feet near the southeastern corner of the State.

In the Franklin Mountains, westernmost Texas, interval C is as much as 450 feet thick; it thins from this area to zero within a few miles to the east, north, and west. The thickness changes are almost balanced by inverse changes in thickness in underlying interval B in the same area. This inverse relation between intervals B and C suggests that the *Las Cruces Formation* and at least the lower cherty part of the *Rancheria Formation* (lower member of the Helms Formation as mapped by King and Knight, 1944), which make up interval C in the Franklin and Hueco Mountains, may be equivalent to beds mapped in interval B and considered Osage and latest Kinderhook in age in nearby areas to the north and east. Moreover, an abundance of chert in the *Rancheria Formation* suggests correlation on a lithologic basis with rocks included in interval B in other parts of Texas and New Mexico.

The upper part of the Hachita Formation of the Escabrosa Group mapped in interval C in southwestern New Mexico has a maximum thickness of about 200 feet, which occurs in southwest Hildalgo County, and it thins northeastward.

LITHOFACIES TRENDS

Interval C is mostly light-gray and light-brownish-gray limestone (pl. 5-B). In only two areas are other constituents of sufficient quantity to affect the lithofacies patterns: (1) in the Hardeman basin of southwestern Oklahoma, where interbeds of mudstone are abundant enough to show with a muddy limestone pattern on the lithofacies map, and (2) in the Hueco and Franklin Mountains area of western Texas, where the lower part of the *Rancheria Formation* and the *Las Cruces Formation* contain more than 10 percent chert and require a chert overprint.

In many parts of the region constituents other than limestone are present but in amounts too small to affect the limestone classification (pl. 5-B; pl. 9-F, secs. *I–I'*,

J–J', *L–L'*). The limestone is commonly cherty and oolitic. The upper part of the interval characteristically is silty or sandy in the Panhandles of Texas and Oklahoma and in eastern Colorado, and oolites are especially common in the middle part of the sequence in beds equivalent to the St. Louis Limestone. Some dolomite is present locally. In northern New Mexico the entire interval is limestone that contains disseminated fine quartz sand.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Oolitic limestone that characterizes interval C was deposited in clear agitated water in an environment (pl. 11, fig. 3) that probably was much like that which prevailed during the preceding interval B. An eastward increase in the amount of mudstone in north-central Texas indicates a source of terrigenous sediment to the east. Land inferred in northern New Mexico during interval B may have persisted and furnished some quartz sand to interval C, although the direction to any such land is not shown in the facies patterns. It seems certain that the extent of the sea was much greater than the present limit of interval C rocks, and the sea may have covered the entire region.

PALEOTECTONIC IMPLICATIONS

In most of the region no notable tectonic event interrupted sedimentation from interval B to interval C. The region was a stable shelf except for the eastern Panhandle areas of Texas and Oklahoma where an east-trending trough continued to sink during interval C time (pl. 10, fig. 3). During the last part of interval C, perhaps in late Meramec time, the region was uplifted sufficiently for the sea to withdraw. Erosion was deepest, and hence the amplitude of uplift probably was greatest, in the southeastern part of the Fort Worth basin and perhaps in adjacent parts of the Ouachita structural belt, in the southern part of the Midland basin, and in parts of trans-Pecos Texas, including the Marathon area. In these areas any previously deposited rocks of intervals C and B were completely stripped away. Confirmation of uplift in central Texas is provided by thickened masses of Chappel Limestone in sinkholes in the Llano region. These masses are overlapped by the Barnett Formation of interval D, which is itself slightly thickened over the central parts of the sinkholes. Cloud and Barnes (1946, p. 110–111), who described these relations, pointed out that formation of the sinkholes, collapse of the Chappel Limestone into them, and subsequent virtual removal of the Chappel from areas between sinkholes must have been a post-Chappel and pre-Barnett sequence of events brought on by subaerial exposure.

INTERVAL D

FORMATIONS INCLUDED

Interval D consists of rocks assigned an uppermost Meramec and Chester age (pls. 6-A, 15). The interval characteristically contains a basal mudstone unit and overlying limestone unit in most of the region including western Oklahoma, north-central and western Texas and southeastern New Mexico. In the Fort Worth basin (fig. 33) these units are the Barnett Formation and overlying *Comyn Limestone* or their equivalents as correlated in cross sections of the Fort Worth Geological Society (1954) (figs. 37*A*, *C*). Unnamed lithologic counterparts of these two units comprise interval D in the Palo Duro and Hardeman basins and in the Anadarko and Dalhart basins in the Oklahoma and northern Texas Panhandles. The Barnett and an unnamed overlying limestone are the units assigned to interval D in the Midland and Delaware basins and in eastern trans-Pecos Texas farther south. A shale, generally called the Barnett Formation, makes up the interval where it is present in the Val Verde and Kerr basins. The Helms Formation (as restricted by Laudon and Bowsher, 1949) makes up interval D in western trans-Pecos Texas and south-central New Mexico, and the Paradise Formation is interval D in southwestern New Mexico.

The *Stanley Shale* makes up interval D in the northern part of the Ouachita structural belt, and the lower part of the Tesnus Formation composes the interval in the western part of the structural belt.

The age and correlation of some of the units here assigned to interval D have been debated for many years. A brief mention of paleontologic evidence supporting the present assignment, therefore, seems appropriate.

The Barnett Formation crops out in the Llano region central Texas, where it has been assigned ages ranging from late Osage to early Pennsylvanian. Conflicting opinions of the earlier workers were summarized by Cloud and Barnes (1946, p. 53–59) and by Hass (1953, p. 69–72).

Hass (1953) found in the Barnett Formation everywhere in the Llano region an upper conodont zone he called Meramec in age, and he found in the southwestern part of the region in beds assigned to the Barnett (Cloud and Barnes, 1946, p. 56) a lower conodont zone he called Osage in age. Beds containing the lower zone are reported only in the southwest part of the Llano region (fig. 32) and are in a limestone sequence atypical of the Barnett in surrounding areas. For the purposes of the present work, the limestone sequence containing the lower zone is separated from the Barnett and is assigned to interval B. Of 18 species of conodonts listed

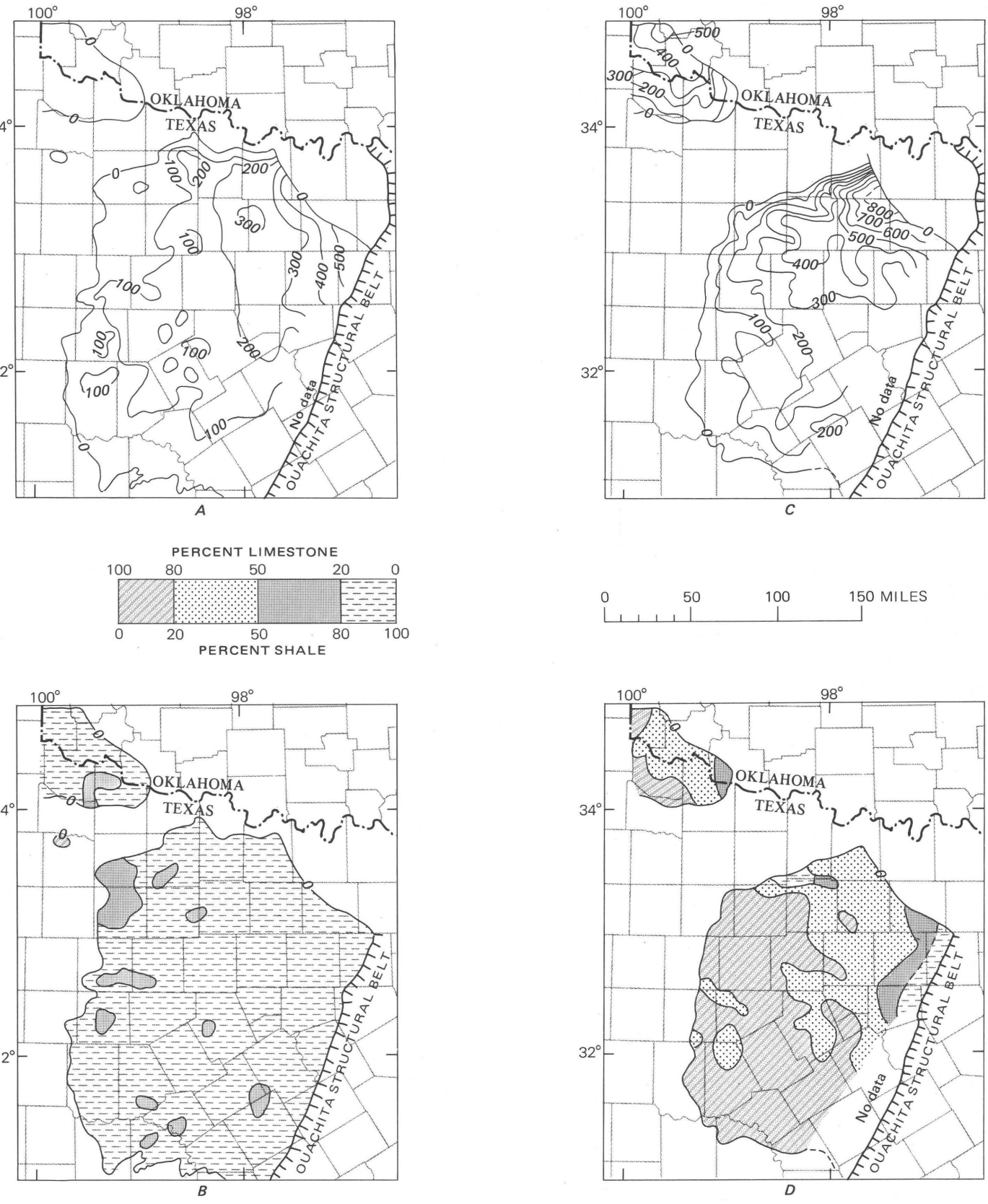

FIGURE 37. — Thickness and lithofacies of Barnett Formation and *Comyn Limestone* and equivalent rocks in north-central Texas and southwestern Oklahoma. *A*, Thickness, in feet, of Barnett Formation and equivalent rocks. *B*, Lithofacies of Barnett Formation and equivalent rocks. *C*, Thickness, in feet, of *Comyn Limestone* and equivalent rocks. *D*, Lithofacies of *Comyn Limestone* and equivalent rocks.

by Hass (1953, table 1) in his upper faunal zone of the Barnett Formation, 7 were found by Rexroad and Collinson (1961, pl. 1) and by Collinson, Scott, and Rexroad (1962, chart 4) to be confined to rocks of Chester age in the upper Mississippi Valley and Illinois basin; only one of the species listed by Hass (1953), namely *Gnathodus texanus* Roundy, occurs in Illinois in rocks older than the Ste. Genevieve Limestone of latest Meramec age. Rexroad and Clarke (1960, p. 1204–1206) noted a remarkably close correspondence between the Barnett conodont fauna and conodonts from the Glen Dean Formation and its equivalents of mid-Chester age in Illinois and adjacent areas. They concluded that the part of the Barnett containing the upper faunal zone of Hass (1953, pl. 1) is Chester in age, with the possibility that it could represent deposition into the Pennsylvanian.

Most paleontologists concerned with the Barnett point out a close correspondence of conodont and (or) shelly faunas among the Barnett Formation of Texas, *Delaware Creek Member* of the Caney Shale of Oklahoma, and the Moorefield, Ruddell, Batesville, and Fayetteville Formations of Arkansas (Giry, in Roundy and others, 1926, p. 3, 4; Miller and Youngquist, 1948, p. 651; Plummer, 1950; Hass, 1953; p. 72). Correlation of the Barnett with these formations accords fairly well with assignment of the Barnett to latest Meramec and Chester.

Hass (1950, p. 1580; 1956, p. 31–32) stated that conodonts from the lower part of the Stanley Shale in Arkansas and Oklahoma are conspecific with those in the upper faunal zone of the Barnett Formation. Ammonoids reported near Little Rock, Ark., indicate that the upper part of the Stanley Shale and the overlying Jackfork Sandstone in that region are Early Pennsylvanian (Morrow) in age (Gordon and Stone, in U.S. Geological Survey, 1967, p. A77). Because of thrust faults at some places and erosion beneath an unconformity at the base of Cretaceous rocks at other places, no complete sections of the *Stanley Shale* are known in Texas. For the purposes of the present work, all of the *Stanley* in Texas, including any part that may be earliest Pennsylvanian, is assigned to interval D.

The subsurface *Comyn Limestone* was originally considered laterally continuous with the Marble Falls Limestone that crops out in the Llano region and hence Pennsylvanian in age (Cheney, 1940, 1947; Cheney and Goss, 1952). As Turner (1957, p. 63–64) pointed out, however, the *Comyn* appears to be conformable with the underlying Barnett Shale of Mississippian age but is unconformably overlain and truncated by Pennsylvanian rocks in the northern and western parts of the Fort Worth basin. It is assigned to the Mississippian System on this basis.

Limestone overlying the Barnett Formation in the Midland basin yields fossils of Chester age (Paleontological Laboratory, Midland, Tex., unpub. reports fo Martin, Dawson, Andrews, and Gaines Counties, Tex.)

A Chester age for the lower part of the Tesnus For mation in the Marathon region of west Texas i established by conodonts from less than 100 feet abov the base (S. P. Ellison, Jr., written commun., 1967) Pennsylvanian fossils are found near the top (Davi White, in King, 1937, p. 61; Bruce Harlton, in Powers 1928, p. 1066).

Fossils of Chester age collected from the Helms For mation have been listed by King and Knight (1944) Laudon and Bowsher (1949), and Plummer (1950).

The Paradise Formation in southwestern New Mex ico ranges in age from late Meramec to middle Cheste (Hernon, 1935; Gillerman, 1958, p. 30; Zeller, 1965, p 29–30).

UPPER BOUNDARY

Interval D is overlain unconformably by Pennsylva nian rocks of Morrow or Atoka age in most parts of th region. Pennsylvanian rocks as young as Des Moine overlie interval D along the south flank of the Re River arch. Permian rocks of the Wolfcamp Series un conformably overlie interval D locally in the vicinity o the Central Basin platform and in trans-Pecos Texas at one locality in southern Reeves County, Tex., Per mian rocks of the Leonard Series rest on interval D.

In much of Texas and western Oklahoma, the base o the Pennsylvanian System is marked by sandstone sandy mudstone, or sandy limestone, and, generally the contact of these rocks with mudstone or limeston containing little or no sandstone in interval D is readil apparent. In the southeastern part of the Fort Wort basin, however, limestone of the *Comyn Limestone* i overlain directly by limestone of the Pennsylvania Marble Falls Limestone without any lithologic brea that could be detected in the sample logs. Where this i thought to be the relation, the contact is drawn ar bitrarily within the limestone sequence.

Similarly, in parts of the Midland basin, west Texas limestone of Atoka age rests on limestone of Lat Mississippian age without any sandy beds to mark th contact. Fusulinids of Atoka age near the base of th Pennsylvanian rocks (Paleontological Laboratory, Mid land, Tex., unpub. reports) give control for locating th boundary.

In the eastern part of the Fort Worth basin, near th front of the Ouachita structural belt in Coryell, Bosqu Hill, and Johnson Counties, the contact betwee Mississippian and Pennsylvanian rocks projects int the lower part of a sequence several thousand feet thic of dark-gray mudstone, fine-grained sandstone, an siltstone containing near the base a few discontinuou beds of limestone. No clear lithologic distinction coul be made within the lower part of this sequence betwee

rocks of Mississippian and Pennsylvanian age, and if an unconformity between the systems is present, it could not be recognized. The thickness and lithofacies of Mississippian rocks in this area are too uncertain to be shown on isopach and facies maps.

The upper boundary of the Helms Formation with overlying Pennsylvanian rocks in west Texas and southern New Mexico is an unconformable contact of mudstone and sandy mudstone containing a few limestone lenses in the Helms that underlies shale and thick-bedded corraline limestone in the Pennsylvanian.

Sandstone and conglomerate of Cretaceous age overlie the *Stanley Shale* in most of the Ouachita structural belt. The *Jackfork Sandstone*, considered here as Pennsylvanian in age, conformably overlies the *Stanley* in the Ouachita Mountains in southern Oklahoma, but it is definitely reported in Texas only in Fannin County (Flawn and others, 1961, p. 18). Mississippian rocks grade upward into Pennsylvanian rocks within the Tesnus Formation in the Marathon area.

THICKNESS TRENDS

The originally deposited thickness of interval D is nearly everywhere modified by post-Missippian erosion (pl. 9-F, secs. *I–I'*, *J–J'*, *L–L'*). The present zero edges of the interval (pl. 6-A) outline large basinal areas where Mississippian rocks were protected, by post-Mississippian downwarping, from severe erosion in Pennsylvanian and Early Permian time.

Within the Ouachita structural belt, the *Stanley Shale* has a partial thickness of 6,700+ feet in western Williamson County, Tex. (Flawn and others, 1961, p. 333). Unknown amounts have been eroded away beneath Cretaceous rocks here and at other places in the structural belt.

In areas north and west of the Ouachita structural belt, the maximum thickness of rocks in interval D is about 2,000 feet in the Delaware basin in Reeves and Ward Counties, Tex. The interval is no more than 750 feet thick in the Midland basin in Midland County, Tex. Maximum thicknesses for interval D elsewhere in Texas and western Oklahoma are 1,280 feet at the eastern end of the Anadarko basin in northeastern Hemphill County, Tex., and about 1,200 feet in the northeastern part of the Fort Worth basin adjacent to the Muenster arch in southwesternmost Cooke County, Tex. A post-Mississippian fault along the south flank of the Muenster arch abruptly terminates Mississippian rocks northeast of the last named locality.

In south-central New Mexico interval D is no more than about 200 feet thick. In southwestern New Mexico, subsurface data give a thickness of 430 feet for interval D. It is absent in the central and northern parts of the State.

Although interval D rocks do not extend into northeastern New Mexico from the Anadarko basin, they do extend north from the Panhandles of Texas and Oklahoma into the southeastern corner of Colorado as a wedge less than 100 feet thick.

LITHOFACIES TRENDS

In the area of the Texas and Oklahoma Panhandles, the lower 50–100 feet of interval D characteristically is gray, green, brown, or reddish-brown mudstone interbedded with yellowish-gray oolitic limestone. Some calcareous, locally glauconitic, sandstone commonly is also present. The overlying part of the interval consists of light-gray to brownish-gray, locally oolitic, limestone interbedded with some gray or greenish-gray mudstone. Maher and Collins (1949) noted a close lithologic resemblance to the approximately equivalent Fayetteville Shale and overlying Pitkin Limestone of Arkansas. Patterns shown for interval D on the lithofacies map (pl. 6-B) and the cross sections (pl. 9-F) reflect the depth of erosion of the limestone sequence at the top of the interval beneath Pennsylvanian rocks. No patterns that clearly are regional depositional facies patterns are evident.

In the Fort Worth basin, interval D is dark-brownish-gray to dark-gray petroliferous mudstone at the base, grading upward in the upper part to dark-gray to dark-brownish-gray, locally oolitic, limestone interbedded with mudstone. Glauconite and grains of phosphatic material are locally present in the mudstone at outcrops in the Llano region (Defandorf, 1960, p. 25). The interval is more than 50 percent limestone in an irregular area in the central part of the Fort Worth basin (pl. 6-B). Interbeds of mudstone increase eastward in both the Hardeman and Fort Worth basins; in the northeastern part of the Fort Worth basin, the interval is locally more than 80 percent mudstone. The change in facies occurs wholly within the *Comyn Limestone*, as shown in figures 37*B* and 37*D*. A predominance of mudstone over limestone on the north, west, and south sides of the Fort Worth basin is due largely to removal by erosion of the *Comyn Limestone* along these margins of the basin rather than to a facies change.

Mudstone in interval D thickens southward in the southern part of the Midland basin in Reagan and Upton Counties, Tex. In the Kerr and Val Verde basins, farther south, dark-gray mudstone makes up most of interval D. Information from drilling is inconclusive in determining if the southward increase in mudstone is due to depositional thickening within the Barnett Formation or is a facies change from limestone to shale within the overlying part of interval D similar to the facies change eastward in the *Comyn Limestone* in the Fort Worth basin. The latter possibility seems more likely.

In southeastern New Mexico and western trans-Pecos Texas, interval D consists of greenish-gray and gray mudstone, some brown muddy sandstone, and a few lenticular beds of argillaceous limestone. This facies appears to grade northward into argillaceous limestone near the northern limit of the interval in the south-central and southeastern parts of the State. Limestone is the prevalent rock type in interval D in southwestern New Mexico.

The *Stanley Shale* and the laterally equivalent lower part of the Tesnus Formation, which compose interval D in the Ouachita structural belt, are dark-gray or greenish-gray micaceous mudstone thinly interbedded with lesser amounts of light-gray or pale-greenish-gray argillaceous feldspathic sandstone and siltstone (King, 1937, p. 55–61; Flawn, 1961, p. 73; Goldstein and Hendricks, 1962, p. 390–393). These formations are shown generally just to the west and north of the Ouachita structural belt, from the Llano region to the Marathon area (pl. 6-B). Within the belt, and generally to the south and east of the belt, no regional facies patterns are shown because of the extremely limited data.

SOURCES AND ENVIRONMENTS OF DEPOSITION

The *Stanley Shale* and Tesnus Formation are widely regarded as deepwater deposits carried into place mainly by turbidity currents (Cline and Shelburne, 1959, p. 205–206; Goldstein, 1961, p. 43; King, 1961, p. 184; Goldstein and Hendricks, 1962, p. 394). Much of the sediment in the *Stanley Shale* traveled into the region from the northeast along the axis of the geosyncline as shown by the orientation of flute casts and other directional sedimentary features in outcrops in the Ouachita Mountains in Oklahoma and Arkansas (Shelburne, 1960; Briggs and Cline, 1963). Land areas south or southeast of the Ouachita Mountains in east Texas or Louisiana have also been suggested as possible contributors of sediment (Miser, 1921; Harlton, 1934, fig. 3, p. 1027; Sellards, 1932, p. 21; King, 1961, p. 184; Goldstein, 1961, p. 43; Klein, 1966, p. 316). Land areas in northern Mexico may have provided the sediment for the Tesnus Formation.

The Barnett Formation and its lithogenetic equivalents in Texas, Oklahoma, and New Mexico are unlike the *Stanley Shale* and Tesnus Formation in that they lack convolute bedding, graded bedding, loadcasts, and other sedimentary features thought to typify turbidites. Apparently, the Barnett, Helms, and other mudstone units at the base of interval D were deposited on a shelf or in a basin area marginal to the geosyncline in shallower water than were the *Stanley* and Tesnus Formations. The eastward increase in mudstone in the *Comyn Limestone* and in thickness of the Barnett Formation indicate that the source of the mudstone in these formations was, like the *Stanley*, to the east or northeast. Black muds possibly stirred up by sediment flows in deeper water to the east apparently drifted slowly and intermittently westward or southwestward and were broadly deposited in shallower flanking parts of the sea. Black petroliferous mudstone and the presence locally of glauconite and phosphatic material in the Barnett suggest rather slow deposition under reducing conditions early in the interval. The supply of mud lessened and bottom waters evidently became more agitated and probably more aerated later in the interval, as shown by the decrease in mudstone and increase of locally oolitic limestone in the *Comyn*.

A similar history accounts for interval D in the southern parts of the region, except there the clay and mud that make up the Helms and laterally continuous units at the base of the interval spread northward from deeper water areas in which the Tesnus was deposited.

There is little evidence in the facies or thickness relations to indicate land in Texas during deposition of interval D. Deposits that now are isolated from one another by post-Mississippian erosion probably were once parts of a continuous blanket of sediment.

Depositional thinning of uneroded parts of the interval northward from the southern border of New Mexico and westward in northwestern Texas and western Oklahoma can be interpreted as evidence for land in New Mexico; however, the size and outline of any such land is highly conjectural. The contribution of terrigenous sediment from any such land to interval D apparently was slight.

The paleogeographic map of the area (pl. 12, fig. 4*A*) shows Texas covered by the sea, with the possible exception of some low land in the area adjoining central eastern New Mexico. New Mexico, conversely, is shown as possibly 70 percent land interspersed with shallow seas in the southern, southeastern, and extreme northeastern parts. Southeastern Colorado and the Panhandle of Oklahoma were also covered by shallow marine waters. The sea was probably deep along the axis of the Ouachita geosyncline (pl. 10, fig. 4). In areas of pronounced subsidence in the ancestral Delaware and Anadarko basins, sedimentation probably kept pace with subsidence and shallow-water conditions prevailed.

PALEOTECTONIC IMPLICATIONS

Interval D was marked in Texas by a rapid sinking of the Ouachita geosyncline (Llanoria geosyncline of Sellards, 1934, p. 37–38), a structural and topographic trough (pl. 10, fig. 4) whose deposits now follow a sinuous course westward across southern Arkansas and southeastern Oklahoma into northeast Texas, and from there south and southeastward in an arc through Texas into Mexico (Miser, 1929, p. 7; Sellards, 1934, p. 37–38; King, 1961, p. 182; Flawn and others, 1961, p.

13). Within the geosyncline were deposits several thousand feet thick of the *Stanley Shale* and the lower part of the Tesnus Formation. The original position of the axis of the geosyncline is now in doubt because later compressive forces have displaced the geosynclinal deposits northwestward or westward perhaps several tens of miles (Hendricks, 1959; King, 1961, p. 187). A rather abrupt increase in the rate of thickening of interval D (pl. 6-A) in the northeastern part of the Fort Worth basin, however, suggests that in Dallas and Collin Counties the border zone between the west side of the Ouachita geosyncline and the flanking shelf may have about coincided with the present front of the Ouachita structural belt.

Sedimentation lagged behind downwarping in the geosyncline, and, during the later part of the interval, deepening water in the geosyncline trapped increasing amounts of southeastward-moving sediment and prevented all but minor amounts from reaching the craton. Oolites in much of the *Comyn Limestone* and in laterally equivalent unnamed limestone units in Texas and Oklahoma indicate a shallow-water environment for the craton, which requires a marked difference in rates of subsidence between the geosyncline and craton during the later part of interval D.

In early interval D time, as the axial part of the geosyncline sank, the flanking cratonic margin, which included the areas of the present Fort Worth, Kerr, Val Verde, and Marfa basins, was also carried down below sea level, and interval D sediments began to accumulate. A basinal area of somewhat indefinite shape was reactivated at the site of the post-Mississippian Delaware basin in Reeves and Ward Counties and parts of adjacent counties, trans-Pecos, Texas. The east-trending Anadarko basin continued to sink in the northeastern Texas Panhandle. A southeast-trending area of retarded subsidence, which separated the two basins, roughly followed the axis of the old Texas arch from east-central New Mexico to the Llano region in central Texas (fig. 34). A saddle crossed the central part of this arch and connected the Anadarko and ancestral Delaware basins along the trend of the Dickens trough.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

The total thickness of Mississippian rocks is shown on plate 7. Disconnected areas in which Mississippian rocks are now found are remnants of once more extensive sheets of rocks that were partly dissected and beveled by erosion at the ends of intervals C and D during Mississippian time and by erosion on structurally active features during Pennsylvanian and Permian time. The most marked effects were produced by events that occurred after the Mississippian Period.

The most complete record of Mississippian deposition is found in the Anadarko basin where about 3,100 feet of Mississippian rocks, representing all the intervals, is preserved. In the Delaware basin, as much as 2,500 feet of Mississippian rocks is preserved. In contrast, a thickness of 6,700+ feet of rocks occurs in an incomplete section of the *Stanley Shale* of predominantly Late Mississippian age within the Ouachita structural belt.

PALEOTECTONIC IMPLICATIONS

The tectonic pattern early in the period was one of very mild orogenic activity. There is little evidence of structural movements of any great amplitude anywhere in the region. Epeirogenic uplifts sufficient to raise all of central Texas, much of New Mexico, and parts of Oklahoma and Colorado above sea level are recorded by regional unconformities at the base of intervals B and D — that is, at the base of the Chappel and Barnett Formations and their equivalents. Remnants of pre-Chester Mississippian rocks are too limited to more than very broadly identify and outline areas of maximum uplift during the first of these epeirogenies. Thickness trends in the Chappel suggest that erosion during this time was greatest in the southeastern part of the Fort Worth basin, east or southeast of the Llano region, and in the vicinity of the Pecos arch; all these areas lie at the edge of the present-day Ouachita structural belt.

Rocks of intervals B, C, and D all thicken into the west end of the Anadarko basin in Lipscomb and Wheeler Counties, Tex.; this basin, therefore, was subsiding moderately during most of Mississippian time. A basinal area roughly coincident with the post-Mississippian Delaware basin in southeastern New Mexico and adjacent Texas also was moderately active during interval B early in the period, and again during interval D in Late Mississippian time. The area of downwarping in this ancestral Delaware basin probably encompassed the area of the present-day Central Basin platform.

Greatly contrasting thicknesses of Mississippian rocks across the west front of the Ouachita structural belt reflect the most important tectonic event of the period — namely, the beginning, probably in latest Meramec time, and the subsequent rapid deepening to several thousand feet, in Chester time, of the Ouachita geosyncline. Concurrently, the bordering part of the craton west and north of the geosyncline sank slowly and unevenly. Because of post-Mississippian structural deformation and erosion, the border zone between the craton and the geosyncline cannot now be located, except that it was within the present Ouachita structural belt. The outer or southeastern margin of the geo-

syncline, if it was at all well defined in Texas, likewise cannot now be located.

Except in the Ouachita geosyncline, deposition of Mississippian rocks was concluded near the end of Mississippian time by regional uplift and gentle folding. Some of the structural elements that govern the present distribution of Mississippian rocks began to form at this time, and their development continued in later Paleozoic time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Pennsylvanian rocks overlie the Mississippian at most places west and north of the Ouachita structural belt (pl. 8). The oldest Pennsylvanian rocks, in general, occur along the periphery of the region in Texas, Colorado, Oklahoma, and southern New Mexico; the youngest Pennsylvanian rocks are found in contact with the Mississippian on the Matador and Red River arches, in the northern part of the Concho shelf, and in an isolated area north of the Central Basin platform. Permian rocks lie on the Mississippian at a few places in trans-Pecos Texas and in the vicinity of the Central Basin platform.

Unnamed rocks of Morrow age overlie unnamed Mississippian rocks at most places in eastern Colorado and in the Panhandles of Oklahoma and Texas north of the Matador arch. The Marble Falls Limestone of Morrow age overlies the *Comyn Limestone* or Barnett Formation of Chester age in the eastern part of the Fort Worth basin and in the Kerr basin. Pennsylvanian rocks of Atoka age, assigned to the *Big Saline Formation* in the western Fort Worth basin, to the unnamed limestone in the Midland basin, or to the *Atoka Formation* in a narrow band adjacent to the Ouachita structural belt in southwestern Texas, rest on the Mississippian. On the northern part of the Concho shelf, at the south side of the Red River arch, and at a few other smaller areas, limestone or shale assigned to the Strawn Group of Des Moines age overlies the Mississippian System.

In the Delaware basin and in areas along the New Mexico–Mexico boundary, rocks of Morrow age are in unconformable contact with rocks of Chester, Meramec, or Osage age. The Pennsylvanian rocks have been assigned to the La Tuna Member of the Magdalena Formation in the Franklin Mountains, the Magdalena Formation in the San Andres Mountains, and the Horquilla Limestone in the Big Hatchet Mountains and nearby areas.

In the mountains of north-central New Mexico, Mississippian rocks are overlain by the Sandia Formation of Morrow and Atoka age. The Lower Pennsylvanian Molas Formation overlies the wedge of Leadville Limestone in the subsurface of San Juan County, N. Mex. The Pennsylvanian part of the Fountain Formation overlies the rocks of interval C farther north in the mountains of south-central Colorado.

In the eastern part of the Ouachita structural belt, the Mississippian *Stanley Shale* grades upward into Pennsylvanian rocks in the Jackfork Sandstone in Fannin County, Tex. Lower Cretaceous rocks of the Trinity Group truncate the *Stanley Shale* in an east-trending band that extends along the Texas-Oklahoma border from eastern Grayson County to western Bowie County, and in a southward-widening band that extends from central Dallas County southward across Bell County. South and east of these areas, Mississippian rocks, where present, are overlain by rocks of Middle and Late Jurassic age (M. E. MacLachlan, in McKee and others, 1956). The lower, Mississippian part of the Tesnus Formation grades upward into the upper part of the formation of Pennsylvanian age in the Marathon area.

PALEOTECTONIC IMPLICATIONS

The area west and north of the Ouachita structural belt was broadly uplifted at about the end of Mississippian time with the result that the sea withdrew from much of the region into the bordering Ouachita geosyncline. Generally, uplift was greatest and land areas persisted longest in the southern part of the Concho shelf and along the trends of the Muenster and Red River arches and probably parts of central and northeastern New Mexico. Erosion in these areas during Early Pennsylvanian time eventually stripped off all previously deposited Mississippian rocks. The Muenster and Red River arches were bordered by faults that were active intermittently beginning early in Pennsylvanian time, and along which the cumulative displacements ultimately were several hundred to several thousand feet.

Downwarping along the inner margin of the Ouachita structural belt was renewed early in the Pennsylvanian Period, probably in response to compressive forces acting westward or northward within the structural belt (Flawn, 1959, p. 24–25; King, 1961, p. 187; Adams, 1962, p. 377). The sea began a slow advance across north-central and west Texas from the northeast, east, and south. The Concho shelf and adjacent regions to the west were flooded by Middle Pennsylvanian (early Des Moines) time, and the Muenster and Red River arches were flooded and buried somewhat later in the period. These features ceased to exert any influence on sedimentation thereafter.

Compression, uplift, and faulting in the frontal zone of the Ouachita structural belt continued into Middle Pennsylvanian time in the eastern part of the belt

(Flawn, 1959, p. 26; Hendricks, 1959, p. 54; King, 1961, p. 187). A similar diastrophism probably began somewhat later in the Marathon area and was concluded later (King, 1961, p. 187). The compressive forces obliterated the Ouachita geosyncline, and the region of the structural belt became a rising landmass that supplied detritus to areas in the interior. In the broad central regions of Texas and in New Mexico and Colorado, local basins and intervening uplifts began to form generally in Atoka or Des Moines time. These structural movements continued sporadically, generally with increasing intensity, during Late Pennsylvanian and Permian time.

Erosion in the Ouachita structural belt during Pennsylvanian and Permian time eventually uncovered Mississippian and older rocks there. Orogenic movements in the structural belt ceased during Permian time, and the region was leveled. Beginning in the Jurassic the region was warped downward, and deltaic, fluviatile, and shallow-water marine deposits accumulated in a sea that flooded across Texas from the Gulf of Mexico.

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North-Central New Mexico

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

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

NORTH-CENTRAL NEW MEXICO, AN ALTERNATE INTERPRETATION OF THE MISSISSIPPIAN

By AUGUSTUS K. ARMSTRONG

ABSTRACT

Initial Mississippian deposits are Kinderhook or Osage in age and are unconformable on older rock Precambrian in age. These Mississippian rocks were laid down during transgression of the sea across an Early Mississippian surface of low relief.

Interval A has an irregular thickness and distribution in south-central New Mexico and is the result of shallow epicontinental deposition on an irregular surface. In southwestern New Mexico interval A consists of shallow shelf carbonates more than 250 feet thick.

Intervals B and C are mostly limestones deposited in a shallow epicontinental sea. In the extreme south-central part of the State a deeper water carbonate basinal facies was deposited, and to the west a shallow shelf carbonate facies developed with the deposition of 800–900 feet of sediments. A major marine transgression, of shallow marine and evaporite deposits, in late interval B, a succeeding regression and a hiatus, and then another major marine transgression in middle interval C time are represented by a 10- to 60-foot-thick section that covered parts of north-central New Mexico. Low-lying land areas were present in northern and central parts of the State adjacent to the epicontinental seas.

Near the end of interval C (late Meramec), the northern and central parts of the State were slightly raised, and the sea withdrew to the southwest and south.

Sediments deposited during interval D (beginning in latest Meramec time) in the southwestern and south-central parts of the State were terrigenous material mixed with marine carbonates. Arenaceous carbonates of early Chester age are found in the northern and central parts of the State.

Near the end of the Mississippian Period north-central New Mexico became positive and the seas withdrew, exposing the Mississippian carbonate rocks to vadose weathering and extensive erosion.

INTRODUCTION AND REGION DEFINED

This chapter presents an interpretation of the Mississippian in New Mexico that is an alternate to that presented in chapter J. Although this interpretation is acknowledged in chapter J, some of the details are expanded here. The major difference between the two interpretations concerns the age assignment of rocks in north-central New Mexico — rocks that are assigned a Devonian(?), Kinderhook, Osage, and Meramec age (pl. 15, col. 88) in chapter J are assigned to the Osage, Meramec, and Chester in this chapter (table 3). In order to permit the reader to compare the differences, the Mississippian of north-central New Mexico is discussed in context with the rest of the State — excepting the southeastern and east-central parts (fig. 38).

PALEOGEOLOGY

UNITS UNDERLYING THE MISSISSIPPIAN

Precambrian igneous and metamorphic rocks underlie all the Mississippian in northern and central New Mexico.

LOWER BOUNDARY OF THE MISSISSIPPIAN

Mississippian rocks of late Osage age unconformably overlie Precambrian metamorphic and igneous rocks in north-central New Mexico, but in the southern part of

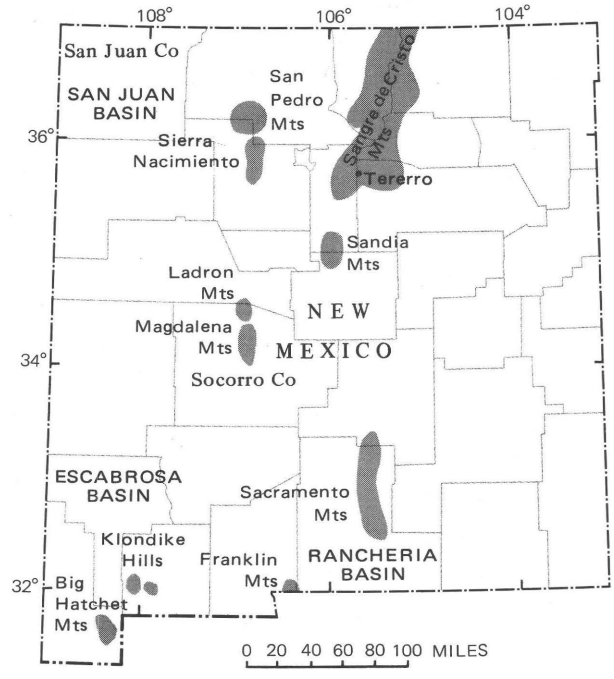


FIGURE 38. — Index map showing localities mentioned in text.

the State they rest unconformably on Upper Devonian sedimentary rocks. The hiatus is shortest in the southwestern part of the State; there, Kinderhook carbonate rocks unconformably overlie the Upper Devonian.

PALEOTECTONIC IMPLICATIONS

Probably as a result of epeirogenic uplift, marine waters withdrew from most, if not all, of New Mexico near the end of Devonian time. At the end of Devonian time New Mexico was nearly a level plain except for two low broad land areas composed of Precambrian rocks — the Zuni-Defiance highlands of northwestern New Mexico and the broad northeast-trending Pedernal highlands of the northeastern part of the State (fig. 39). Neither of the land areas was tectonically active in pre-interval D Mississippian time, and neither contributed appreciable amounts of terrigenous clastic material to the Mississippian epicontinental seas.

INTERVAL A

FORMATIONS INCLUDED

Stratigraphic units of Kinderhook age composing interval A are restricted to southern New Mexico (pl. 15, cols. 82-84; table 3; fig. 39). In southwestern New Mexico and adjacent Arizona the lower part of the Keating Formation is assigned to interval A. The *Caballero Formation* of Kinderhook age is present in south-central New Mexico (pl. 15, cols. 83, 84).

UPPER BOUNDARY OF INTERVAL A

The Keating Formation in southwestern New Mexico shows continuous sedimentation across the Kinderhook-Osage boundary (intervals A and B). Laudon and Bowsher (1949) indicated a short hiatus between the *Caballero Formation* and the overlying Lake Valley Limestone of Osage age in the areas of outcrop.

THICKNESS TRENDS

Interval A is thickest in the southwestern part of the State. In outcrops in south-central New Mexico, the *Caballero Formation* is generally less than 60 feet thick. Rocks of Kinderhook age are not recognized in the Franklin Mountains, where Lane (1974) reported conodonts of Late Devonian age at the top of the Percha Shale, unconformably overlain by the *Las Cruces Formation* having at its base a conodont fauna not older than late Osage age. In contrast to the interpretation given in chapter J, rocks of Kinderhook age are not recognized in north-central New Mexico.

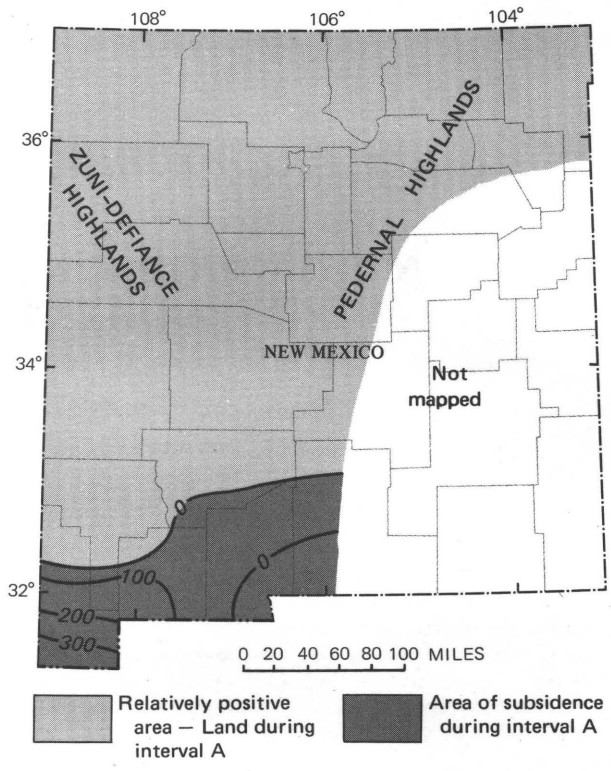


FIGURE 39. — Paleotectonic map of New Mexico during interval A (Kinderhook) showing restored isopachs (in feet).

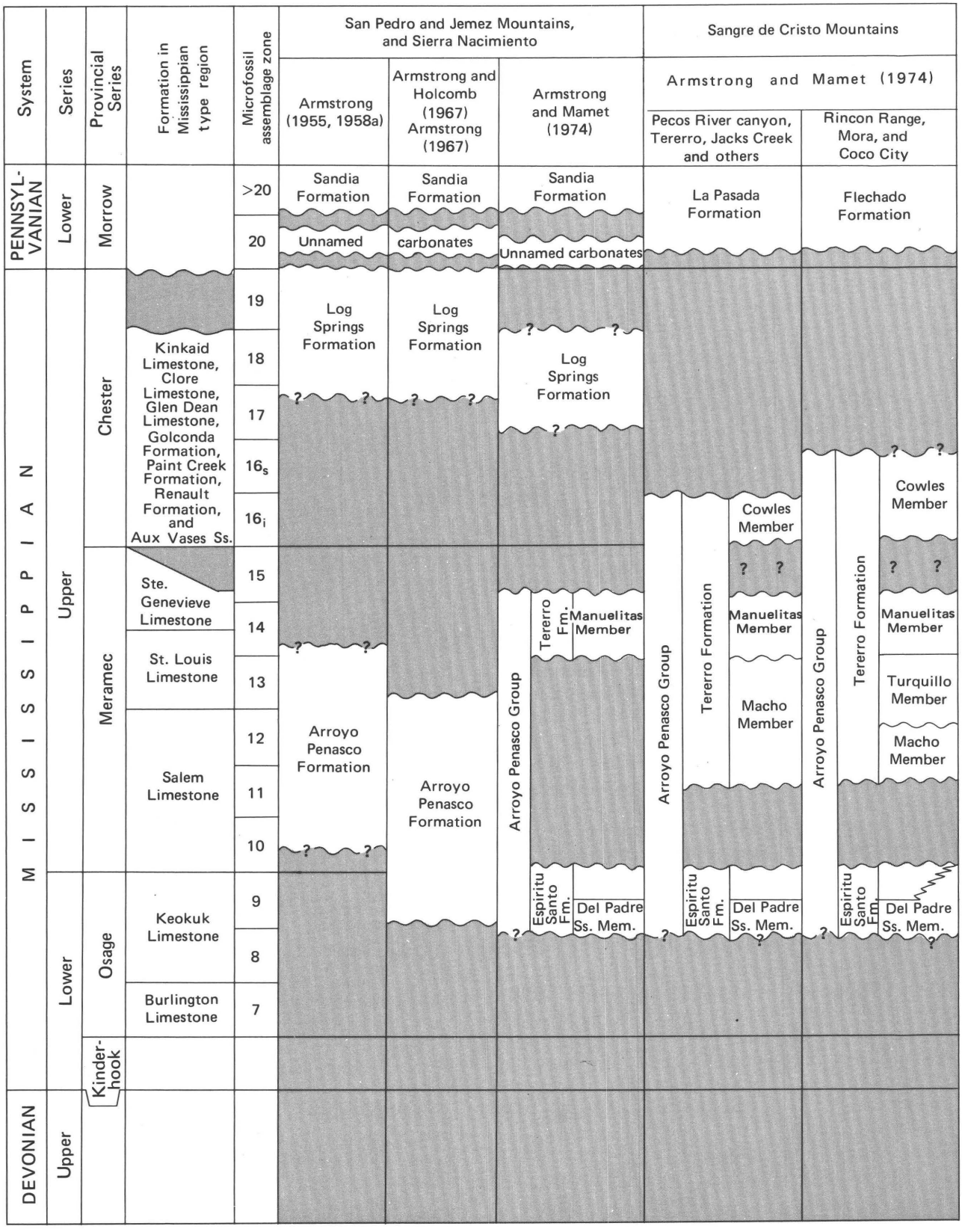
LITHOFACIES TRENDS

The Keating Formation is relatively chert-free carbonate consisting of limestone, dolomitic limestone, and lesser amounts of dolomite. The *Caballero Formation* of the south-central area is thin-bedded bioclastic limestone, admixtured and interbedded with argillaceous and arenaceous quartzose terrigenous clastics.

SOURCE AND ENVIRONMENTS OF DEPOSITION

Interval A of southwestern New Mexico is represented by carbonates deposited in shoal-water oolitic banks, cyclically alternating with stromatolitic microdolomites deposited in subtidal to intertidal environments. Brachiopods, corals, and Foraminifera are locally common and indicate that parts of the depositional cycles formed in the open marine environments. The *Caballero Formation* was deposited on a very shallow marine shelf on which marine organisms flourished. The sources of the terrigenous clastics in the *Caballero* were underlying Devonian rocks and stream-transported sediments brought in from the Precambrian terrain of central and northern New Mexico.

TABLE 3. — *Correlation chart of Mississippian rocks in north-central New Mexico, an alternate interpretation*



PALEOTECTONIC IMPLICATIONS

The northern three-fourths of the State was tectonically stable and nearly flat during interval A. In southwestern New Mexico a negative area or shallow basin developed which was to persist well into Chester time and which extended southward into Mexico and west into Arizona.

INTERVAL B

FORMATIONS INCLUDED

Named stratigraphic units included in New Mexico are the Leadville Limestone of the subsurface in San Juan County which may include some rocks of Kinderhook age, the *Caloso Formation* and Kelly Limestone of Socorro County (Armstrong, 1958b, 1962), the Lake Valley Limestone of south-central New Mexico, the *Las Cruces Formation*, Franklin Mountains, west Texas (fig. 38), and parts of the Keating and Hachita Formations of the Escabrosa Group, southwestern New Mexico (pl. 15, cols. 83–86). The basal 10–20 feet of the Arroyo Penasco Group, the Espiritu Santo Formation (Armstrong and Mamet, 1974) in north-central New Mexico, is very latest Osage age (table 3). Diagnostic microfossils from the Espiritu Santo Formation, identified by B. L. Mamet, are *Calcisphaera laevis* Williamson, *Carbonella* sp., *Earlandia* sp., *Endothyra* sp., *Inflatoendothyra* sp., *Inflatoendothyra "inflata"* (Lipina) OBJ (*I. eospiroides* Skipp, in McKee and Gutschick), *Latiendothyra* of the group *L. parakosvensis* (Lipina) [*L. skippae* (Armstrong)], *Parathurammina* of the group *P. cushmani* Suleimanov, *Parathurammina suleimanovi* Lipina, *Priscella* sp., *Septaglomospiranella* sp., *Septatournayella* sp., *Spinoendothyra bellicostata* Lipina), *Spinoendothyra spinosa* (Chernysheva), and *Tournayella* sp.

UPPER BOUNDARY OF INTERVAL B

In southwestern New Mexico the upper boundary of interval B is gradational within a sequence of massive encrinites in the upper half of the Hachita Formation (Armstrong, 1962, 1970).

Conodont studies of Lane (1974) and unpublished microfossil studies indicate continuous deposition from the *Las Cruces Formation* into the overlying *Rancheria Formation* of Meramec age in the Franklin Mountains of west Texas (pl. 15). The *Rancheria Formation* overlaps northward onto an apparent eroded surface of the Lake Valley Limestone of Osage age (Yurewicz, 1973).

The Lake Valley Limestone in its area of outcrop in central New Mexico is overlain unconformably by Pennsylvanian rocks (pl. 15). Younger Mississippian rocks undoubtedly were once present over much of the region (fig. 41) but were removed by Late Mississippian–Early Pennsylvanian erosion.

In north-central New Mexico the Espiritu Santo Formation is separated from the middle Meramec Tererro Formation by a hiatus (table 3) which represents most of Salem-age time (Armstrong and Mamet, 1974).

THICKNESS TRENDS

Interval B is present in its originally deposited thickness in southwestern New Mexico (fig. 40), where it is overlain by Meramec rocks of interval C. Elsewhere in the State it has been reduced by postdepositional erosion. Interval B is thick in southwestern New Mexico. It thins rapidly northward in west-central New Mexico, and northward from south-central New Mexico, by truncation beneath Pennsylvanian sedimentary rocks.

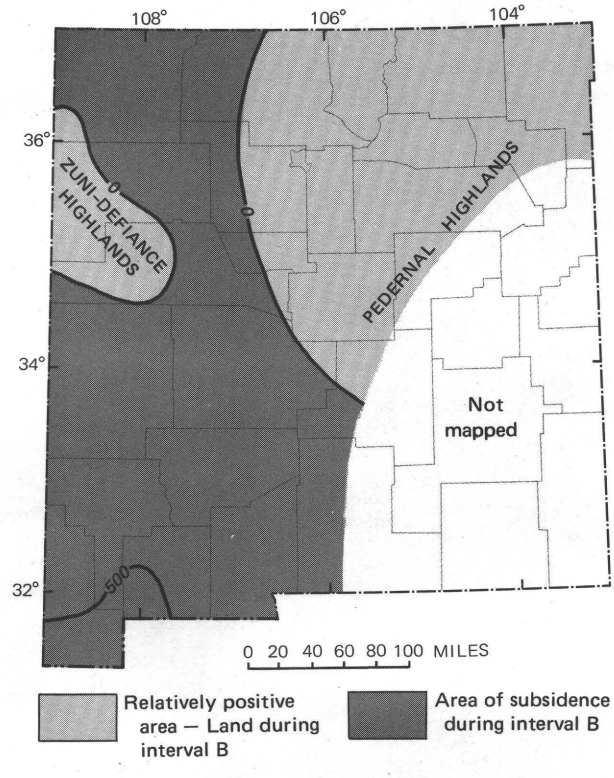


FIGURE 40. — Paleotectonic map of New Mexico during interval B (Osage) showing restored isopachs (in feet).

LITHOFACIES TRENDS

In south-central New Mexico, interval B consists mostly of cherty micrites and crinoidal limestones. In the Sacramento Mountains bioherms having fenestrate bryozoan cores are common (Laudon and Bowsher, 1941, 1949; Pray, 1958, 1961). In southwestern New

Mexico thick Osage sequences consist in the lower part of cherty dark pelletoid-micritic-bioclastic limestones and in the upper part of massive light-colored crinoidal bioclastic limestone and very minor amounts of dolomite. In west-central New Mexico, in the Magdalena, Lemitar, and Ladron Mountains, the *Caloso Formation* is calcispheroid pelletoid stromatolitic micrite, and the overlying Kelly Formation is crinoidal limestone. The Leadville Limestone in the subsurface of northwestern New Mexico is a cyclic sequence of crinoidal limestone, micritic limestone, and dolomite.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Western New Mexico was mostly covered by shallow epicontinental seas during interval B time. The seas transgressed eastward from the Cordilleran miogeosyncline through the Paradox basin of Utah, around the north flank of the Zuni-Defiance highlands, and into the areas of the present-day San Juan Basin. Seas advanced, also, from southern New Mexico northward, and at the end of Osage time they had flooded much of southern, western, and northern New Mexico west of the Pedernal highlands. Areas of Precambrian rocks were exposed in low-lying land areas, and these areas contributed the terrigenous clastic material to the basal sandstones of the Espiritu Santo Formation.

In southwestern New Mexico carbonate production kept pace with regional subsidence. The Osage rocks of south-central and southwestern New Mexico were deposited in the open sea; they contain very little dolomite and have no sedimentary structures indicating an intertidal environment. The *Caloso Formation* (Armstrong, 1958b, 1963) of west-central New Mexico, on the other hand, probably was deposited in or near the subtidal zone. The overlying Kelly Limestone was deposited in an open-marine shoal-water environment, probably a considerable distance from shore. The northward time equivalent to the Kelly Limestone, the Espiritu Santo Formation of north-central New Mexico, was deposited in a subtidal to supratidal environment (fig. 40). The Leadville Limestone of the subsurface of northwestern New Mexico was deposited near shore.

PALEOTECTONIC IMPLICATIONS

Epeiric subsidence, which began in late Kinderhook time in southern New Mexico, continued with only minor fluctuations during Osage time. Marine transgressions by late Osage time had advanced over most of New Mexico. At the end of interval C, marine waters had retreated off of central and northern New Mexico, exposing the thin carbonate sediments to vadose weathering.

INTERVAL C

FORMATIONS INCLUDED

In southwestern New Mexico the upper part of the Hachita Formation of the Escabrosa Group is assigned to interval C (Armstrong, 1962, 1970; Zeller, 1965). In south-central New Mexico and extreme western Texas, the *Rancheria Formation* is assigned to interval C; it contains megafossils (Laudon and Bowsher, 1949) and conodont faunas (Lane, 1974), both of Meramec age.

The Manuelitas Member of the Tererro Formation of the Arroyo Penasco Group (Armstrong and Mamet, 1974) of the San Pedro Mountains and Sierra Nacimiento in north-central New Mexico contains a microfauna of Meramec age. In the Sangre de Cristo Mountains, the Turquillo and Manuelitas Members of the Tererro Formation (table 3) contain a rich microfossil assemblage of Meramec age.

The Turquillo and Manuelitas Members of the Tererro Formation are shoaling water arenaceous carbonate rocks deposited over the northern part of the State during a middle Meramec marine transgression.

Foraminifers are abundant in the Turquillo Member. In particular, *Endothyranopsis* of the group *E. spiroides* (Zeller), *Eoendothyranopsis scitula* (Toomey), *Eoendothyranopsis hinduensis* Skipp (in McKee and Gutschick, 1969), *Eoendothyranopsis prodigiosa* (Armstrong), and primitive *Endothyranopsis* and biseriamminids are present. The dasyclad *Koninckopora* is widespread. The fauna indicate a Meramec, Salem/St. Louis boundary age equivalent.

The Manuelitas Member is rich in Foraminifera. The association of *Eoendothyranopsis macra* (Zeller), *Eoendothyranopsis* of the group *E. ermakiensis* (Lebedeva), *Eoendothyranopsis prodigiosa* (Armstrong), *Endothyranopsis* of the group *E. compressa* (Rauzer-Chernoussova and Reitlinger), and *Globoendothyra paula* (Vissarionova) indicates a Meramec, St. Louis age equivalent (Mamet, in Armstrong and Mamet, 1974).

UPPER BOUNDARY OF INTERVAL C

In southwestern New Mexico the encrinites of the Hachita Formation in interval C are overlain gradationally by micrite, oolite, and bioclastic calcarenite of the Paradise Formation in interval D.

The upper boundary of interval C in south-central New Mexico and west Texas appears to be gradational between the *Rancheria Formation* and the Helms Formation. In the Sangre de Cristo Mountains of north-central New Mexico, the Manuelitas Member (interval C) is disconformably overlain by the Cowles Member of the Tererro Formation, of early Chester age. In the

Sierra Nacimiento and the San Pedro Mountains, the Manuelitas Member is overlain disconformably by the Chester age red beds of the Log Springs Formation (table 3).

THICKNESS TRENDS

Interval C is thickest in south-central and southwestern New Mexico (fig. 41). In north-central New Mexico it has been extensively eroded in Late Mississippian–Early Pennsylvanian time, and ranges in thickness from 0 to 55 feet.

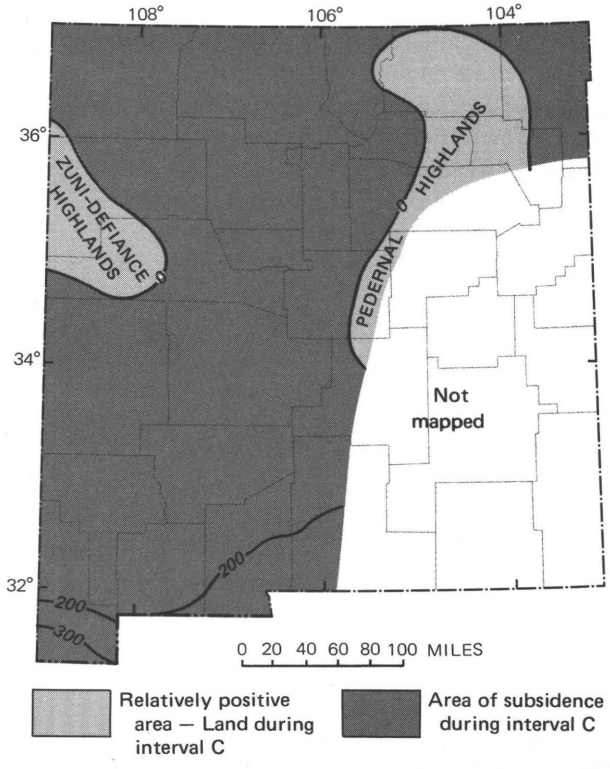


FIGURE 41. — Paleotectonic map of New Mexico during interval C (Meramec) showing restored isopachs (in feet).

LITHOFACIES TRENDS

Interval C in southwestern New Mexico is light-gray massive encrinite, which is oolitic and pelletoid. The *Rancheria Formation* in south-central New Mexico is cherty pelletoid dolomitic arenaceous micritic limestone. Terrigenous material is mostly silt-sized quartz and makes up 20–30 percent of the limestone. Interval C of north-central New Mexico consists of recrystallized limestone, micrite, oolitic limestone, and microdolomite, all containing varying amounts of silt to fine quartz grains.

SOURCES AND ENVIRONMENTS OF DEPOSITION

Interval C represents the time of maximum marine transgression in New Mexico. Only the residual Zuni-Defiance and Pedernal highlands stood above the surrounding seaways. Most of interval C in southwestern New Mexico was deposited on an open-marine shelf in relatively high-energy shoal-water environment. The *Rancheria Formation* of south-central New Mexico contains features indicating deposition in relatively deep water (Yurewicz, 1973; Lane, 1974).

A marine transgression over north-central New Mexico occurred in Meramec time with the deposition of oolitic sand and pelletoid-micritic limestone and fine quartz sand in shallow water offshore. The quartz sand is believed derived from the Pedernal highlands to the east (fig. 41).

PALEOTECTONIC IMPLICATIONS

New Mexico was tectonically stable to slightly negative during interval C. The southern part of the State was the area of greatest subsidence. Relatively thick carbonate sections accumulated in the Escabrosa-Rancheria basins. In the north-central part of the State, subsidence was slight. The marine water retreated from north-central New Mexico in late Meramec time.

INTERVAL D

FORMATIONS INCLUDED

Interval D is confined to the south-central, southwestern, and north-central parts of New Mexico. In south-central New Mexico it consists of the Helms Formation and in southwestern New Mexico, of the Paradise Formation. The Paradise Formation (pl. 15, col. 84) includes beds of latest Meramec age as well as beds of late Chester age. The Cowles Member in the Sangre de Cristo Mountains was assigned by Baltz and Read (1960) as Meramec in age on the basis of stratigraphic interpretation (pl. 15, col. 88). However, the Cowles contains primitive *Neoarchaediscus* and *Zellerina*, indicating that the unit is early Chester equivalent and assignable to interval D. The Log Springs Formation is known only in the Sierra Nacimiento, Sandia, Jemez, and San Pedro Mountains, where it is 6–50 feet thick and rests with a marked unconformity on various beds of the Arroyo Penasco Group. It is composed of continental clastic red beds and is post-early Chester in age and, inasmuch as it is overlain with a hiatus by Morrow age beds, it must be Chester equivalent.

UPPER BOUNDARY OF INTERVAL D

Interval D is overlain unconformably by Pennsylvanian rocks of Morrow and Atoka age. In the Franklin Mountains thin-bedded limestones in the Helms are overlain by massive Morrow limestone along a paracon-

formity. In the Sacramento Mountains the top of the Helms Formation is an angular unconformity at the base of Pennsylvanian pebble conglomerates.

The Paradise Formation in the southwestern part of the State in the Klondike Hills and Big Hatchet Mountains is unconformably overlain by thick-bedded Pennsylvanian quartzitic sandstone and pebble conglomerate, which contain impressions of "*Lepidodendron.*" These Pennsylvanian beds are overlain by marine carbonate rocks of Morrow age.

THICKNESS TRENDS

The originally deposited thickness of interval D is everywhere modified by Late Mississippian and Pennsylvanian erosion. Interval D sediments at one time undoubtedly extended farther north than shown by the zero isopach in figure 42, and probably covered much of the State.

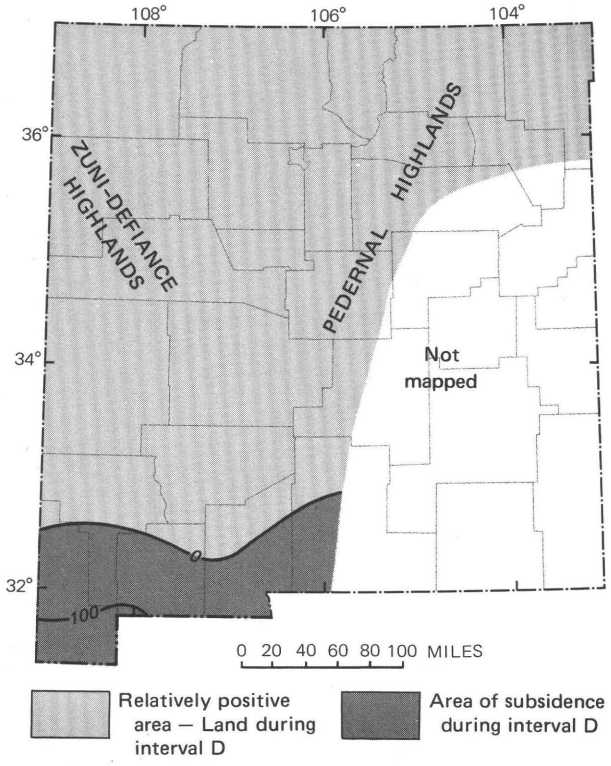


FIGURE 42. — Paleotectonic map of New Mexico during interval D (Chester) showing partially restored isopachs (in feet).

In south-central New Mexico the Helms Formation is generally less than 100 feet thick, and it thins northward to zero within the Sacramento Mountains (Pray, 1961, p. 69). The Paradise Formation is progressively thicker and younger at its top southward in southwestern New Mexico. In north-central New Mexico the lower Chester Cowles Member of the Tererro Formation is less than 30 feet thick. The Log Springs Formation of late Chester age in the Sierra Nacimiento and in the San Pedro Mountains is generally less than 50 feet thick.

LITHOFACIES TRENDS

The Helms Formation consists of greenish-gray silty calcareous shale and, near the top, thin-bedded pelletoid to micritic argillaceous limestones. The Paradise Formation is a complex suite of thin- to medium-bedded ooid pelletoid micritic and bioclastic limestone interbedded with marl, shale, and quartz siltstone. In north-central New Mexico, in the Sangre de Cristo Mountains, the lower Chester Cowles Member of the Tererro Formation consists of siltstone, calcareous siltstone, shale, and pelletoidal fine-grained silty limestone and ostracod micrite. The Log Springs Formation is composed of clastic red beds, hematitic shale, siltstone, sandstone, and conglomerate of continental origin.

SOURCE AND ENVIRONMENTS OF DEPOSITION

Shale and silty shale in the lower two-thirds of the Helms Formation are interpreted as deposited in moderately deep water. More fossiliferous micritic thin-bedded limestone near the top of the formation indicates shallower water, perhaps closer to the shore.

A shallow-water nearshore environment characterized the southwestern part of the State in which terrigenous clastic sediments and marine carbonates were sorted and mixed by currents and waves. Abundant ooid limestone suggests periods of shoaling or development of tidal bars. An abundant brachiopod, echinoderm, and bryozoan fauna indicates well-circulated clean water. Siltstone and sandstone beds commonly have abundant plant remains, indicating nearby land, probably to the north. The siltstone and pelletoidal silty limestone of the Cowles Member of the Tererro Formation represent shallow-water marine sedimentation with a terrigenous source probably from the Pedernal highlands to the east. The Log Springs Formation is formed in part from a reworked regolith and in part from stream transported material derived from elevated adjacent highlands composed of Mississippian carbonate and exposed Precambrian terrane (Armstrong, 1967; Armstrong and Mamet, 1974).

PALEOTECTONIC IMPLICATIONS

New Mexico was tectonically nearly dormant for most of interval D. The southern part of the State sank slightly, and the northern part was stable but marked by a marine transgression, depositing the lower Chester Cowles Member of the Tererro Formation. By late Chester time probably all of New Mexico was emergent except for the south-central region. In late Chester

time, northern and central New Mexico were folded and faulted, and the Mississippian carbonate rocks were removed from broad areas (Armstrong, 1958b, 1962, 1967; Armstrong and Mamet, 1974).

In late Chester time interbedded limestone and gypsum in the Espiritu Santo Formation of late Osage age in the Sangre de Cristo Mountains area were removed by vadose waters, and adjacent overlying carbonates collapsed forming the breccia that constitutes the Macho Member of Tererro Formation. Sinkholes, solution cavities, and pisolites were formed. In the southern Sierra Nacimiento a 30- to 50-foot-thick accumulation of rounded chert nodules, Precambrian rock pebbles and cobbles, terra rosa soils, and hematitic clays, all belonging to the Log Springs Formation, formed by weathering on top of the Mississippian carbonates. Subsequently, the Log Springs deposits were covered beneath marine carbonates of Morrow age (Armstrong, 1967).

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

The most complete record of Mississippian deposition is in the Big Hatchet Mountains in southern New Mexico, where 1,400 feet of shallow-shelf, predominantly carbonate rocks is present. Disconnected areas of Mississippian rocks in central and northern New Mexico are remnants of extensive sheets now dissected and beveled by erosion. Although relatively thin, the Arroyo Penasco Group of north-central New Mexico spans a considerable part of Mississippian time.

PALEOTECTONIC IMPLICATIONS

The tectonic pattern for pre-interval D time was of slight orogenic activity. The most notable events were slow subsidence, which was greatest in the south-central and southwestern parts of the State, and flooding by epicontinental seas. The Arroyo Penasco Group of north-central New Mexico records three marine transgressions and regressions. Subsidence ceased in early Chester time in northern and central New Mexico and was followed by regional uplift and gentle folding. Continued uplift and folding brought Mississippian deposition to an end in southwestern New Mexico by late Chester time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

Pennsylvanian rocks unconformably overlie the Mississippian at most places in New Mexico. In north-central New Mexico Mississippian rocks in the San Pedro Mountains, in the Sierra Nacimiento, and in the Sangre de Cristo Mountains are overlain by the Pennsylvanian Sandia Formation. The Pennsylvanian Molas Formation overlies the Leadville Limestone in the subsurface of San Juan County.

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Arizona

By EDWIN D. MCKEE

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–L

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

ARIZONA

By EDWIN D. MCKEE

ABSTRACT

Areas of land and sea were well established in Arizona early in Mississippian time and apparently remained in about the same places throughout most of that period. The size of the landmasses fluctuated greatly, however, as the sea transgressed and regressed across the region at least three times. Positive elements that persisted were the Defiance-Zuni element in the northeast corner of the State and probably the Ensenada element in the southwest. Other parts of the State were shelves, periodically covered with shallow water, that merged into geosynclines to the northwest and southeast. The geosynclines were covered by deeper, more permanent water bodies.

Seas began to advance across a broad land area in Arizona from the northwest and southeast during interval A but did not reach a maximum size until sometime in interval B when they connected across central Arizona and covered most of the State. During that interval, more than 500 feet of sediment was deposited in the northwest corner and more than 700 feet in the southeast corner. During interval C the sea regressed and was considerably restricted. The record of interval D is scant, but remnants of Chester age rocks suggest that the sea again advanced and later retreated across the shelf; it remained long enough for the accumulation of thin deposits probably over large areas.

Mississippian rocks in Arizona are mostly carbonates. Dolomite formed large in the shallow-water areas bordering landmasses, and limestone formed mostly farther offshore. Terrigenous detrital material is so scarce in all intervals that it seems very doubtful that land areas in this region were appreciably elevated at any time during the Mississippian. Land probably was low and flat throughout the period.

Rocks underlying the Mississippian System, largely of Devonian age, are everywhere separated from the Mississippian by an unconformity representing a brief hiatus. The erosion surface shows only mild relief; apparently, the uplift that took place at this time was regional, but the elevation of the land remained low. Following Mississippian deposition, a period of somewhat more severe erosion took place; in many localities a karst topography developed, and in large areas the upper units of the Mississippian were entirely removed.

REGION DEFINED

The region discussed in the following section is restricted to the State of Arizona. It consists of the southwestern part of the Colorado Plateau, the mountain area of central Arizona, and a part of the Basin and Range province in southern and western Arizona (fig. 43). Rocks of Mississippian age occur in nearly all parts of the State, but in the southwestern and central parts of the State, only widely scattered and poorly preserved remnants are exposed; so the record from these areas leaves much to be desired.

During Mississippian time the Arizona region included a margin of the Cordilleran miogeosyncline to the northwest and the upper end of the Sonoran geosyncline to the southeast (fig. 44). Between these features was a mildly negative shelf area that was in-

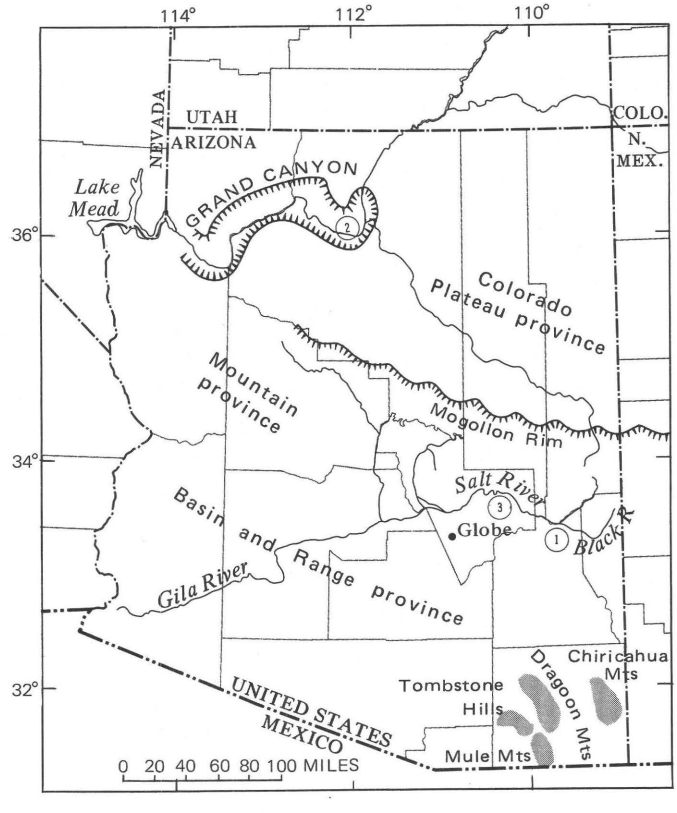

FIGURE 43. — Areas and localities mentioned in the text: (1) Black River Crossing, (2) Bright Angel Trail, (3) Salt River Canyon.

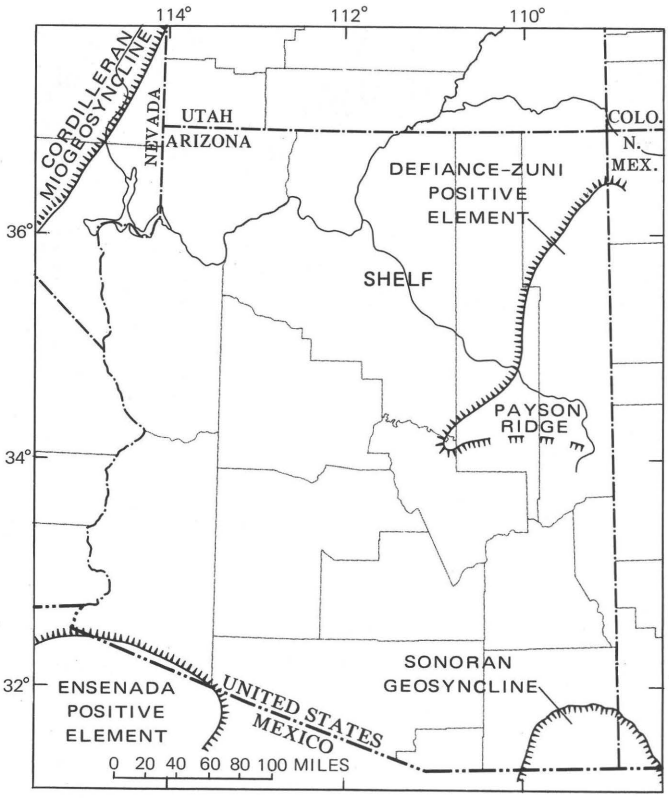

FIGURE 44. — Tectonic elements during Mississippian time. Boundaries are generalized.

termittently beneath the sea and, at the time of maximum marine invasion, was largely submerged. Only a few small positive elements prevailed throughout the period, and these apparently were too low and too restricted to furnish any appreciable amount of detritus. Sediments are almost exclusively carbonate.

PALEOGEOLOGY

UNITS UNDERLYING MISSISSIPPIAN

Throughout Arizona where Mississippian rocks occur, they are underlain by strata of Devonian age, except for a few places in the eastern Grand Canyon area where Cambrian strata are directly beneath Mississippian: one small area near the southwest margin of the Defiance-Zuni positive element where Precambrian rocks underlie them and at two places in east-central Arizona where Mississippian strata were deposited on small hills of Precambrian quartzite that were islands in the Devonian (Teichert, 1965, p. 49).

Formation names applied to the Devonian of Arizona differ largely according to geographic locality. In the northeastern corner of the State, the name Ouray Limestone, from Colorado sections, is used in subsurface work. Throughout Grand Canyon the name Temple Butte Limestone is applied, and in central and most of southern Arizona Martin Limestone is used (pl. 15). In extreme southeastern Arizona the *Portal Formation* directly underlies the Mississippian (Armstrong, 1962, p. 15).

LOWER BOUNDARY OF MISSISSIPPIAN

In the Grand Canyon area of northern Arizona, Devonian and Cambrian strata were widely eroded and beveled before deposition of Mississippian sediment. The surface of erosion shows no conspicuous dissection and no more than slight angular discordance. Uplift of appreciable magnitude could not have taken place at the end of Devonian time (McKee and others, 1969), although apparently the uplift was sufficient for the sea to withdraw from much or all of Arizona.

The Mississippian-Devonian contact in central Arizona, southeast and south of Grand Canyon, is clearly a regional disconformity. The general lack of coarse fragments in basal rocks of the Redwall Limestone at the base of the Mississippian, the sparse amount of reworked insoluble residue, and the rather slight dissection of the Devonian surface all suggest a short interval of mild erosion and a negligible amount of diastrophism during the hiatus.

Details of the stratigraphic record in southern Arizona are more poorly known than those of the areas farther north, but the record seems generally similar. A hiatus between the earliest Mississippian and latest Devonian rocks is recognized (Armstrong, 1962, p. 14), but no marked erosion or discordance was developed between them.

PALEOTECTONIC IMPLICATIONS

The paleogeologic map of units directly beneath the Mississippian System (pl. 2) shows that downwarping occurred and sediment accumulated during Devonian time in nearly the same places as during the Mississippian. The region was elevated slightly and was emergent at the end of Devonian or beginning of Mississippian time.

INTERVAL A

FORMATIONS INCLUDED

Interval A in Arizona consists almost entirely of rocks of Kinderhook age. Part of the Whitmore Wash Member of the Redwall Limestone (pl. 15, col. 116) in the western Grand Canyon and Black River Crossing areas has been dated as Kinderhook in age on the basis of brachiopods and endothyrids (McKee and others, 1969, p. 30, 36). The Escabrosa Limestone, farther south, also contains rocks of Kinderhook age, determined on the basis of conodonts (Ethington, 1965, p. 568) found near Globe. Some rocks of Kinderhook age

are probably in the lower part of the Keating Formation of the Escabrosa Group in southeastern Arizona (A. K. Armstrong, oral commun., 1971), but all the Keating is arbitrarily assigned to interval B. Apparently, rocks of interval A represent the earliest stage in a transgression of the Mississippian sea across Arizona.

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A, as determined from fossils, lies between the highest zone of Kinderhook and the lowest zone of Osage faunas. In the Redwall this boundary is entirely within the Whitmore Wash Member and cannot be selected on the basis of lithology. In the Escabrosa the boundary has not been determined precisely but probably is near the base.

THICKNESS TRENDS

Interval A is limited laterally by sections in which fossils of Osage age occur at or near the formation base. Interval A in northwestern Arizona seems to thin eastward from 40 to 50 feet near the Nevada line to zero about 30 miles farther east (pl. 3-A). In east-central Arizona the rocks of comparable age are also less than 100 feet thick; they approach a zero thickness not far north of the Salt River Canyon.

LITHOFACIES TRENDS

Interval A is composed entirely of carbonate rock in its limited Arizona exposures (pl. 3-B). Although these rocks apparently formed at or near a margin of the sea, they are almost completely free of terrigenous material.

SOURCES AND ENVIRONMENTS

A seaway extended a short distance into the western Grand Canyon area from Nevada, and a second seaway occupied the eastern part of central Arizona and connected with open water either to the southeast or to the southwest (pl. 3-A). Deposits in the connecting part of the sea seem to be missing; to the east, the nearest rocks of definite Kinderhook age are in southwestern New Mexico (lower member of Escabrosa Limestone (Zeller, 1965) and lower part of Keating Formation (Armstrong, 1970)) and in south-central New Mexico (*Caballero Formation*). Clastic grains of calcite that form most of the carbonate rock indicate shallow seas and an environment of moderate energy in the waters that advanced eastward and northward during Kinderhook time. Brachiopods and endothyrids, locally common, suggest seawater of normal salinity (pl. 11, fig. 1). Apparently, no areas of high relief were nearby, and carbonate rock formed nearly everywhere on most of the surface upon which deposition occurred.

PALEOTECTONIC IMPLICATIONS

Interval A represents the first stage of a major transgression across parts of Arizona (pl. 10, fig. 1). Although interval A was restricted to small areas of deposition, the subsidence and consequent transgression continued to a maximum in later Mississippian time during which much of the State was submerged by marine waters.

INTERVAL B

FORMATIONS INCLUDED

Most of the Mississippian rocks in Arizona are of Osage age and are assigned to interval B. In the northern and central parts of the State, this interval consists of most of the Whitmore Wash Member (except certain basal beds of Kinderhook age assigned to interval A) and all of the Thunder Springs and Mooney Falls Members (McKee, 1963) of the Redwall Limestone (pl. 15, col. 116). Farther south, it consists of the Keating Formation and the lower part of the Hachita Formation of the Escabrosa Group (Armstrong, 1962, 1970).

UPPER BOUNDARY OF INTERVAL B

In northern Arizona the contact between intervals B and C is placed at the boundary between the Mooney Falls and Horseshoe Mesa Members of the Redwall Limestone. Throughout western Grand Canyon this contact, as determined by faunal zones (McKee and others, 1969, p. 129), is only a short distance above the Osage-Meramec boundary. In other areas where the Redwall is present, this contact, because it is readily recognizable, is also used even though faunal control is lacking.

In southeastern Arizona the boundary between rocks of Osage age and those of Meramec age is the contact of the lower and upper parts of the Hachita Formation of the Escabrosa Group as described by Armstrong (1962) at Blue Mountain in the Chiricahua Mountains. This boundary is recognized as the division plane between intervals B and C.

In the Dragoon Mountains and Tombstone Hills and possibly in the Mule Mountains of southern Arizona, the boundary between intervals B and C probably is near the top of the Mississippian sequence. In these localities, rocks believed to be younger than Osage, but of pre-Chester age, are tentatively recognized near the top of the main rock unit (Gilluly and others, 1954, p. 12), but the thickness of strata of Meramec age is not known. The main rock unit is assigned entirely to interval B for purposes of preparing the isopach maps. The presence of Meramec strata is indicated by a plus sign (+) at appropriate control points on the map of interval C (pl. 5-A).

THICKNESS TRENDS

Interval B of the Mississippian System is represented in Arizona by an isopach pattern basically like that of other Paleozoic units. In the northeastern part of the State, a zero isopach outlines the Defiance-Zuni positive element (pl. 4-A). The thickness of the interval increases northwestward from this line toward the boundary of Arizona with Utah and Nevada, and also southward toward the southeastern corner of the State. Maximum thicknesses for interval B of about 600 and 700 feet are recorded in opposite corners of the State.

LITHOFACIES TRENDS

Almost all of interval B in Arizona is composed of carbonate rock (pl. 4-B), the only noteworthy exception being chert that occurs as abundant thin beds or layers of nodules within some rock units. Evaporites are absent and terrigenous detrital sediments of any type are very uncommon either in pure form or as a constituent of the carbonates. The carbonate rock in some areas is limestone and in others is dolomite (fig. 45). The dolomite apparently is largely restricted to nearshore areas where the water was shallow and had more than average salinity (McKee and others, 1969, chap. 2, figs. 6, 9, 10).

SOURCES AND ENVIRONMENTS

Within northern Arizona, the transgression that began during interval A time continued during the time of interval B, and the rocks of the upper part of the Whitmore Wash Member of the Redwall were

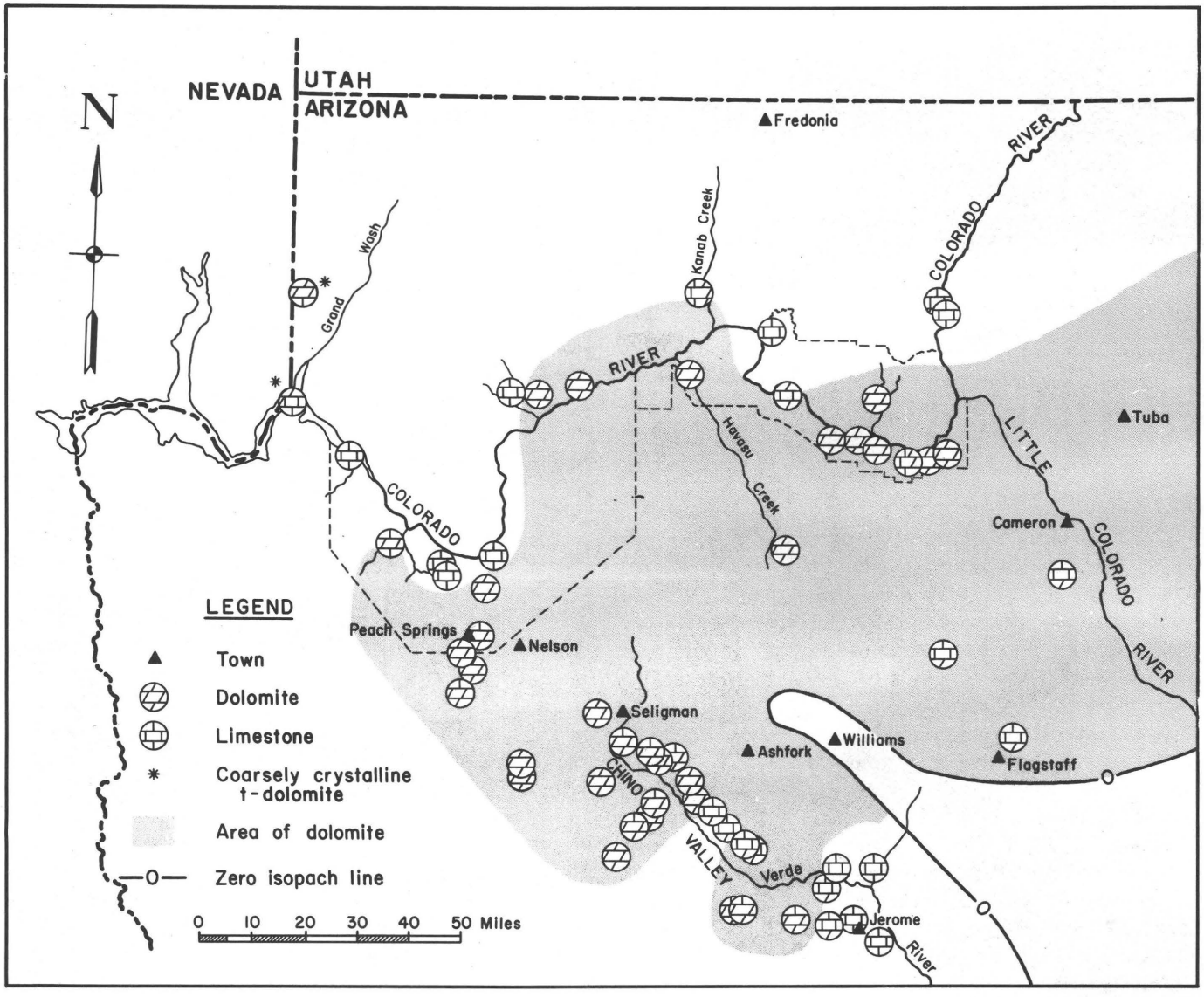

FIGURE 45. — Distribution of limestone and dolomite in Whitmore Wash Member of Redwall Limestone in northwestern Arizona. Rock symbols (or symbol) in each circle indicate approximate proportions of carbonate-rock types in a stratigraphic section. From McKee, Gutschick, and others (1969, p. 25); reprinted with permission of The Geological Society of America.

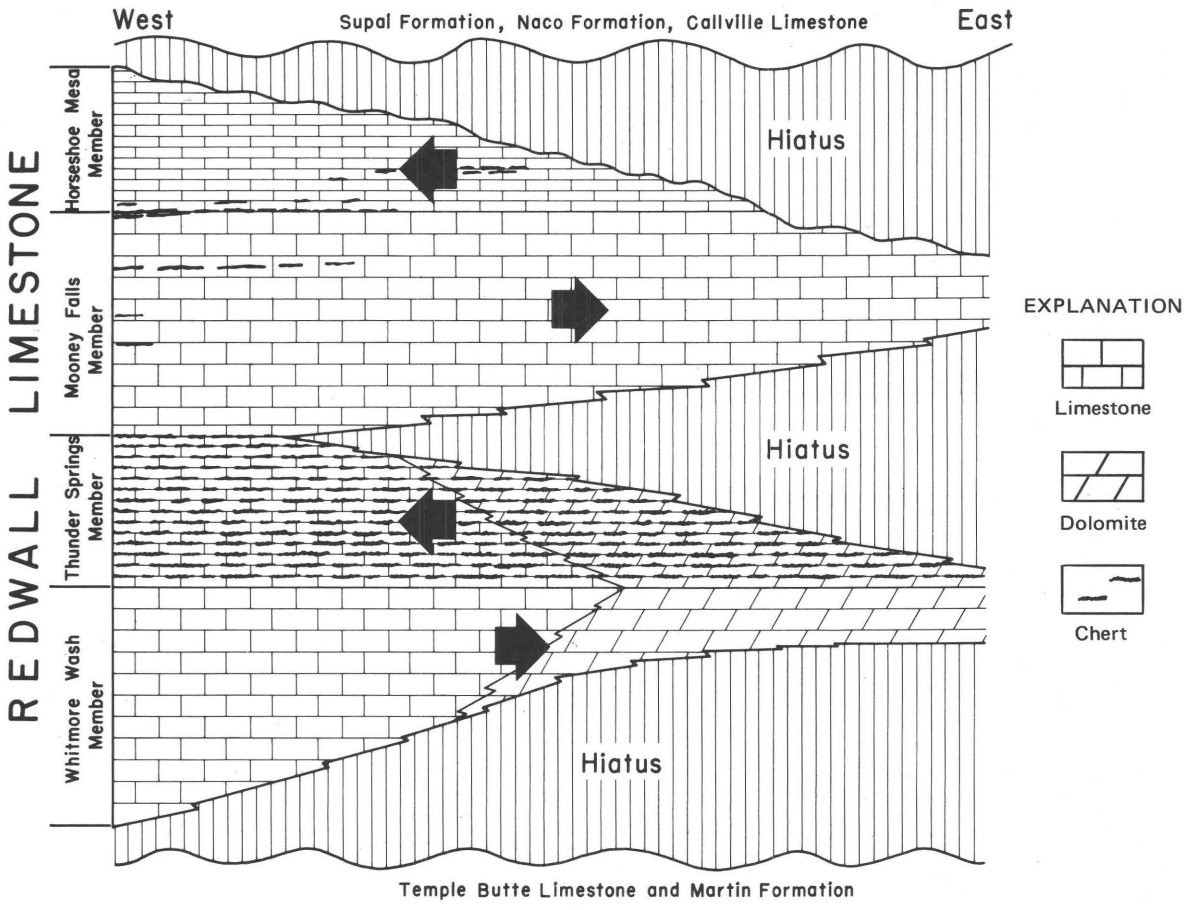

FIGURE 46. — Schematic generalized east-west cross section showing stages of transgression and regression as represented by members of the Redwall Limestone. Horizontal distance is about 200 miles; maximum total thickness is about 800 feet. From McKee, Gutschick, and others (1969, p. 63); reprinted with the permission of The Geological Society of America.

deposited. This transgression was followed by a regression of the strandline during which the Thunder Springs Member was deposited. A second major transgression followed and the Mooney Falls Member was deposited. Thus, two large transgressions and a major regression are recorded in the rocks of interval B (fig. 46) in northern Arizona.

The sea transgressed across Arizona from two directions, the northwest and southeast. The southwestern limit of the seaway is unknown, but, on the basis of thickness trends, it probably was in northwestern Mexico (McKee, 1951, p. 485).

Because terrigenous sediments are very uncommon in rocks assigned to interval B of the Mississippian in Arizona, the absence of nearby high landmasses or of the deltas of major river systems may be assumed (pl. 12, fig. 2). The carbonate rocks of interval B accumulated under normal marine conditions (pl. 11, fig. 2). They contain a high proportion of clastic carbonate particles — both biogenic and inorganic — which suggest various degrees of current and wave energy on the sea floor, but they also contain much aphanitic limestone that is interpreted as having formed from lime mud under quiet water conditions. The contrasting environments apparently alternated in a cyclic succession, possibly the result of changes in sea level (McKee, 1960).

PALEOTECTONIC IMPLICATIONS

Interval B represents the part of the Mississippian System in Arizona which was characterized by maximum sinking of shelves in the northwestern and southeastern parts and by the greatest extent of the seas. The simple depositional convergence of interval B suggests that the Defiance-Zuni positive element of northeastern Arizona remained emergent, but it probably was not highly elevated at this time (pl. 10, fig. 2). The Payson ridge, a prong extending southwestward from the Defiance-Zuni positive area, tended to divide the interval B sea and was completely covered only during the time of Mooney Falls deposition (pl. 9-F, sec. *N–N'*; McKee and others, 1969, p. 571–574).

INTERVAL C

FORMATIONS INCLUDED

Interval C of the Mississippian System in Arizona consists of the Horseshoe Mesa Member of the Redwall Limestone (pl. 15, col. 116) in the Grand Canyon area. Although the extreme uppermost part of the Mooney Falls Member of the Redwall contains fossils of Meramec age in western Grand Canyon, the entire member is assigned to interval B in this study because no practical method was found of locating consistently the B-C boundary within the member. In southeastern Arizona, interval C consists of the upper part of the Hachita Formation of the Escabrosa Group (Armstrong, 1962, p. 5; 1970), and farther west in southern Arizona it consists of beds of small but otherwise unknown thickness containing a probable Meramec fauna (Gilluly and others, 1954, p. 12). In central Arizona along the Mogollon Rim interval C once was present but has been removed by erosion as shown by residual fragments of rock from the Horseshoe Mesa Member of the Redwall preserved in many places as residuum on a karst surface (McKee and others, 1969, p. 70–72).

UPPER BOUNDARY OF INTERVAL C

The upper boundary of interval C, nearly everywhere that this interval occurs in Arizona, is a surface of erosion. In the Grand Canyon area, thickness trends suggest a fairly even surface, though in one place (Bright Angel Trail) a remnant of limestone of Chester age, which is assigned to interval D, is locally preserved above limestone of interval C (pl. 9-F, sec. *I–I'*).

In extreme southeastern Arizona the base of the mudstone-bearing Paradise Formation of interval D forms the upper boundary. Elsewhere, Pennsylvanian rocks of Morrow age, generally consisting of mudstone and sandstone, overlie limestone of interval C.

THICKNESS TRENDS

Interval C is distributed generally in the same areas as underlying interval B, although C is far less extensive. The maximum recorded thicknesses are in the northwest and southeast parts of the State and are about 200 feet (pl. 5-A).

Interval C originally must have been more extensive than at present. Apparently, its margins have been eroded back across much of central and western Arizona. South of Grand Canyon along the Mogllon Rim interval C is absent, but in southeastern Arizona and from there eastward into New Mexico, the interval is represented by Meramec age limestone of very irregular thickness. Much of the interval in that area has been removed by erosion (Armstrong, 1962, p. 24).

LITHOFACIES TRENDS

Interval C in Arizona consists entirely of carbonate rock — mostly aphanitic limestone. Because no terrigenous sediment is present, lithofacies trends are not apparent on plate 5-B.

SOURCES AND ENVIRONMENTS

Seaways extended southeastward from the northwestern corner of the State and northwestward from the southeastern corner during interval C.

The Defiance-Zuni positive element in northeastern Arizona almost certainly was emergent during interval C, as in earlier Mississippian time, but probably was very low.

The fine-grained carbonate sediment comprising interval C in Arizona is interpreted as having originally been a lime mud deposited in a marine environment of low energy (pl. 11, fig. 3). Evidence is available (McKee and others, 1969, p. 590) that the mud formed during the regressive stage of a shelf sea following the major transgression represented by the upper part of interval B.

PALEOTECTONIC IMPLICATIONS

No major tectonic events in Arizona are indicated by the sedimentary record of interval C. Although a marine environment prevailed in some parts of the region during an early stage, the region was elevated and the sea had withdrawn entirely long before the end of interval C time. A lack of detrital sediment in deposits of this interval suggests that no land areas were being differentially uplifted in the region. Areas that both previously and subsequently were negative or positive apparently were stable at this time (pl. 10, fig. 3).

INTERVAL D

FORMATIONS INCLUDED

Rocks of Chester age, included in interval D, once had widespread distribution in Arizona, but have subsequently been largely removed by pre-Pennsylvanian or Early Pennsylvanian erosion.

In northern Arizona a single isolated element of Chester rocks occurs at the top of the Redwall Limestone near the Bright Angel Trail in Grand Canyon (pl. 9-F, sec. *I–I'*). This 6½-foot limestone unit has been dated on the basis of both brachiopods and endothyrids (McKee and others, 1969, p. 74).

At Black River Crossing in east-central Arizona, a second isolated remnant possibly is also assignable to interval D. Residual chert filling and overlying Redwall karst structure at this locality contain the coral *Aulina* which is considered by Sando (1963) to be of either Meramec or Chester age.

In extreme southeastern Arizona the Paradise Formation is well established as Chester in age (Hernon, 1935, p. 657). This formation is assigned to interval D.

UPPER BOUNDARY OF INTERVAL D

The upper boundary of interval D in southeastern Arizona and adjacent parts of New Mexico is a surface of erosion, but the hiatus between the Chester and the lowest Pennsylvanian (Morrow) rocks which lie above was thought by Armstrong (1962, p. 27) to have been very short. He believed "only latest Chester and very earliest Pennsylvanian time" to be unrepresented in the rocks. In the Chiricahua Mountains, Ariz., the basal Pennsylvanian is described as a brown to gray marine mudstone overlain by massive limestones; this sequence sharply overlies the Paradise Formation of interval D.

THICKNESS TRENDS

Interval D extends into southeastern Arizona and southern New Mexico from the south and apparently was deposited in the northern end of an embayment that was largely in Mexico. Sediments in the Arizona part of this embayment range in thickness from more than 200 feet (pl. 6-A) in the southeast corner of the State (Epis, 1956) to 30 feet near the Chiricahua Mountains to the north (Armstrong, 1962).

Although the remnants of interval D are too small for thickness trends to be determinable in northern Arizona, the isolated occurence of 6 feet of interval D limestone in eastern Grand Canyon suggests that a sheet of limestone once extended into this area from the west during interval D time.

LITHOFACIES TRENDS

In southern Arizona a major change in the character of sedimentation and environment of deposition is represented by interval D. Rocks of this interval include sandstone and mudstone as well as carbonate types (pl. 6-B) in cyclic sequences, according to Armstrong (1962, p. 24). Marine and terrigenous units interfinger. These sequences of sandstone, mudstone, and carbonate rock contrast with the pure limestone-dolomite sequence of the underlying earlier Mississippian and have a greater similarity to the Pennsylvanian strata above.

In northwestern Arizona the single known remnant of interval D is limestone, but it is too small to justify generalizations concerning lithology of the larger sheet that was originally deposited.

SOURCES AND ENVIRONMENTS

Little information is available on the depositional environment of interval D in southeastern Arizona. Because both marine and continental deposits are present and because the marine invasions are known to have come from the southeast (pl. 11, fig. 4), uplift and erosion must have taken place in areas located in other directions. This uplift and erosion resulted in accumulation of terrigenous sediment following a long time of exclusively carbonate deposition. However, no data on paleocurrent directions are as yet available to show the location of any such area or areas.

PALEOTECTONIC IMPLICATIONS

Cyclic and fluctuating sedimentation in interval D suggests crustal instability in the southeastern part of Arizona and adjoining New Mexico (pl. 10, fig. 4). Emergence and uplift in various, as yet unidentified, areas doubtless account for an influx of terrigenous materials into the southern seaway during Chester time; delineation of rather well-defined positive elements recognized throughout Pennsylvanian time probably also began during Chester time.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

Mississippian rocks have their maximum thicknesses in Arizona in opposite corners of the State (pl. 7). In the northwest corner, a thickness in excess of 800 feet occurs in the vicinity of Lake Mead; in the southeast corner, the thickness is more than 1,200 feet.

Zero isopachs delimiting the two principal depositional areas outline positive elements in the northeast and southwest. Near the northern part of the east border of the State is the Defiance-Zuni positive element onto which the Mississippian strata lapped, and from which a structural ridge, partly submarine, extended far westward across the State. Another positive element probably was present in the northwestern part of Mexico as suggested by probably southward thinning of Mississippian rocks in the southwestern part of Arizona.

Mississippian rocks are incompletely represented in the area extending from the southern margin of the Colorado Plateau diagonally southeastward across

Arizona. Because of extensive pre-Pennsylvanian erosion, the total thickness southwest of the Defiance-Zuni positive element and near central Arizona is a considerably reduced record of original sedimentation.

PALEOTECTONIC IMPLICATIONS

The isopach map of the Mississippian System (pl. 7) shows clearly that both the Cordilleran miogeosyncline to the northwest and the Sonoran geosyncline to the southeast extended into Arizona and were actively subsiding during Mississippian time. Considerable amounts of sediment accumulated during at least three major transgressions and regressions of the sea. Positive elements, although emergent, could not have been elevated appreciably, as indicated by an almost complete lack of terrigenous detritus in most sediments that are preserved. Only in interval D in southeastern Arizona are appreciable amounts of terrigenous sediment preserved in the Mississippian section.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Almost everywhere in Arizona that Mississippian strata are present, rocks of Early Pennsylvanian age are present above them (pl. 8). In a few small areas later erosion has completely removed Pennsylvanian rocks; in some of these places Tertiary gravels or lavas rest on the Mississippian.

Pennsylvanian strata directly above the Mississippian are of Morrow age in the northwest (basal Supai Formation) and in the southeast (Black Prince Limestone; basal Horquilla Limestone, in extreme southeast). The Black Prince Limestone was originally considered to be Late Mississippian or Early Pennsylvanian(?) in age (Gilluly and others, 1954), but later faunal studies (Nations, 1963) show that it is Early Pennsylvanian (Morrow) in age. The oldest Pennsylvanian beds in northeastern Arizona (Molas Formation and basal part of lower Hermosa Formation) and in east-central Arizona (lowest part of Naco Formation) are believed to be of Atoka age.

PALEOTECTONIC IMPLICATIONS

Although scattered traces of young Mississippian (Chester) strata are known from various parts of Arizona and Lower Pennsylvanian strata overlie the Mississippian in most places, a widespread hiatus is present between Mississippian and Pennsylvanian rocks. Evidence of subaerial erosion during the time represented by the hiatus consists of a conspicuous karst surface on which were developed abundant sinkholes, caverns, and deposits of solution breccia over wide areas.

Mississippian and Pennsylvanian strata occupy similar areas in Arizona and apparently were deposited in the same major basins and geosynclines; likewise the positive elements that defined the margins of these periodically sinking areas seem to have been in much the same places during both systems. The hiatus between systems represents a time when the region was elevated above sea level and subjected to long-continued erosion, largely by solution, but it was a time of major change in the tectonic patterns of the region.

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Central Rocky Mountains and Northern Colorado Plateau Region

By WILLIAM W. MALLORY

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–M

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

CENTRAL ROCKY MOUNTAINS AND NORTHERN COLORADO PLATEAU REGION

By WILLIAM W. MALLORY

ABSTRACT

Westward retreat of the sea from the craton at the close of Devonian time was followed in Kinderhook (interval A) and Osage (interval B) time by a new advance of the sea. An east-thinning wedge of carbonate rock was deposited which overlapped rocks of Precambrian to Late Devonian age. This widespread deposit bears the names Madison Limestone (locally Madison Group) in Wyoming, Pahasapa Limestone in the Black Hills area, Guernsey Formation (upper part) in the Hartville area, Leadville Limestone in western Colorado, Deseret Limestone and Madison Limestone equivalent in the Uinta Mountains area, and Mooney Falls, Thunder Springs, and Whitmore Wash Members of the Redwall Limestone in southern Utah.

The Meramec (interval C) was a time of emergence; rocks of this interval are limited to extreme northern and western Wyoming, to the Uinta Mountains area in northern Utah, and to south-central Utah. In Wyoming, limestone of the interval C part of the Madison locally contains evaporites. In the Uinta Mountains area the Humbug Formation contains sandstone, shale, and carbonate rock, possibly derived from the weathering of Precambrian rocks in southeastern Wyoming. In south-central Utah the Redwall is dominantly limestone. The sea retreated completely from Wyoming in Meramec time, and a land surface with as much as several hundreds of feet of relief was developed on the Madison. Deposition in eastern Utah was continuous from Meramec (interval C) to Chester (interval D) time. In latter Chester time (interval D) the sea advanced into western Wyoming, invaded valleys developed on the Madison, and submerged all but the highest hills. A we-l-sorted sandstone, the Darwin Sandstone Member of the Amsden Formation, was deposited in this shallow sea with an irregular distribution. The sand was probably derived from a distant northeast source. Interval D in the Uinta Mountains area is composed of sandstone, shale, and limestone in the Doughnut Formation and has limited extent. The sea continued to expand in Wyoming from Late Mississippian into Early Pennsylvanian time, and Lower Pennsylvanian rocks cover most of the region except western Colorado and eastern Utah.

REGION DEFINED

The Central Rocky Mountains and northern Colorado Plateau region includes the following areas (fig. 47):

Wyoming: Entire State except that part west of the 110° meridian.

South Dakota: Fall River, Custer, and western Pennington Counties (southern part of the Black Hills area).

Colorado: All of the State west of the Front Range, Wet Mountains, and Sangre de Cristo Range.

Utah: The Uinta Mountains, the Uinta basin, and the rest of the southern part of the State except the western tier of counties.

PALEOGEOLOGY

Rocks beneath the Mississippian System range in age from Late Devonian to Precambrian (pl. 2).

In northwestern Wyoming rocks of the Mississippian System lie disconformably on the Cottonwood Canyon Member of the Madison Limestone of Devonian and Mississippian age. In central Wyoming they lie on the Devonian Darby Formation, the Ordovician Bighorn Dolomite, Cambrian sandstone correlative with the Deadwood Formation of the Black Hills area, and, locally, the Cambrian Gallatin Limestone (pl. 2). In southeastern Wyoming the upper part of the Guernsey Formation, the Madison Limestone and an underlying unnamed sandstone are assigned to the Osage; the age of the upper part of the Guernsey is not definitely known (pl. 15, col. 94), and it may include beds of Kinderhook age. The lower part of the Guernsey is shown as Devonian by Love, Henbest, and Denson (1953). In northeastern Wyoming and the Black Hills area, rocks of Mississippian interval A rest on closely associated dark shale of the Englewood Formation identified as Early Mississippian and Late Devonian (Klapper and Furnish, 1962).

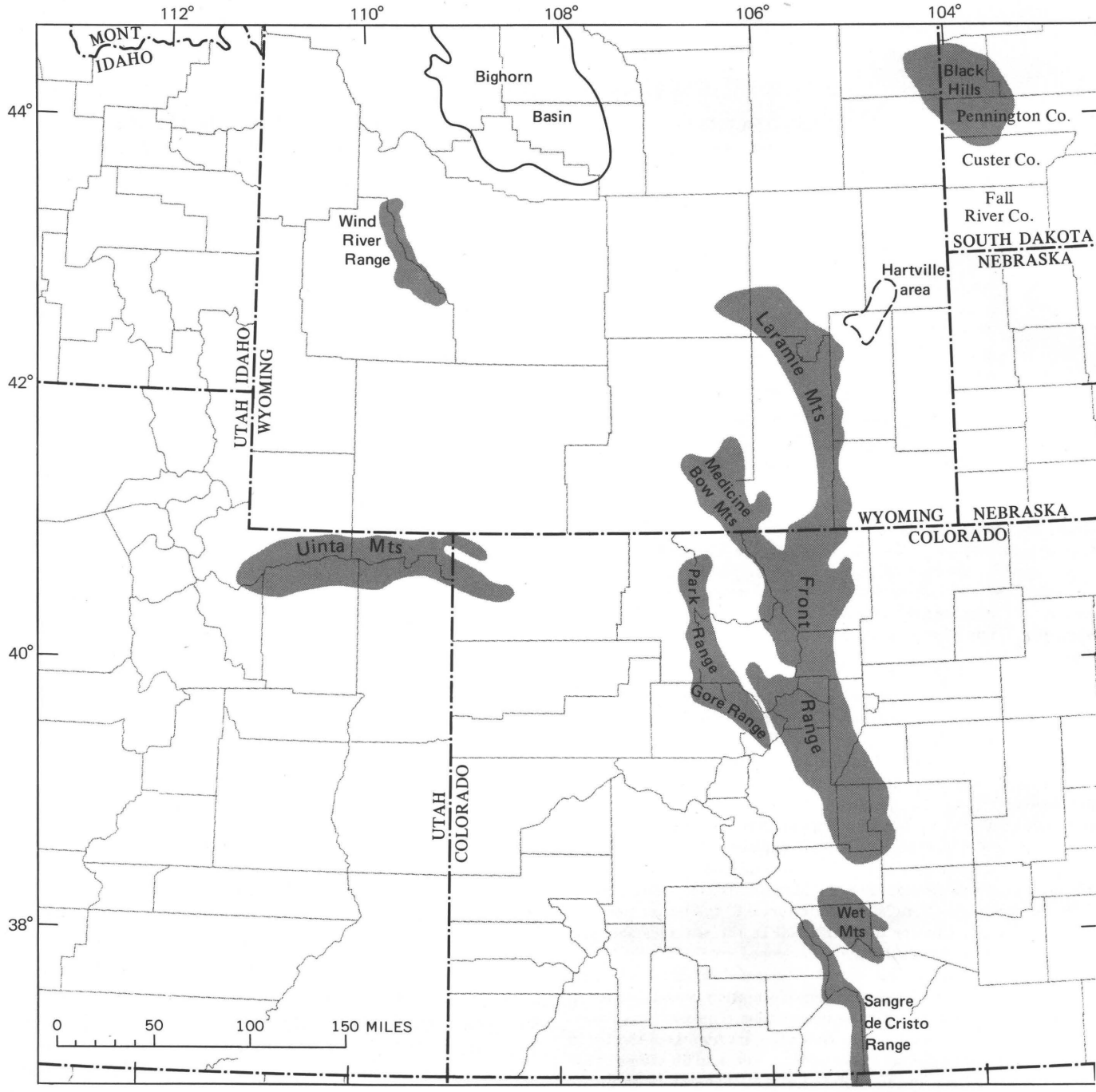

FIGURE 47. — Geographic features mentioned in text.

In northwestern Colorado and the Uinta Mountains area, the Mississippian System rests on the Lodore Formation and Tintic Quartzite of Cambrian age. In central Colorado Mississippian rocks rest on the Dyer Dolomite Member of the Chaffee Formation (Upper Devonian). In southwestern Colorado the Mississippian Leadville Limestone rests on the Ouray Limestone. Baars and Knight (1957, p. 121) stated that the Ouray fauna has both Devonian and Mississippian aspects, s the systemic boundary lies within the Ouray.

The contact of the Mississippian on older rocks i everywhere a disconformity, except possibly in south western Colorado where the contact is within the Oura Limestone and in northwestern Wyoming where th systemic boundary is within the Cottonwood Canyo Member of the Madison Limestone.

INTERVAL A

FORMATIONS INCLUDED

Identification of rocks of Kinderhook age, which make up interval A in most of Wyoming, is based on fieldwork in 1965 and 1966 by W. J. Sando (written commun., 1966), who found macrofossils of Kinderhook age at the base of the Madison Limestone, usually in a dolomite unit less than 100 feet thick. Where diagnostic fossils are not available, as in subsurface sections, dolomite at the base of the Madison has been arbitrarily assigned to interval A.

In the Black Hills area the lower part of the Pahasapa Limestone is included in interval A.

In the Hartville uplift area of southeastern Wyoming (pl. 15, col. 94), the Guernsey Formation may contain beds of Kinderhook age, but definitive evidence is lacking. None of the Guernsey is shown on the maps of interval A (pls. 3-A, 3-B).

In the Paradox basin (fig. 48), Baars (1966, p. 2093–2101) recognized within the Leadville Limestone of Mississippian age a regionally persistent intraformational disconformity, and he informally designated the part below the disconformity as the lower member of the Leadville and the part above as the upper member. The lower member was considered by Baars as the age equivalent of the Whitmore Wash and Thunder Springs Members of the Redwall Limestone in northern Arizona. He stated that primitive endothyrids closely resembling *Chernyshinella* occur in the lower member and that no more advanced forms were found. On this evidence he considered that the lower member of the Leadville (and the upper part of the gradationally underlying Ouray Limestone) are of Kinderhook age (Baars, 1966, p. 2089, 2101).

McKee(1960) considered that only the lower part of the Whitmore Wash Member is of Kinderhook age (pl. 15, col. 116). L. D. Holcomb (oral commun., 1968) called attention to the difficulty of placing a contact between the Leadville Limestone and the Ouray Limestone and pointed out that a major fraction of the strata which contain the primitive endothyrid fauna may belong in the Ouray. It is reasonably certain that rocks of Kinderhook age are present in the Paradox basin, but their thickness and extent are controversial, and all the Leadville arbitrarily has been placed in interval B.

THICKNESS AND LITHOFACIES TRENDS

In northern and western Wyoming rocks of interval A compose a broad wedge which thins east and south from 200 feet at the Idaho State line to zero in the central, southwestern, and northeastern parts of the State (pl. 3-A). Everywhere in Wyoming interval A is carbonate rock (pl. 3-B).

ENVIRONMENTS OF DEPOSITION, PALEOTECTONIC IMPLICATIONS, AND PALEOGEOGRAPHIC INTERPRETATIONS

The relation of the preceding topics to the continental inundation which began in Kinderhook time and culminated in Osage time will be most conveniently discussed in the section on interval B.

INTERVAL B

FORMATIONS INCLUDED

Interval B rocks are the most widespread part of the Mississippian System in the region. In most of Wyoming it consists of the middle part of the Madison Limestone (pl. 15, col. 95). In southeastern Wyoming it makes up the Guernsey Formation (part), and in the Black Hills it makes up all but the basal few feet of the Pahasapa. In northeastern Utah and extreme northwestern Colorado, the Deseret and Madison Limestones compose interval B. In southeastern Utah the Mooney Falls, Thunder Springs, and Whitmore Wash Members of the Redwall Limestone are assigned to the interval. Throughout most of western Colorado the Leadville Limestone is assigned to interval B. At the base of the Leadville is the Gilman Sandstone Member, a zone of yellow or light gray sandstone or dolomitic sandstone 5–50 feet thick but usually 10–20 feet thick. The Gilman rests with inconspicuous disconformity on the Dyer Dolomite Member of the Chaffee Formation. The Gilman has been recognized with certainty only in central Colorado. Limestone, locally cherty in the upper part and having a basal feldspathic, conglomeratic sandstone, extends as a fingerlike projection from southwestern Wyoming into eastern Colorado (Maughan, 1963, p. C23–C26). Fossils at the base of the limestone section (notably *Chonopectus fischeri*) establish a Mississippian, probably Kinderhook age for the sequence (Maughan, 1963, C23–C26). The sequence, because it is thin and its age assignment is somewhat uncertain, is included with interval B.

The ages of the upper and lower parts of the Guernsey have not been determined with certainty, but fossils from the middle part suggest an Early Mississippian age (Love and others, 1953). The Mississippian part of the formation has been assigned to interval B, although beds of Kinderhook age may be present.

UPPER BOUNDARY OF INTERVAL B

Throughout most of Wyoming and nearly all of western Colorado the upper boundary of interval B is

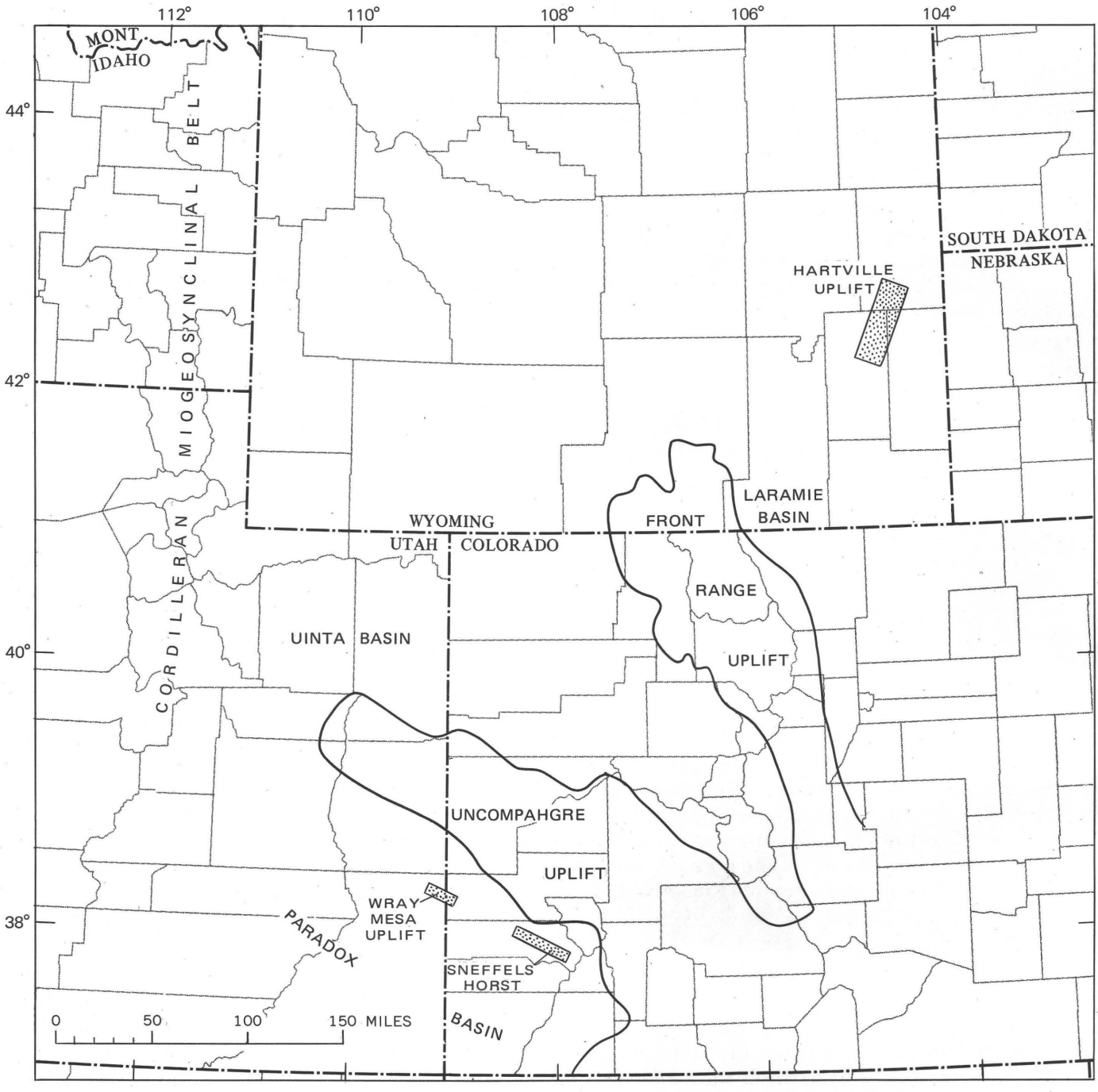

FIGURE 48. — Some late Paleozoic tectonic features mentioned in text. The Front Range and Uncompahgre uplifts are elements of the Ancestral Rocky Mountains of Pennsylvanian time.

conspicuously disconformable. In many places solution and erosion have carved an irregular surface commonly having tens or scores of feet of relief. In central and western Wyoming the boundary is further emphasized by the lithologic contrast of white- or salmon-colored sandstone of the overlying Darwin Sandstone Member of the Amsden Formation of Chester age with limestone of the Madison. In eastern and southern Wyoming the irregular upper surface of interval B is the systemic boundary, and rocks of Pennsylvanian age, usually some shade of red, contrast markedly with the gray Mississippian limestone.

In nearly all of western Colorado the upper surface of the Leadville Limestone is the systemic boundary, and it exhibits the same topographic irregularity present on the Madison in Wyoming. Overlying Pennsylvanian rocks are dark-gray shale or red beds.

In northwestern Wyoming and in central and northeastern Utah where rocks of interval C are present, the upper boundary of interval B is less conspicuous. In Wyoming the base of interval C is placed at the base of a widespread solution zone in which brecciated limestone blocks are mixed in a sandy or shaly matrix. The top of the Mooney Falls Member of the Redwall Limestone, which is the top of interval B (McKee, 1960, 1963), can be traced in wells northward from outcrops in the Grand Canyon, Ariz., into central Utah. Diagnostic criteria in Utah for separating limestone of the Mooney Falls from limestone in the overlying Horseshoe Mesa Member of the Redwall are not abundant, but careful matching of subsurface logs provides correlations which can be accepted with a fair degree of confidence.

In northeastern Utah carbonate rocks of interval B contrast strongly with the mixed clastic lithology of the overlying Humbug Formation; hence, the upper boundary of interval B, though conformable, is clearly discernible.

THICKNESS TRENDS

Interval B forms a wedge thickening westward across western Colorado and eastern Utah toward the Cordilleran geosyncline farther west (pl. 4-A; pl. 9-F, sec. *N–N'*). Rate of thickening increases fairly constantly westward, and interval B attains a maximum thickness of 1,400 feet or more in central Utah. Isopachs terminate abruptly at the Uncompahgre uplift in central Colorado. Gradual thinning seems to take place at the west margin of the Front Range uplift in north-central Colorado (fig. 48). Isolated areas near the southern margin of the Uncompahgre uplift where interval B rocks are missing are at the sites of the Wray Mesa uplift and the Sneffels horst (Baars, 1966), uplifted blocks active in pre-Late Cambrian and in post-Mississippian time.

In Wyoming interval B exhibits considerable irregularity of thickness. From isolated maximums of more than 600 feet in northwestern, northeastern, and southwestern Wyoming, the interval thins southeasterly to a highly irregular zero limit in the southeast corner of the State. An arcuate area of relatively thin interval B rocks, 100–300 feet thick, extends through central Wyoming. In the southeastern corner of Wyoming the interval is less than 100 feet thick.

LITHOFACIES TRENDS AND SEDIMENT SOURCE

Nearly everywhere in the region interval B consists mostly of carbonate rock, but it contains minor quantities of sandstone or shale (pl. 4-B). In western Colorado, the basal Gilman Sandstone Member of the Leadville Limestone is a persistent unit but so thin that it does not affect the lithofacies patterns of interval B. In central-southeastern Wyoming, a thin basal sandstone composes a large enough fraction of the interval to affect the map patterns. This sandstone unit was assigned a Cambrian age for many years, but Maughan (1963) stated that it intertongues with the Guernsey Formation, and Agatston (1954, p. 514) reported Mississippian fossils. The distribution of the sandstone indicates that an area in central southeastern Wyoming and north-central Colorado was emergent late in Kinderhook or early in Osage time and was high enough to supply small quantities of sand to the surrounding sea.

That part of the Madison Limestone assigned to interval B consists almost entirely of thin- to medium-bedded, dark-colored limestone (Strickland, 1956). Its lower part is nearly everywhere argillaceous, commonly cherty and thin bedded. Traces of anhydrite are present in the northern part of the Bighorn basin. In north-central and eastern Wyoming, the dark, thin- to medium-bedded, argillaceous limestone facies is replaced by a medium- to thick-bedded light-colored bioclastic limestone and dolomite facies that is difficult to distinguish from the overlying Mission Canyon.

The lower part of the Guernsey Formation in the Hartville area is thin-bedded fine-grained silty purple to gray dolomite interbedded with fissile dolomitic purple shale and siltstone. Sparse lenses and nodules of pink chert are present. Pink arkose containing abundant pink feldspar and white quartz grains and pebbles occurs at the base. The upper part of the Guernsey is gray cherty coarsely crystalline thick bedded dolomite that grades upward into blue-gray cherty limestone. Dark-brown quartzitic cherty concretions are common. Oolitic and cavernous beds are present on outcrops and in subsurface sections.

The Pahasapa is a massive gray limestone similar to the Madison. The Deseret is blue-gray thin-bedded cherty dolomitic limestone, not easily distinguished from the underlying Madison equivalent. The Deseret contains beds of both Osage and Meramec age (pl. 15, col. 110) but is shown entirely on maps as interval B (pls. 4-A, 4-B).

The Whitmore Wash Member of the Redwall Limestone as seen in cuttings from deep wells in southeast Utah is white, gray, or tan oolitic dolomite

and limestone. Algal remains are common. The Thunder Springs Member is cream-colored to brown cherty oolitic limestone and dolomite. Crinoidals are common. The Mooney Falls Member is tan to gray coarsely oolitic slightly argillaceous limestone and dolomite. Oolites are especially common in the upper part.

The Leadville Limestone in the type locality and in many other parts of the Colorado mountains (Knight and Baars, 1957; Mayer, 1964; Rothrock, 1960; Bloom, 1961) has been extensively dolomitized and mineralized. Where not altered, it is commonly a blue-gray to gray-brown massive medium to coarsely crystalline limestone. Black chert is locally present (Merrill and Winar, 1958, p. 2114).

Rocks in Wyoming included in interval B have been described by Strickland (1956); in Colorado, by Knight and Baars (1957), Mayer (1964), Rothrock (1960), and Bloom (1961); in Utah, by Reid (1954) and Heylmun (1961); in the Four Corners area, by Knight and Baars (1957), Parker (1961), and Parker and Roberts (1963, 1966); and in the Paradox basin, by Baars (1966).

ENVIRONMENT OF DEPOSITION

A sea advanced eastward over the Central Rocky Mountains region, beginning in Kinderhook (interval A) time and culminating in nearly complete inundation of the region in Osage (interval B) time (pl. 11, figs. 1, 2). During much of interval B the entire area of deposition was a shallow to moderately deep sea. The water was unusually clear and free from mud or sand.

PALEOTECTONIC IMPLICATIONS

The region during intervals A and B was structurally stable except for a slow, steady foundering of the crust which allowed the sea to invade almost the entire region. In north-central Colorado and southeastern Wyoming, a lowland remained emergent (pl. 10, fig. 2).

Subsidence during interval B was greatest in a belt trending northeast across central Utah, an area that was emergent in interval A time. Rocks assigned to interval B in this trough have a maximum thickness of about 1,400 feet. The orientation of the trough and the unusual thickness of rocks deposited in it, compared with the thickness of rocks in the rest of the region, suggest that the trough was either a segment of the Cordilleran miogeosyncline or a parallel, marginal cratonic trough transitional in tectonic activity with the larger miogeosyncline. Because thickness of interval B rocks in western Utah ranges generally from 500 to 1,000 feet, the trough in central Utah can be considered a part of the miogeosynclinal belt for this part of Mississippian time.

PALEOGEOGRAPHIC INTERPRETATIONS

The Early Mississippian sea invaded an almost level plain in Wyoming, Colorado, and eastern Utah which had a gentle slope northwestward (pl. 12, figs. 1, 2). Only southeastern Wyoming and north-central Colorado remained above water. The central part of this area probably was high enough to be an area of low hills.

INTERVAL C

FORMATIONS INCLUDED

Interval C, which includes rocks mostly of Meramec age (pl. 15) is present in northwestern Wyoming as the Bull Ridge Member of the Madison Limestone (pl. 15, col. 95; Sando, 1968, p. 1855–1856), in the Uinta Mountains area as the Humbug Formation (pl. 15, col. 110; Kinney, 1955), and in south-central Utah as the Horseshoe Mesa Member of the Redwall Limestone (McKee, 1963).

UPPER BOUNDARY OF INTERVAL C

The upper boundary of interval C is determined in different areas by different criteria. In southern Utah the upper limit of the carbonate rock of the Horseshoe Mesa at the top of interval C contrasts with the overlying red beds of Pennsylvanian age. In the Uinta Mountains area a high sandstone content of the Humbug Formation distinguishes it from less sandy limestone in the overlying Doughnut Formation of interval D. In northwestern and western Wyoming limestone of the Bull Ridge Member of the Madison contains brecciated solution zones and solution fillings of sandstone, shale, and traces of anhydrite. This unit contrasts sharply with the overlying Darwin Sandstone Member of the Amsden Formation of interval D where the Darwin lies locally on the Madison or with red beds of Pennsylvanian age which lie on the Madison elsewhere.

THICKNESS TRENDS

Interval C rocks (pl. 5-A) in north-central Utah fill an irregular trough and attain a maximum thickness of about 1,000 feet in central Utah. In south-central Utah thicknesses are irregular, but they average about 100 feet. The margins of interval C in eastern Utah and northwestern Colorado are highly irregular owing to extensive solution and erosion. Several isolated outliers of interval C, probably erosion remnants, have been identified in this area. The eroded edges of carbonate and evaporite rocks that are widespread in Idaho and Montana extend into northern and central-western Wyoming, where they are less than 200 feet thick.

LITHOFACIES TRENDS

The lower part of the Bull Ridge Member of the Madison consists of reddish or yellowish thin-bedded silty dolomite, siltstone, and shale that commonly contains angular fragments of limestone derived by solution from overlying strata. Irregular bodies of red sandstone were deposited in solution cavities after post-Madison, pre-Amsden exposure are also common. The upper part is limestone and dolomite which was partly or completely brecciated by collapse, probably due to solution of and removal of anhydrite. Stromatolite structures are evident in the upper part of the member at many localities (Sando, 1968, p. 1856).

The Humbug Formation differs in composition from place to place, but, in general, it is composed of interbedded carbonate rock, sandstone, and locally of thin mudstone beds. The Horseshoe Mesa Member of the Redwall as seen in cuttings from deep wells in southeast Utah is cream-colored to tan limestone and dolomite. Algal remains and crinoidals are common.

The Horseshoe Mesa Member of the Redwall Limestone is homogeneous carbonate, but it grades northward in the Uinta Mountains area into the lithologically diverse Humbug Formation (pl. 5-B). In north-central Utah the lithofacies is muddy carbonate, but these rocks grade west, north, and east into calcareous sandstone. In northern and western Wyoming the highly irregular erosional edge of the interval makes restoration of facies trends difficult. Considerable muddy carbonate is present in northwestern Wyoming. Traces of anhydrite suggest that breccia zones in the Bull Ridge Member of the Madison were once occupied by extensive anhydrite and gypsum deposits which have since been removed by solution.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The lithofacies patterns in the Humbug Formation suggest that muddy carbonate formed in the central part of a small landlocked bay (pl. 11, fig. 3). Sandstone at the margins of the bay probably was derived from a land area in southeastern Wyoming. Local beds of gypsum and anhydrite attest to poor circulation of the sea in northwestern Wyoming.

PALEOTECTONIC IMPLICATIONS

The highly irregular edges of the Horseshoe Mesa Member of the Redwall in eastern Utah and the Bull Ridge Member of the Madison Limestone in northern and western Wyoming indicate that the original extent of these rocks was greater than their present extent (pl. 10, fig. 3). The same reasoning probably applies to the Humbug Formation, although control is scarce in southwestern Wyoming for determining the present extent of equivalents of this unit. Epeirogenic uplift caused the sea to regress during interval C time. At the close of the interval the sea had withdrawn westward into the Cordilleran miogeosyncline, and the region remained emergent long enough to cause weathering and erosion of a highly irregular surface on the top of the Madison and Leadville Limestones throughout most of Wyoming and Colorado.

PALEOGEOGRAPHIC INTERPRETATIONS

Most of the eastern part of this region (pl. 12, fig. 3) was a low plain, which was subaqueous early in the interval and subaerial at the end. A central part in the vicinity of the Laramie basin may have been high enough to be classified as low hills and to have been a source of detritus for the Humbug Formation.

INTERVAL D

FORMATIONS INCLUDED AND UPPER BOUNDARY OF INTERVAL D

Two formations or parts of formations compose interval D, the Darwin Sandstone Member of the Amsden Formation in western Wyoming and the Doughnut Formation in the Uinta Mountains area (pl. 15, cols. 95, 110). The Darwin is the basal member of the Amsden Formation and lies with marked disconformity on the Madison Limestone. At an obscure exposure in the Wind River Range, on Cherry Creek, macrofossils were collected from two ledges in the lower part of the overlying Horseshoe Shale Member of the Amsden Formation (Shaw and Bell, 1955). These have been identified as Late Mississippian (Chester), or transitional Mississippian-Pennsyvlanian. Other collections in western Wyoming support this age assignment for the Horseshoe (Sando, 1967, p. D32–D34); hence, the underlying Darwin is considered of Chester age.

The upper boundary of the interval is marked by lithologic contrast with overlying rocks but is everywhere conformable or apparently conformable. Well-sorted, white-, pink-, or salmon-colored sandstone in the Darwin Sandstone Member of the Amsden Formation contrasts with bright-red shale and siltstone of the Horseshoe Shale Member of the Amsden. Sandstone and shale of the Doughnut contrasts with an overlying sequence of limestone and red beds in the Round Valley Limestone or the Morgan Formation which contain Pennsylvanian Foraminifera.

THICKNESS TRENDS

Isopachs for interval D define a northeast-trending

trough in northeast Utah (pl. 6-A). The Doughnut Formation, which fills the trough, thickens southwest into the Cordilleran miogeosyncline and attains a thickness of more than 1,000 feet in central Utah. The Darwin is variable in thickness and irregular in extent.

LITHOFACIES TRENDS

The Darwin Sandstone Member at most places is gray, white, or cream-to-salmon sandstone; locally, it is brick red or has brick-red blotches or specks. It is crossbedded and well sorted and is composed almost entirely of fine to medium quartz sand. The cement is silica or, locally, calcite. The Darwin is remarkably homogeneous in composition throughout its extent (pl. 6-B). Color variations have no discernible pattern.

The Doughnut Formation in the Uinta Mountains area is typically black shale, but it also contains red, green, brown, and purple mudstone and beds of limestone and dolomite. Locally, carbonate rock is the principal constituent. The facies pattern is clearly related to the shape of the trough in which the Doughnut Formation was deposited. The central part is muddy carbonate, whereas to the northwest and southeast, on the margins of the trough, carbonate rock grades into mudstone; in extreme northwestern Colorado, the mudstone grades into muddy sandstone.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The sea advanced westward in Wyoming during interval D time over a surface of appreciable relief. Sand brought in by a river system flowing west or southwest was dropped in quiet water in estuaries and bays which earlier had been river valleys cut on the Madison Limestone at the end of interval C time (pl. 11, fig. 4). The sand, as suggested by its high degree of sorting and its homogeneity, probably was derived from a preexisting sandstone outside the Rocky Mountains region. A possible source was Ordovician sandstone exposed in the Hudson Bay–Lake Superior region (Mallory, 1967, p. G25–G27, fig. 16).

Terrigenous material in the Doughnut Formation may have been derived locally, as suggested by the fact that grain size coarsens toward the margins of the deposit. Possibly, the detrital material was derived from the area of low hills identified during Osage time in south-central Wyoming and north-central Colorado. This area appears to have been exposed throughout Mississippian time, and during interval D time it could have provided detrital material that was transported by rivers across the intervening area.

PALEOTECTONIC IMPLICATIONS

Subsidence was renewed and became more intense in the northeast-trending trough that originated in Meramec time in northeastern Utah (pl. 10, fig. 4; pl. 9, sec. *B–B'*). The Darwin Sandstone Member, which forms a thin blanket on the karst surface of the Madison in Wyoming, indicates relative tectonic stability for that area. Tectonism in Wyoming consisted of slow downwarping, sufficient to allow eastward advance of a shallow sea. If the marine transgression was due to eustatic rise, then the Wyoming shelf may have been tectonically neutral.

PALEOGEOGRAPHIC INTERPRETATIONS

During latter Chester time an invading sea penetrated into western and central Wyoming from the west, initially following drainage channels eroded in the Madison Limestone during late Meramec and early Chester time. The shoreline must have been highly irregular since its outline was controlled by the old dendritic drainageways. Estuaries and bays formed initially, but later they coalesced into an open seaway that covered all of western Wyoming (pl. 12, fig. 4*C*).

The area now occupied by ranges of the Rocky Mountains in southern Wyoming and northern Colorado was probably a land area of low hills in Chester time.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

The bulk of the Mississippian rocks in the region is assignable to interval B, and the thickness map of all Mississippian rocks (pl. 7) resembles the interval B isopach map (pl. 4-A). In western Colorado and eastern Utah the system is represented by a simple westward-thickening wedge, except for the area of the Uncompahgre uplift and nearby smaller areas where Paleozoic rocks were removed by erosion in Pennsylvanian and later time. Near the northwestern tip of the Uncompahgre uplift, the rate of westward thickening increases markedly into the Cordilleran miogeosyncline. In Wyoming the system thickens generally northwestward from a zero limit in the south Central Rocky Mountains region to about 1,000 feet in the northwestern part of the State. A northeast-trending band of thin strata, however, is present in the central and southwest part of the State. An adjacent band of thick strata is conspicuous in the Uinta Mountains area reflecting the presence of a northeasterly trending trough that developed in the latter half of Mississippian time.

GEOLOGIC UNITS DIRECTLY ABOVE THE MISSISSIPPIAN SYSTEM

Mississippian rocks nearly everywhere in Wyoming, northwestern Colorado, and northeastern Utah are overlain (pl. 8) by rocks of Morrow (Pennsylvanian) age. In southeastern Utah and southwestern Colorado, Pennsylvanian rocks younger than Morrow rest on the Mississippian System in many places, including the margins of the Ancestral Rocky Mountains where Mississippian strata are locally overlain by rocks of Atoka, Des Moines, Missouri, and Virgil age, and in southeastern Utah where in the central part of the Paradox basin rocks of Atoka age directly overlie the Mississippian System. On the northeastern and southwestern margins of the Paradox basin rocks of Atoka age are absent and rocks of Des Moines age overlie the Mississippian. In south-central Utah Pennsylvanian strata are absent and Permian rocks lie on Mississippian.

SUMMARY

Mississippian rocks record two marine advances and an intervening retreat. In Kinderhook and Osage time, the sea overlapped the region from the west, covering rocks of Precambrian to Late Devonian age and depositing a wedge of carbonate rock which thickens generally north and northwest. In Meramec time the sea retreated. Strata of Meramec age are limited to western and northern Wyoming and to central and northeastern Utah and environs. Most of these strata are carbonate rocks, but sandstone and shale are present in the Uinta Mountains region.

A second marine invasion, which began in Chester time in western Wyoming, culminated in complete inundation of previously weathered and eroded Mississippian strata in most of Wyoming and the northeastern Utah–northwestern Colorado area by the end of Early Pennsylvanian (Morrow) time. A broad land area in southwestern Colorado and southeastern Utah remained emergent during Chester and Morrow time; Mississippian limestone in this area was deeply weathered to form a solution breccia of limestone nodules and residual red clay, the Molas Formation. On the eastern and western margins of the Paradox basin, weathering of Mississippian rocks continued into Middle Pennsylvanian (Des Moines) time.

During Mississippian time the region was tectonically stable except for gentle epeirogenic warping possibly combined with eustatic changes in sea level.

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Northern Rocky Mountains and Adjacent Plains Region

By ALBERT E. ROBERTS

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–N

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

NORTHERN ROCKY MOUNTAINS AND ADJACENT PLAINS REGION

By ALBERT E. ROBERTS

ABSTRACT

At the close of the Devonian Period and the beginning of the Mississippian Period a thin layer of black marine shale was deposited over much of the Northern Rocky Mountains and Adjacent Plains region; then widespread Kinderhook seas covered the entire region and deposition was predominantly limestone. In the Osage seas, deposition produced cyclic alternating layers of dolomite and limestone, and during Meramec time dolomite, anhydrite, and salt were laid down in restricted seas. The Williston basin and Central Montana trough deepened and broadened during the Mississippian Period. Magnesian content of carbonates deposited in these areas gradually increased until late Meramec time. Regional tectonic movements in late Meramec and early Chester time are believed to have initiated a change from deposition of carbonates and evaporites to deposition of terrigenous sediments characteristic of the Chester.

Fossils in the basal Lodgepole Limestone indicate that during the Kinderhook the entire area was inundated in a very short period of time by a northeastward-transgressing sea. Continuity of the carbonate and interbedded shale sequence suggests that the region was tectonically stable. Deposition in the eastern part of the region was on a broad shelf of low relief in an epeiric sea. The western part of the region was a miogeosyncline in which a thick and lithologically uniform sequence of very fine grained carbonaceous clastics was deposited suggesting a rather constant downwarping and filling. Sediments were derived from a landmass of low relief to the northwest or southwest.

Tectonic and sedimentary differentiation of the shelf from the adjacent miogeosyncline to the west and the Williston basin to the east began in Osage time. On the shelf area normal marine conditions predominated everywhere during Osage time and extensive carbonate sequences were deposited that were considerably thicker than those deposited during Kinderhook. These carbonate sediments are relatively uniform in lithology and thickness, and tectonic and sedimentary changes are recognized only on a regional scale.

The Northern Rocky Mountains and Adjacent Plains region was within the equatorial belt probably between 5° N. and 10° N. paleolatitude in which the climate probably was warm during Meramec time. Deposition for most of the region consists of carbonate and carbonate-evaporite sequences. These characteristic lithofacies are considered to have formed on a nearly level sea floor during cyclic rising and lowering of the sea. Types of fossils and the presence of little or no terrigenous clastic material within the carbonate and carbonate-evaporite sequences suggest a marine environment distant from the shoreline; the remote land areas probably had very low relief.

A major change in the tectonic activity of the Mississippian occurred at the close of the Meramec or the beginning of the Chester as indicated by a change in the kind and extent of deposits and by the facies patterns of the earliest units of the Chester. Areas emerged that had previously been submerged. Detrital sediments of the Big Snowy Group or its stratigraphic equivalents of Chester age were probably derived through erosion of Mississippian and older Paleozoic rocks from these uplifted areas. The upper beds of the pre-Chester Mission Canyon Limestone or Charles Formation were truncated, and, locally, a karst topography was developed. Similarly, in north-central Utah at the west end of the Uinta Mountains, a red clastic unit, interpreted as a karst deposit, commonly occurs at the base of the Doughnut Formation (equivalent to part of the Manning Canyon Shale and part of the Great Blue Limestone).

Regional diastrophism and uplift ended deposition of Mississippian rocks. Much of south-central Montana was gently elevated, and Mississippian rocks that had been recently deposited were partly eroded. Widespread karst deposits in much of Montana were formed at this time. The Mississippian-Pennsylvanian boundary is established in Montana, Idaho, and westernmost Wyoming at a regional unconformity between the Big Snowy Group of Chester age and the unconformably overlying Amsden Group of Pennsylvanian age.

REGION DEFINED

The region of this chapter, the Northern Rocky Mountains and Adjacent Plains region, includes Idaho, western and central Montana, north-central Utah, and western Wyoming (fig. 49). During deposition of the Mississippian System this region was predominantly a marine shelf with the subsiding intracratonic Williston basin on the east side and the subsiding Cordilleran miogeosyncline on the west (fig. 50).

In central and eastern Idaho, western Montana, and western Wyoming, Mississippian rocks crop out (pl. 1) on the flanks of individual ranges that compose the Northern Rocky Mountains. Mississippian rocks are also exposed in Montana in ranges of the plains area east of the Rocky Mountains. These include the Little Belt, Big Snowy, Little Rocky, and Pryor Mountains, and the Beartooth Range. Rocks of Mississippian age also form prominent outcrops in the Crawford Mountains and the Wasatch and Bear River Ranges of north-central Utah. Throughout much of the Plains area Mississippian rocks are present in the subsurface and are known from extensive drilling for petroleum.

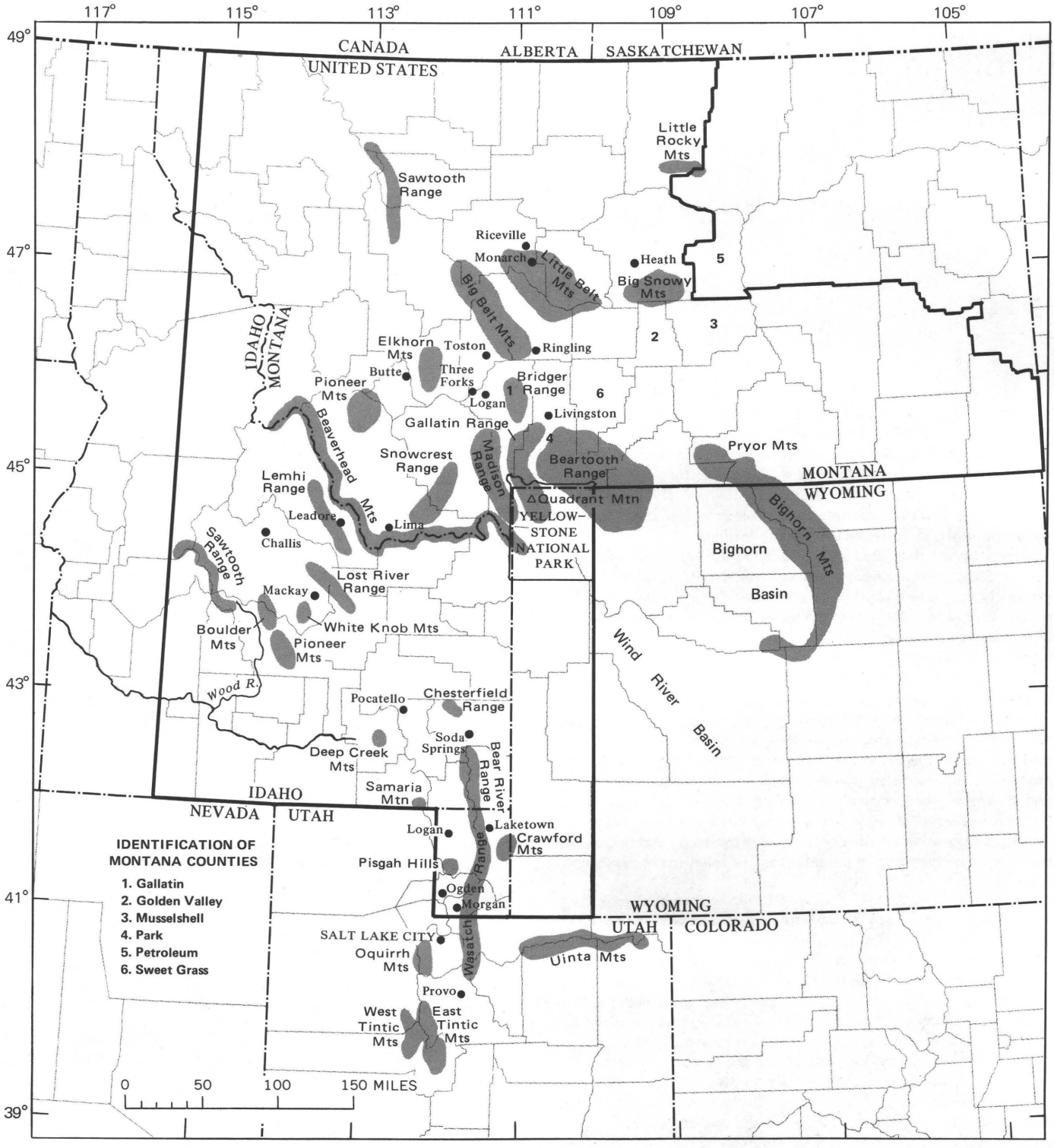

FIGURE 49. — Geographic features in Northern Rocky Mountains and Adjacent Plains region mentioned in text. Solid line indicates region of this chapter.

NOMENCLATURE

The rock-stratigraphic units (group, formation, member, bed), time-stratigraphic units (series), and interval assignments (A, B, C, D) of the Mississippian System for this region are summarized on plate 15 (cols. 96, 101–109). The selected reference sections are modified from the following authors: (96) Sando and Dutro, 1960; (101) Maughan and Roberts, 1967; (102) Mudge, Sando, and Dutro, 1962; (103) Roberts, 1961,

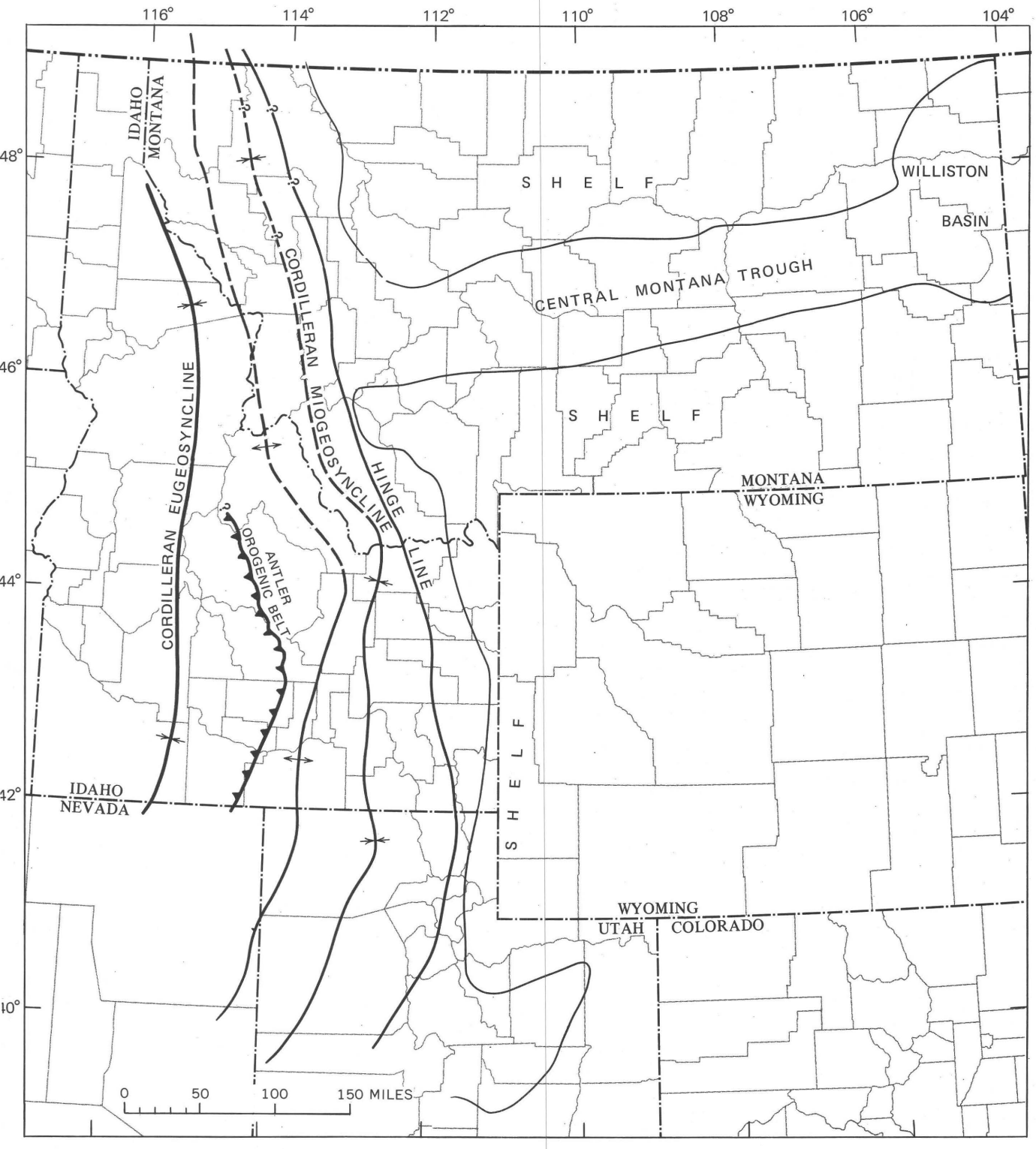

FIGURE 50. — Some structural elements in Northern Rocky Mountains and Adjacent Plains region mentioned in text.

.966; (104) Huh, 1967; Mamet, Skipp, Sando, and Mapel, 1971; Sandberg, 1975; (105) Carr and Trimble, .961; (106) Dutro and Sando, 1963; (107) Williams, .948; Mullens and Izett, 1964; (108) Sando, Dutro, and Gere, 1959; and (109) Mullens and Laraway, 1967.

For convenience in the following discussion I classify individual Mississippian sections as being in the miogeosynclincal area or the shelf area. This classification is difficult in southeastern Idaho and north-central Utah because this area was the site of a north-trending

tectonic hinge belt or line (fig. 50) which gradually migrated eastward during the Mississippian Period. Thus, during interval A the area was part of the shelf, but by the time of interval D it had become part of the miogeosyncline. Moreover, the area continued to be tectonically active in post-Mississippian time; thus, the stratigraphic record has been reduced or destroyed locally.

AGE AND CORRELATION

SHELF AREA

Throughout the Northern Rocky Mountains and Adjacent Plains region the most commonly used nomenclature for the cratonic platform or shelf deposits includes the term "Madison" in some form or is correlated with the Madison. The Madison Limestone in southwestern Montana, as defined by Peale (1893, p. 33), was assigned by Walcott (in Peale, 1893, p. 39) to Lower Carboniferous. Faunas collected from the Madison Limestone in Yellowstone National Park were regarded by Girty (1899, p. 484) as possibly equivalent to the major part of the Mississippian but as having its strongest affinities with the Kinderhook. The Madison Limestone in the Little Belt Mountains was subdivided by Weed (1899a, p. 2) into the Paine Shale for the lower part, the Woodhurst Limestone for the middle part, and the Castle Limestone for the upper part. Faunas from the Paine Shale and the Woodhurst Limestone were assigned to the Kinderhook or Chouteau age according to Schuchert and Girty (in Weed, 1900, p. 293). The Madison Formation in the Little Rocky Mountain region in northern Montana was designated a group and divided into the Lodgepole Limestone and the overlying Mission Canyon Limestone by Collier and Cathcart (1922, p. 173).

The nomenclature of the Madison in southwestern Montana was integrated by Sloss and Hamblin (1942, p. 314). They proposed retention of group rank for the Madison with Collier and Cathcart's formational subdivisions of Lodgepole and Mission Canyon and the adoption of Weed's Paine Shale and Woodhurst Limestone as members of the Lodgepole Limestone. The stratigraphic section of Madison rocks exposed along the north bank of the Gallatin River near Logan, Mont., was intepreted by Sloss and Hamblin (1942, p. 313) as the type Madison section.

The Madison of southwestern Montana was summarized by Weller and others (1948, p. 138) as of late Kinderhook and Osage age. Laudon (1948, p. 295) and Holland (1952, p. 1714) concluded that faunas from the type Madison were indicative of Kinderhook age. Dolomite beds in the upper part of the Mission Canyon Limestone in southwestern Montana were assigned to the Meramec by Sloss and Moritz (1951, p. 2157) who tentatively correlated them with dolomite beds of the Charles Formation of central Montana. The Lodgepole and Mission Canyon Limestones in the Bridger Range represented continuous deposition through all Kinderhook and perhaps earliest Osage time according to Laudon and Severson (1953, p. 507). Williams (in Klepper and others, 1957, p. 18) recognized beds of Meramec age in the upper part of the Mission Canyon in western Montana. The Madison Group at the type section was subdivided into faunal zones (Sando and Dutro, 1960, p. 122; Sando, 1960, p. B227); the zones were correlated with late Kinderhook, Osage, and probable Meramec of the type Mississippian.

Mississippian rocks equivalent to the Madison Group (Sando, 1960, p. B226) exposed in the Sawtooth Range of northwestern Montana were named the Hannan Limestone by Deiss (1941, p. 1896) and were later described briefly by him (Deiss, 1943, p. 228). He (1933, p. 45) considered this sequence to be almost equivalent in time to the Kinderhook and Osage Series. Sloss and Hamblin (1942, p. 311) concurred with Deiss but recognized that some of the upper limestone beds might be of Late Mississippian age. Sloss and Laird (1945) subdivided the Hannan Limestone in northwestern Montana into units MC, MB_2, MB_1, and MA, in ascending order. Unit MC was assigned a Kinderhook age and was considered equivalent to the Paine Shale Member of the Lodgepole Limestone. Unit MB_2 was assigned an Osage age and was considered equivalent to the Woodhurst Limestone Member of the Lodgepole. Unit MB_1 was assigned an Osage age and was considered equivalent to the Mission Canyon Limestone. Unit MA was assigned a Meramec age and was considered equivalent to the Charles Formation. Mudge, Sando, and Dutro (1962, p. 2003) abandoned the Hannan, and the rocks are assigned to the Madison Group in part of this area and divisible into the Allan Mountain Limestone (equivalent to the Lodgepole Limestone) and the Castle Reef Dolomite (equivalent to the Mission Canyon Limestone). Unit MA of Sloss and Laird (1945) was named the Sun River Member of the Castle Reef Dolomite by Mudge, Sando, and Dutro (1962, p. 2003).

The Madison Group in southwestern Montana is divided into the Lodgepole Limestone, of Kinderhook and early Osage age, and the Mission Canyon Limestone, a lower member of late Osage age and an upper member of Meramec age (pl. 15, col. 103). Fossils found in the Madison Group in this area include corals, brachiopods, crinoids, bryozoans, and conodonts. On the basis of faunal collections, the Kinderhook-Osage boundary is placed at or near the contact between the Woodhurst Limestone and Paine Shale Members of the Lodgepole (Sando and Dutro, 1974), and the Osage-Meramec boundary may be placed at the base or the top

of the solution breccia that lies at the base of the upper member of the Mission Canyon Limestone. Inasmuch as the base of the solution breccia is a more consistent stratigraphic horizon than the top, the Osage-Meramec boundary is placed for convenience at the base. The upper member of the Mission Canyon Limestone in western Montana correlates with the Charles Formation in central and eastern Montana and unit MA of Sloss and Laird (1945) or with the Sun River Member of the Castle Reef Dolomite in northwestern Montana.

The Big Snowy Group, established by Scott (1935), consists of the Kibbey and Otter Formations, previously named by Weed (1900) for units in the Little Belt Mountains, and the Heath Formation which Scott recognized and named as a unit overlying the Otter in the Big Snowy Mountains. The Kibbey Formation is generally considered of Chester age inasmuch as in most places it rests conformably on beds of Meramec age. Locally, it rests unconformably on a karst surface. However, in some places the basal Kibbey interfingers with the uppermost part of the Charles Formation suggesting a local late Meramec age (Maughan and Roberts, 1967). A Chester age is also indicated by fossils from lower strata in the Otter Formation, which are laterally equivalent to the upper part of the Kibbey, and by the gradational contact with the overlying Otter Formation. Faunal collections by Weed (1900, p. 295–296) were assigned to the Carboniferous, apparently Lower Carboniferous. Paleontologic studies by Scott (1942) and Easton (1962) on the fauna of the Otter Formation indicate a marine environment of Chester age. Scott (1935, p. 1031–1032) considered the fauna of the Heath Formation closely related to that of the *Brazer Limestone* of Idaho and the Moorefield Formation of Arkansas and was an age no younger than late Chester. Easton (1962, p. 14), on the basis of abundant fossil collections, also assigned the Heath Formation to the Late Mississippian (Chester). Maughan and Roberts (1967) restricted the Heath to beds of Late Mississippian (Chester) age beneath a Late Mississippian to Early Pennsylvanian regional unconformity.

MIOGEOSYNCLINAL AREA

Deposition in the miogeosyncline in east-central Idaho during the Kinderhook and Osage Series formed a thick flysch deposit that was named the McGowan Creek Formation by Sandberg (1975). This Early Mississippian argillite sequence grades eastward into the shelf units of the Lodgepole Limestone and lower part of the Mission Canyon Limestone.

The first important change in the carbonate deposition in the miogeosynclinal-shelf area of southern Idaho and north-central Utah occurred in the middle of the Osage Series with deposition of a black phosphatic shale unit (fig. 52). This shale is the basal part of a carbonate sequence named the Deseret Limestone by Gilluly (1932). Morris and Lovering (1961, p. 93) used the name Deseret in the Tintic area, central Utah, and Mullens and Laraway (1967) extended the Deseret to the Morgan area in north-central Utah. Equivalent strata in south-central Idaho were named the lower member of the Deep Creek Formation by Carr and Trimble (1961, p. C181).

Throughout central Utah there is a thick quartzitic sandstone sequence with interbedded limestone and dolomite that conformably overlies the Deseret Limestone. Exposures of this stratigraphic unit in the East Tintic Mountains were named the "Humbug intercalated series" by Tower and Smith (1899, p. 625–626) and later redefined as the Humbug Formation by Gilluly (1932, p. 26–29). Crittenden (1959, p. 67) observed that the sandstones are not restricted to the Humbug and that similar sandstone bodies occur in abundance in both the upper part of the Deseret Limestone and in the middle third of the Great Blue Limestone. Mullens and Laraway (1967) extended the Humbug Formation to north-central Utah, near Morgan, and subdivided the Humbug into two members — a lower member of sandstone and an upper member of limestone.

The Humbug-Deseret sequence in the Chesterfield Range of southeastern Idaho was named the Little Flat Formation by Dutro and Sando (1963). The Little Flat is correlative to the west with the Deep Creek Formation of Carr and Trimble (1961). The Great Blue Limestone in the Chesterfield Range was named the Monroe Canyon Limestone by Dutro and Sando (1963).

The Humbug Formation in Big Cottonwood Canyon in the Wasatch Range near Salt Lake City is unconformably overlain by a sequence of black shale and interbedded dark-gray limestone named the Doughnut Formation by Crittenden, Sharp, and Calkins (1952, p. 10). In northeastern Utah east of the Wasatch Range, equivalent strata of the Great Blue Limestone are primarily black shale and interbedded carbonate and fine-grained clastics. Marine faunas from the Doughnut Formation are similar to those from the Great Blue and Manning Canyon (Crittenden, 1959, p. 73). Therefore, it seems likely that the Doughnut Formation is the eastern or shoreward stratigraphic equivalent of the Great Blue Limestone and part of the Manning Canyon Shale.

The name of Great Blue Limestone was applied informally to a Carboniferous limestone sequence exposed in the Oquirrh Mountains of central Utah by Spurr (1895). This sequence was later redefined by Gilluly (1932, p. 29–32), and the shales in the upper

part of the sequence were separated from it as the Manning Canyon Shale (1932, p. 31). Baker, Huddle, and Kinney (1949) identified the thick nearly homogeneous series of thin-bedded limestones that conformably overlies the Humbug Formation in the Wasatch Range of north-central Utah with the Great Blue Limestone and the overlying shale and interbedded limestone section as the Manning Canyon Shale. Terminology of the Great Blue Limestone and Manning Canyon Shale was extended to the Pocatello, Idaho, area in south-central Idaho by Carr and Trimble (1961) and to the Samaria Mountain area in southeastern Idaho by Beus (1968).

The Great Blue Limestone of central Utah is correlative with (1) the Monroe Canyon Limestone, in the Chesterfield Range of southeastern Idaho (Dutro and Sando, 1963, p. 1967), (2) units 2, 3, and 4 of the *Brazer Formation* in the Pisgah Hills of north-central Utah (Williams and Yolten, 1945, p. 1145), (3) members B and C of the *Brazer Limestone* in the Bear River Range of north-central Utah (Mullens and Izett, 1964, p. S10), and (4) part of the Doughnut Formation and the upper member of the Humbug Formation in the Morgan area of north-central Utah (Crittenden, 1959, p. 69).

The Manning Canyon Shale conformably overlies the Great Blue Limestone from north-central Utah to south-central Idaho. Locally, near the margin of deposition, at the east end of the Uinta Mountains, Colorado, equivalent rocks lie unconformably on the Humbug Formation. Girty (in Gilluly, 1932, p. 33–34) considered the Manning Canyon Shale to be Upper Mississippian and Lower Pennsylvanian in age. Later paleontologists concur in showing the age to be latest Chester and early Morrow. The systematic boundary divides the formation approximately into an upper third that is Pennsylvanian and lower two-thirds that is Mississippian. In east-central Idaho, rocks of the upper part of the Scott Peak, and the entire South Creek, and Surrett Canyon Formations are of Chester age (Mamet and others, 1971), and they lithologically grade eastward onto the shelf into units of the Big Snowy Group.

The Manning Canyon Shale of north-central Utah is correlative with (1) unit 5 of the *Brazer Formation* in the Pisgah Hills of north-central Utah (Williams and Yolten, 1945, p. 1145), (2) the lower 1,160 feet of the Manning Canyon Shale in the Deep Creek Mountains of south-central Idaho (Carr and Trimble, 1961, p. C182), and (3) the upper part of the Doughnut Formation in the Morgan area of north-central Utah (Crittenden, 1959, p. 69).

PALEOGEOLOGY

LOWER BOUNDARY OF THE MISSISSIPPIAN

In much of the Northern Rocky Mountains and Adjacent Plains region the Devonian-Mississippian boundary lies within a thin unit, generally a dark carbonaceous shale; the contact cannot be easily identified but superficially it appears to be conformable. Detailed stratigraphic and paleontologic studies, however, indicate that this thin lithogenetic unit contains a faunal break or hiatus at the systemic boundary (Sandberg and Klapper, 1967). Because of the difficulty of recognizing the true systemic boundary, the bottom of rocks treated as Mississippian for the purposes of this study has been placed arbitrarily at a conspicuous, sharp lithologic change at the base of the Lodgepole Limestone, the Allan Mountain Limestone, and the McGowan Creek Formation (pl. 15, cols. 96, 101–109). This contact is disconformable in much of the region of east-central Idaho, north-central Utah, and northwestern and central Montana. The contact appears to be conformable in southwestern Montana and western Wyoming where the Cottonwood Canyon Member of the Lodgepole or of the Madison Limestone is present or in north-central and northeastern Montana where the black shale unit of the Bakken Formation is present.

UNITS UNDERLYING THE MISSISSIPPIAN

Rocks that underlie the Mississippian as arbitrarily limited in this study are mostly of earliest Mississippian and latest Devonian age (pl. 2). These are the Bakken Formation in north-central and northeastern Montana, the Exshaw Formation in northwestern Montana, the Sappington Member of the Three Forks Formation in southwestern Montana, the Cottonwood Canyon Member of the Lodgepole Limestone in south-central Montana and adjacent eastern Idaho, the Milligen Formation in central Idaho (Wood River area), the Cottonwood Canyon Member of the Madison Limestone in western Wyoming, and the Leatham Formation in north-central Utah.

In an area extending from northwestern Montana through central Montana into southeastern Montana these thin units of Late Devonian–Early Mississippian age are missing and Mississippian rocks rest on older beds of Late Devonian age — the Three Forks, the Duperow, and the Birdbear Formations. In a small area in central Montana, Devonian and Silurian rocks are missing and Mississippian rocks rest directly on the Red River Formation of Ordovician age and on undifferentiated Cambrian and Ordovician rocks.

PALEOTECTONIC IMPLICATIONS

In Late Devonian time epeiric movements in Montana warped and locally uplifted Upper Devonian beds and exposed them to erosion. These cratonic movements may have been simultaneous with part of the Antler orogeny in the geosynclinal area to the west. Thin uppermost Devonian and lowermost Mississippian

marine rocks then accumulated in downwarped areas, shallow basins that were separated by the broad areas of uplift. The strandline appears to have migrated across wide areas as a result of eustatic changes or slight epeiric movements. Finally, a relatively widespread subsidence of the region marked the beginning of interval A and allowed accumulation of the Lodgepole Limestone and its equivalents.

INTERVAL A

FORMATIONS INCLUDED

SHELF AREA

In the type section of the Madison Group near Three Forks in southwestern Montana, interval A consists of the Paine Shale Member of the Lodgepole Limestone and its stratigraphic equivalents (pl. 15, col. 103). In shelf areas of western Wyoming, north-central Utah, and southeastern Idaho, where the Paine Shale Member has lost its identity, equivalent strata assigned to interval A are generally referred to as the lower member of the Lodgepole Limestone. In northwestern Montana, interval A consists of the lower and middle members of the Allan Mountain Limestone (Mudge and others, 1962).

The type Lodgepole Limestone in the Little Rocky Mountains of north-central Montana consists of thin-bedded carbonate and shale. The lower part of the Lodgepole becomes increasingly silty from this area northward into Canada. In central and southwestern Montana, where the Lodgepole can be subdivided, the Paine Shale Member consists of greenish-gray siltstone intercalated with thin-bedded finely to coarsely crystalline limestone, dolomitic and magnesian limestone, and calcitic dolomite.

In southwestern Montana a dolomitic limestone unit 20 feet thick at the base of the Lodgepole contains about 30 percent brown and yellow chert and is an excellent stratigraphic marker, particularly in structurally complex areas where the Three Forks Formation is faulted or squeezed out (Roberts, 1966, p. B7). Dark-gray to yellow nodular chert also distinguishes the basal Lodgepole in the Bridger Range (Laudon and Severson, 1953, p. 509). Richards (1957, p. 409) described these basal cherty beds, which stand in ledges above the easily eroded Three Forks Shale, in the Beartooth Range southeast of Livingston. Chert is also common in the lower part of the Lodgepole in western Wyoming, north-central Utah, and in the subsurface of central Montana. In eastern Montana this part of the Lodgepole sequence becomes glauconitic.

In north-central Utah, along most of Blacksmith Fork Canyon, in the Bear River Range, the Lodgepole Limestone is conformable with the underlying Leatham Formation, but near the mountain front the Lodgepole rests unconformably on the Beirdneau Sandstone Member of the Jefferson Formation or isolated remnants of the Leatham Formation preserved along the unconformity (Mullens and Izett, 1964).

In the Williston basin the Bakken Formation comprises an upper black shale, a middle fine-grained clastic unit, and a lower black shale. In the Three Forks, Mont., area the Sappington Member of the Three Forks Formation has the same lithologic sequence and has the identicial conodont faunas as the Bakken. The upper part of the Sappington includes the Devonian-Mississippian boundary within it. Because of the dominantly Devonian age, the entire threefold Sappington and its stratigraphic equivalents are excluded from interval A in this report.

MIOGEOSYNCLINAL AREA

In the Lost River Range of central Idaho interval A consists of the lower part of the McGowan Creek Formation (pl. 15, col. 104). In the Pioneer Mountains, Idaho, interval A is an argillite sequence in the basal Copper Basin Group (Paull and others, 1972). Near the Idaho-Montana border in the Beaverhead Mountains gradational and intertonguing relations between the McGowan Creek and Lodgepole have been described by Sandberg, Mapel, and Huddle (1967).

STRATIGRAPHIC RELATIONS

The Paine Shale Member of the Lodgepole Limestone and its stratigraphic equivalents on most of the shelf area of the Northern Rocky Mountains and Adjacent Plains region rests conformably on Devonian formations. The Paine Shale is recognizable only in central and southwestern Montana on the basis of lithology; however, equivalent strata (lower member) of the widespread Lodgepole Limestone is generally fossiliferous, and coral and brachiopod zonation (Sando and Dutro, 1960) was generally used to establish the boundary between intervals A and B. In north-central Utah the Lodgepole Limestone is conformable with the underlying Leatham Formation along most of Blacksmith Fork Canyon in the Bear River Range (Mullens and Izett, 1964, p. S7). Near the mountain front, however, the Lodgepole rests unconformably on the Beirdneau Sandstone Member of the Jefferson Formation or isolated remnants of the Leatham Formation preserved along the unconformity (Mullens and Izett, 1964).

In the miogeosyncline area dark-gray shale and argillite of the lower part of the McGowan Creek Formation were often included with the Devonian Milligen Formation on the basis of lithology (Sandberg and others, 1975). Sandberg, Mapel, and Huddle (1967, p.

C128) described an assemblage of Early Mississippian (Kinderhook) conodonts from the base of the McGowan Creek in the Lost River Range of central Idaho, indicating an exclusively Mississippian age for the McGowan Creek. This assemblage is similar to the conodont fauna from the base of the Lodgepole in Montana.

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A in Montana, western Wyoming, north-central Utah, and southeastern Idaho is in a continuous carbonate sequence. The boundary was established throughout the area in key stratigraphic sections on the basis of fossils, using the coral and brachiopod zonation established by Sando and Dutro (1960), then extended from these sections on lithologic characteristics. In general, the uppermost beds of the Paine Shale Member or its stratigraphic equivalents are thin- to medium-bedded argillaceous dolomite or dolomitic limestone that is slightly less resistant to erosion than the overlying rocks. The basal beds of interval B are generally massive- to medium-bedded limestones having little or no argillaceous content, and they are more resistant to weathering. The basal limestone unit of interval B is oolitic and is laterally persistent throughout much of Montana.

THICKNESS TRENDS

Strata of interval A are thin relative to the other intervals of the Mississippian System in the Northern Rocky Mountains and Adjacent Plains region; much of the thinness is thought to be due to onlap during transgression of a shallow Lodgepole sea. Throughout the shelf area of Montana, western Wyoming, north-central Utah, and southeastern Idaho, the thickness of this interval ranges from about 100 feet to slightly more than 300 feet (pl. 3-A). Lateral changes in thickness are very gradual. In general, the interval has thicknesses of 200 feet or more in a broad irregular east-trending belt that extends from central Montana into east-central Montana. Rocks of interval A thin from this belt southward toward Wyoming and northward toward Canada.

In the miogeosynclinal area of central Idaho, however, the thickness of this interval increases rapidly westward. A maximum thickness of 1,500 feet for interval A is identified in central Idaho. Although control is sparse, maximum thicknesses of 500–600 feet occur in apparently isolated areas of rapid subsidence.

LITHOFACIES TRENDS

Interval A throughout the shelf area of Montana, western Wyoming, southeastern Idaho, and northeastern Utah consists of thin-bedded gray or brownish-gray to dark-gray fine- to medium-grained limestone and dolomite alternating with thin beds of gray to greenish- or yellowish-gray calcareous shale or siltstone assigned to the Lodgepole Limestone (pl. 3-B). The carbonates typically weather light gray or yellowish gray. The dolomites are characteristically argillaceous. The insoluble residue of the dolomites, which is distinctive of all the Madison Group, generally increases as the quantity of magnesium increases (fig. 51). Throughout the carbonate sequence of interval A, chert is also common in thin layers along bedding planes or, less commonly, in nodules or lenses. Chert is particularly abundant in the basal part of interval A and is an excellent stratigraphic marker in central and south-central Montana, western Wyoming, and north-central Utah (Roberts, 1966, p. B7). In southern and southeastern Montana glauconite becomes a common constituent in the basal part of interval A. Some of the carbonate beds contain crinoids, bryozoa, corals, and brachiopods. The shaly partings commonly contain fragments of these fossils. In many places, the basal few feet of the Lodgepole consists of calcarenitic carbonate which represents deposits formed in shallow water during the widespread Kinderhook transgression.

In northern Montana the basal Lodgepole Limestone becomes increasingly silty northward. Thus, in southern Alberta and Saskatchewan, Canada, the base of the Mississippian carbonate equivalent to the Lodgepole is younger and the underlying Bakken Formation or Exshaw Formation thickens northward at the expense of the Lodgepole or its equivalent.

Interval A in the miogeosynclinal area of central Idaho consists of massive- to thin-bedded dark-gray to black shale, claystone, mudstone, and siltstone interbedded with some sandstone and limestone (pl. 3-B). In areas of complex structure the sequence is commonly altered to slate and phyllite, siliceous mudstone and chert, and quartzitic siltstone, quartzite, and limestone.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The textures of the carbonate sequence of interval A suggest a shallow marine low-energy environment (pl. 11, fig. 1). Detrital silt and clay in the sediments of interval A throughout the shelf area of Montana, western Wyoming, southeastern Idaho, and north-central Utah are probably reworked insoluble residues derived from the erosion of older Paleozoic sedimentary rocks exposed in adjacent low-lying land areas. A land area occupied much of southeastern Wyoming and was one of very low relief. The low content of detrital material in interval A marginal to this southern land area indicates very little contribution of detrital material from the south. On the other hand, an increase of detrital material to the north in northern Montana suggests

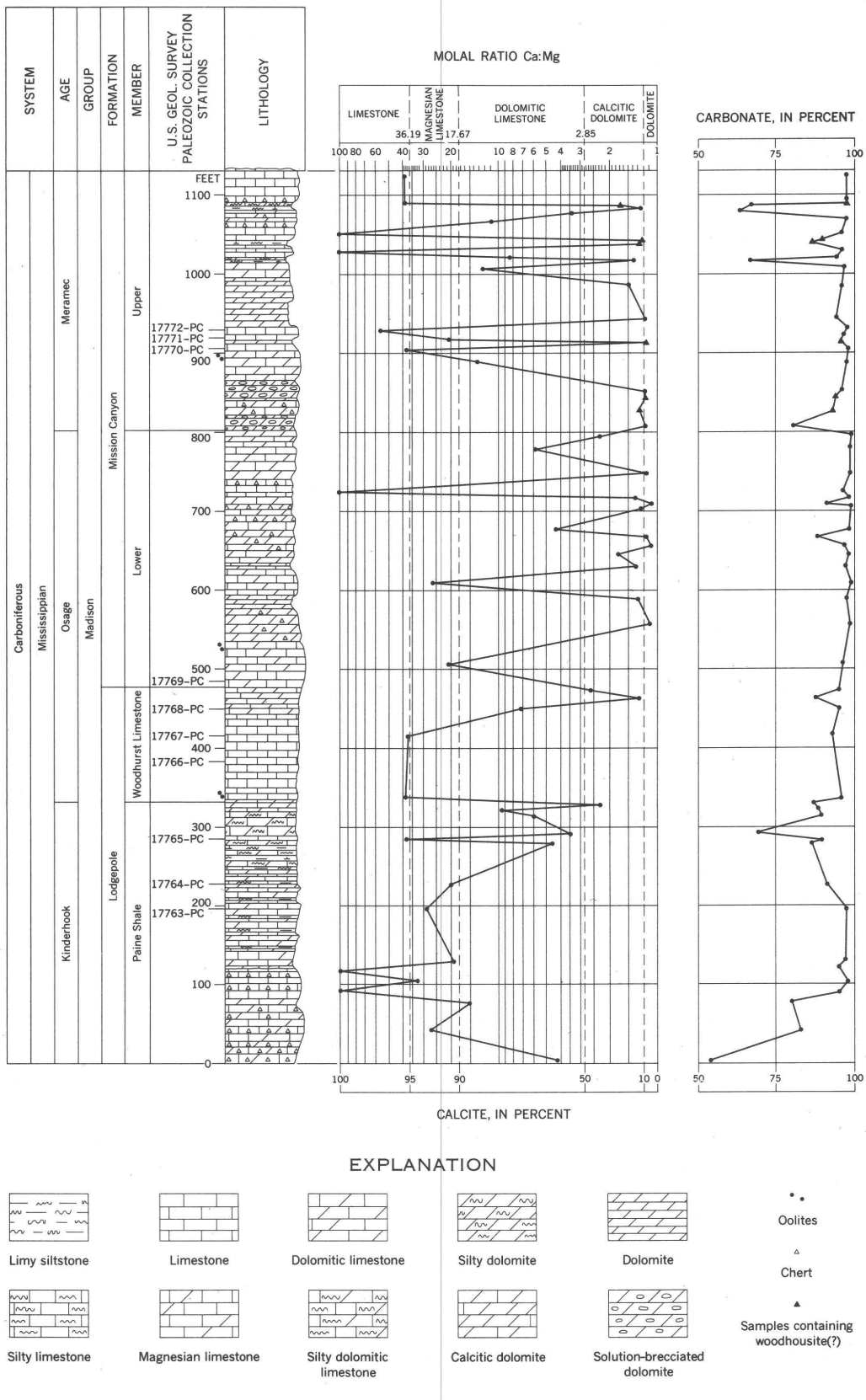

FIGURE 51. — Stratigraphic section of the Madison Group in sec. 35, T. 2 S., R. 9 E., and secs. 1 and 2, T. 3 S., R. 9 E., near Livingston, southwestern Montana. Calcium-magnesium analyses by J. A. Thomas. Reprinted from Roberts (1966, p. B5).

contributions of material from this direction. Any land area in Canada may have been some distance away.

In the miogeosynclinal area of central Idaho, thick very fine grained black carbonaceous sediments were deposited in euxinic and brackish water lagoons and other nearshore environments in rapidly sinking basins (pl. 3-B). The rate of subsidence probably was about equal to the rate of filling. The presence of land to the west is indicated by coal beds in this interval (Umpleby and others, 1930, p. 26) in the Wood River area of central Idaho (pl. 12, fig. 1). Postdepositional eastward thrusting has moved some of these deposits to the east; therefore, the shore was farther west.

PALEOTECTONIC IMPLICATIONS

At the close of the Devonian Period the surface of the Northern Rocky Mountains and Adjacent Plains region was at or near sea level. Fossils in the Paine Shale Member of the Lodgepole indicate that the entire region was inundated in a very short period of time as a result of either epeiric subsidence or eustatic rise of sea level. The regional continuity of the carbonate and interbedded shale sequence suggests that except for the regional downwarping the region was a tectonically stable area (pl. 10, fig. 1) and that a general southeastward thinning of the Paine Shale Member of the Lodgepole (interval A) is due to the southeastward onlap across the Wyoming landmass. This transgression covered all Late Devonian positive areas.

The thick and lithologically uniform sequence of very fine grained carbonaceous clastic sediments of interval A that were deposited in the miogeosynclinal area suggests constant downwarping (pl. 10. fig. 1) and filling.

Although the position of the axis of the eugeosynclinal belt west of the region is concealed by younger deposits, remnants of eugeosynclinal rocks of possible Mississippian age are locally identified in Oregon and Washington (Yates, chap. Q). It may be noteworthy that heavy minerals and chert grains in the coarse clastics of the eugeosynclinal belt in Nevada were derived from the west (Churkin, 1962, p. 587), and, by projection, a subsiding eugeosynclinal belt may have extended northward from Nevada, west of this region, during Mississippian time; sediments filling the belt might have been derived from the west.

INTERVAL B

FORMATIONS INCLUDED

SHELF AREA

The Woodhurst Limestone Member of the Lodgepole Limestone, the lower member of the Mission Canyon Limestone (pl. 15, col. 103), and their stratigraphic equivalents compose interval B on the shelf area of the Northern Rocky Mountains and Adjacent Plains region.

The equivalent of the Woodhurst Limestone Member in northwestern Montana is the upper member of the Allan Mountain Limestone (Mudge and others, 1962). In western Wyoming, north-central Utah, and southeastern Idaho the equivalent unit is the upper member of the Lodgepole Limestone (pl. 15, cols. 105–109).

The upper half of interval B for the shelf area includes the stratigraphic equivalents of the lower member of the Mission Canyon Limestone in the type locality of the Madison Group near Three Forks in southwestern Montana (pl. 15, col. 103). In central Montana, Mission Canyon Limestone is restricted to this interval. In northwestern Montana the equivalent unit is the lower member of the Castle Reef Dolomite (Mudge and others, 1962). Rocks in western Wyoming have the same nomenclature as the type locality of the Madison; however, to the west equivalent strata in the Chesterfield Range of southeastern Idaho are designated the siltstone member of the Little Flat Formation (Dutro and Sando, 1963, p. 1967). Near Pocatello, Idaho, strata equivalent to the lower member of the Mission Canyon Limestone are the lower member of the Deep Creek Formation (Carr and Trimble, 1961, p. C181). In north-central Utah equivalent rocks are lower members of the Brazer Dolomite, *Brazer Limestone*, or *Brazer Formation* (pl. 15, cols. 107–108). Also, near Morgan in north-central Utah, equivalent strata compose the lower part of the Deseret Limestone (Mullens and Laraway, 1967).

MIOGEOSYNCLINAL AREA

In the miogeosynclinal area near Mackay in central Idaho, interval B composes the upper half of the McGowan Creek Formation as used by Sandberg (1975).

STRATIGRAPHIC RELATIONS

The Woodhurst Limestone Member of the Lodgepole Limestone or its stratigraphic equivalent rests conformably on the Paine Shale Member or its stratigraphic equivalent. Assemblages of brachiopods and corals from Livingston, Mont., establish that the Kinderhook-Osage boundary lies at or near the contact between the Paine and Woodhurst (Sando and Dutro, in Roberts, 1966, p. B16–B17). More recent work has refined the age determination, and the Woodhurst is now regarded as containing some beds of latest Kinderhook age (Sando and others, 1969, p. E22). Throughout much of Montana the basal limestone unit of the Woodhurst is oolitic

and the Paine contains numerous beds or partings of shale, making the two stratigraphic units distinguishable. However, in southeastern Idaho, north-central Utah, and western Wyoming, where the typical physical characteristics of the Paine and Woodhurst Members are absent, the boundary between intervals A and B was established entirely on the basis of fossils.

The contact between the Woodhurst Limestone Member of the Lodgepole Limestone or its stratigraphic equivalent and the overlying lower member of the Mission Canyon Limestone or its equivalent is gradational. In general, the contact is placed arbitrarily to divide the more argillaceous and thinner bedded carbonates below from the less argillaceous and thicker bedded carbonates above. Fossils indicate that the Lodgepole–Mission Canyon contact is near the middle of the Osage (pl. 15, cols. 101, 103).

In north-central Utah the upper member of the Lodgepole Limestone is overlain by the Deseret Limestone or the Brazer Dolomite or their stratigraphic equivalents. Straddling the hinge line between the miogeosyncline and the shelf in this area a widespread black phosphatic shale forms the base of the Deseret or Brazer (fig. 52). Blackwelder (1910, p. 543) first described this phosphatic unit from outcrops east of Ogden, Utah. For a comprehensive historical and economic treatment of the phosphatic shale unit the reader is referred to Cheney (1957), Morris and Lovering (1961, p. 93–104), and Mullens and Izett (1964, p. S8, S21–S24). Gilluly (1932, p. 26), in defining the Deseret Limestone in central Utah, established the base of the formation at the base of the phosphatic unit. Williams (1939, p. 8–9, 26–30) described the basal member of the *Brazer Formation* in north-central Utah as a 200-foot sequence of black phosphatic shale, in which thin beds and stringers of oolitic phosphate rock occur. In Utah a phosphatic unit crops out in a number of places: near Laketown (Richardson, 1941, p. 22), near Logan (Williams, 1948, p. 1142), in the Wasatch Range east of Provo (Baker and others, 1949, p. 1174), in the Wasatch Range east of Salt Lake City (Calkins and Butler, 1943, p. 25), in the Crawford Mountains (Sando and others, 1959, p. 2747), and in the East Tintic Mountains (Morris and Lovering, 1961, p. 93–104). In Idaho a phosphatic unit is exposed in Monroe Canyon in the Chesterfield Range near Soda Springs (Mansfield, 1952, p. 20). These occurrences of phosphatic rock are regarded here as correlative.

Faunal evidence from the lower part of the Brazer Dolomite and the Deseret Limestone indicates an Early Mississippian age. Sando, Dutro, and Gere (1959, p. 2755–2758) studied a fauna of bryozoans, brachiopods, and corals from members 1 and 2 of the type Brazer Dolomite at Brazer Canyon in the Crawford Mountains, north-central Utah. They compared the fauna from member 1 with species characteristic of uppermost Kinderhook or lowermost Osage and the assemblage from member 2 with common Osage faunas of the standard series of the Mississippi Valley region.

Carbonate beds within or immediately above the Deseret-Brazer phosphatic shale unit in this area were sampled for conodonts to establish the continuity and age of the unit. A typical sample — collected by J. E. Smedley in 1968 at Old Laketown Canyon from a bed of medium-dark-gray dolomite within the phosphatic shale, 29–33 feet above the base of the Brazer Dolomite, in sec. 32, T. 13 N., R. 6 E., near Laketown, Utah — contained the following conodont fauna identified by C. A. Sandberg (written commun., 1969):

	Number of specimens
Polygnathus cf. *P. communis* Branson and Mehl	128
Spathognathodus pulcher (Branson and Mehl)	14
S. regularis (Branson and Mehl)	2
Doliognathus latus Branson and Mehl	3
D. dubius Branson and Mehl	1
Syprioniodina sp	9

According to C. A. Sandberg (written commun., 1969), *Doliognathus*, as at present known, is restricted to the lower and middle parts of the Osage — that is, Fern Glen Limestone to middle Burlington Limestone of the Mississippi Valley. It characterizes the *Bactrognathus–Polygnathus communis* Assemblage Zone of Collinson, Scott, and Rexroad (1962). In Missouri, it occurs in rocks equated to the upper part of that zone as well as in rocks equated to the lower part of their overlying *Bactrognathus–Taphrognathus* Assemblage Zone (Thompson, 1967). In both the Mississippi Valley and Missouri, *Doliognathus* occurs below and does not overlap the first occurrence of *Taphrognathus*. The abundant *Polygnathus* in this sample is apparently a new species that is also present in the upper part of the McGowan Creek Formation of central Idaho which is regarded as Osage (Sandberg, 1975, p. E9). In North America, *Polygnathus* does not occur above the Osage. In Europe, however, younger occurrences of *Polygnathus* are reported, but the occurrences are not abundant above rocks equivalent to Osage. Sandberg (written commun., 1969), on the basis of the preceding fauna and those from the phosphatic unit in adjacent areas, concluded that the base of the Deseret and Brazer in north-central Utah is most likely middle Osage. The ammonoids *Dzhaprakoceras* and *Beyrichoceras* are present in the beds sampled at Old Laketown Canyon. These ammonoids are common in the lower part of the Deseret Limestone and have been used previously as evidence of a Meramec age.

Primarily on the basis of paleontologic data, some geologists have proposed that the Deseret and the lower

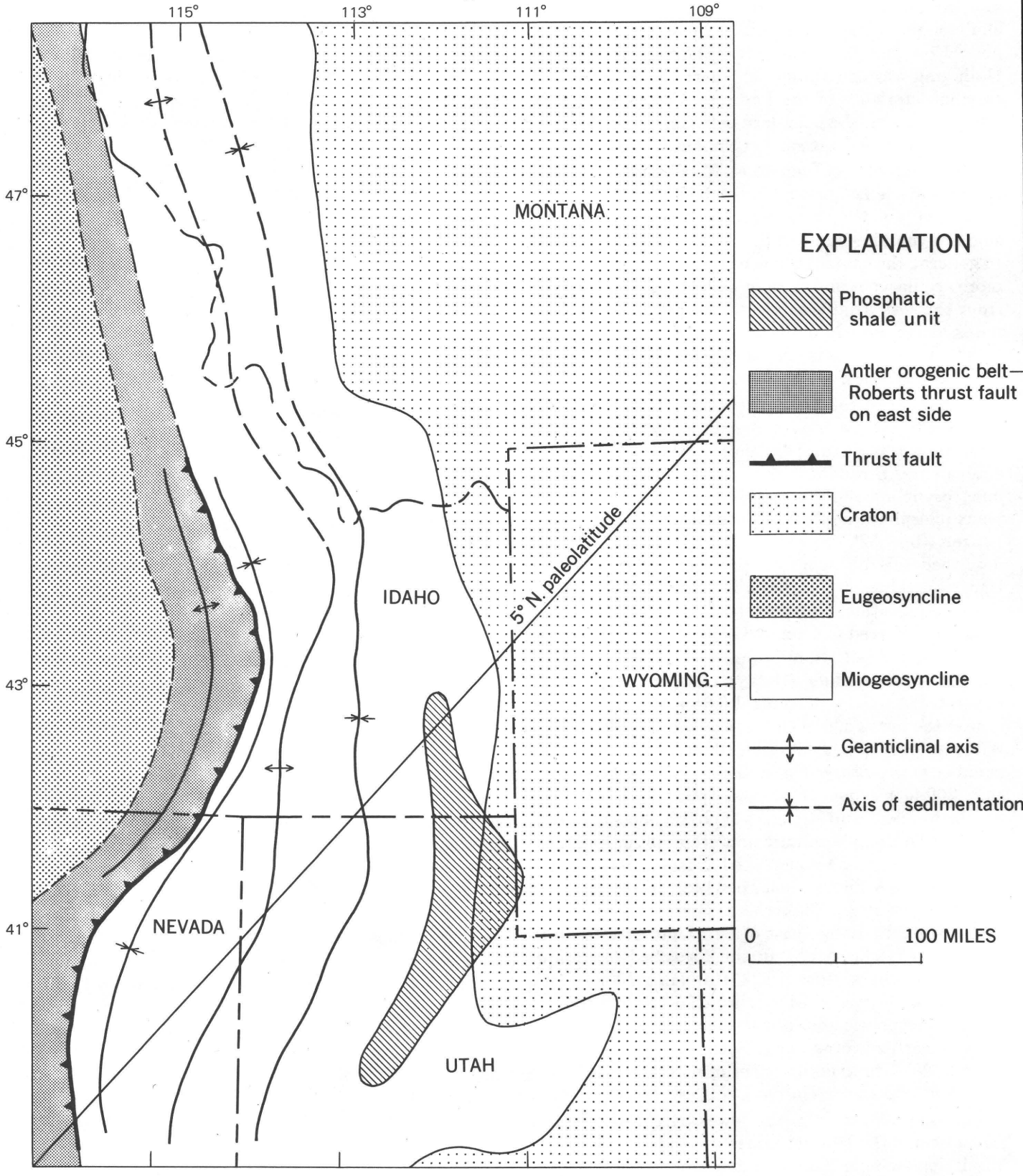

FIGURE 52. — Paleotectonic position of phosphatic shale unit at base of Deseret Limestone–Brazer Dolomite. Compiled by R. P. Sheldon in 1969 from Cheney (1957), Morris and Lovering (1961), Roberts and Thomasson (1964), W. C. Gere (written commun., 1966), and A. E. Roberts (unpub. data, 1966).

member of the Brazer rest unconformably on the Lodgepole Limestone in north-central Utah. This proposed mid-interval B unconformity is not evident in the rocks; thus, the two formations apparently are conformable. Some paleontologic evidence also suggests continuous deposition. In this study the interval B sequence of north-central Utah is interpreted as essentially conformable (pl. 15, cols. 107–109).

UPPER BOUNDARY OF INTERVAL B

The boundary between intervals B and C is assigned to the base of the Charles Formation or the base of the upper member of the Mission Canyon Limestone or their equivalents (pl. 15, cols. 101, 103). The base of the Charles in west-central Montana (including the section in the type well) is placed at the base of the lowest thick anhydrite bed in a sequence of interbedded carbonate and evaporite beds that overlies the Mission Canyon Limestone. Throughout southwestern Montana the leaching of this basal anhydrite has formed a conspicuous laterally continuous solution breccia (Roberts, 1966, p. B17) that provides a marker between intervals B and C. In parts of eastern Montana, where this basal anhydrite is missing, the base of the Charles is generally placed at the base of a persistent dark-gray shale — the *Richey shale* of subsurface usage. This shale is about 200 feet stratigraphically higher than the basal anhydrite.

In north-central Utah the upper boundary of interval B is placed between members 2 and 3 of the type Brazer Dolomite (pl. 15, col. 108) and their stratigraphic equivalents in adjacent areas.

In the miogeosynclinal area of central Idaho the upper boundary of interval B is at the top of the McGowan Creek Formation as it is recognized in the Lost River Range. The upper member of the McGowan Creek is a mudstone overlain by sandy carbonate rocks of the Middle Canyon Formation. Near Pocatello, Idaho, the boundary is placed between the lower and upper members of the Deep Creek Formation.

THICKNESS TRENDS

Interval B thickness trends illustrate the first downwarping of the shelf in central Montana to form the Central Montana trough (pl. 4-A). This tectonic feature extends eastward from the miogeosyncline to the Williston basin. Strata of interval B are thick relative to the other Mississippian intervals throughout the shelf area of Montana, western Wyoming, north-central Utah, and southeastern Idaho. The carbonate sequence deposited in the Central Montana trough ranges in thickness from 800 feet to more than 900 feet. The miogeosyncline was a belt that included many isolated subsiding troughs and basins, and the thickness of interval B in most of these structures (900–1,500 feet) is similar to the thickness in the center of the Williston basin (900–1,250 feet). Downwarping of the central part of the miogeosyncline was greater; in central Idaho, rocks assigned to interval B are more than 2,000 feet thick (pl. 4-A). Along the hinge line between the shelf and the miogeosyncline in western Wyoming, south of the Central Montana trough in southern Montana, and around the periphery of the Williston basin is a fairly continuous reefoid zone indicated by the dual or closed 500-foot isopachs. In this reefoid zone oolite and calcarenite deposits are common. In many places in southeastern Montana these deposits are associated with bioherms, which in some places formed mounds on the sea floor. Thickness trends, lithology, and paleontologic evidence suggest that this widespread sequence of carbonate rock was deposited on a shallow shelf which was gradually downwarping in central Montana. Thickness trends and lithologies in the miogeosyncline suggest a zone of folding being filled by very fine grained clastics in the western part and gradually changing to arenaceous carbonates in the eastern part (pl. 4-B).

LITHOFACIES TRENDS

The interval B sea received a smaller influx of very fine grained detrital material than did the interval A sea; in the resulting clear water, thick relatively pure carbonate rocks characteristic of interval B were deposited. Interval B is not only the most widespread of the Mississippian intervals, but it also has the most consistent lithologic continuity for the shelf area of Montana, western Wyoming, southeastern Idaho, and north-central Utah (pl. 4-B). The dominant lithofacies for interval B consists of massive- to thin-bedded very finely to coarsely crystalline limestone, magnesian limestone, dolomitic limestone, calcitic dolomite, and dolomite. The calcium and magnesium content in these carbonates differs from bed to bed and indicates that these metals were deposited cyclically (Roberts, 1961, p. B294). A few anhydrite beds occur in this carbonate sequence, and in subsequently uplifted areas they have been leached and thus have formed solution breccias. Chert is also common in interval B, particularly in the lower member of the Mission Canyon Limestone and the Deseret Limestone. Oolite beds, in many areas laterally persistent for miles, are characteristic of this interval in southern Montana and western Wyoming.

In the miogeosynclinal area of east-central Idaho, interval B, the upper part of the McGowan Creek Formation, consists of massive- to thin-bedded dark-gray to black carbonaceous shale, mudstone, and siltstone with some quartzose sandstone and argillaceous limestone.

The sequence is calcareous in part and contains laminae of chert locally.

A persistent dark-gray shaly unit which commonly contains phosphate is present at the base of the Deseret Limestone and the Brazer Dolomite or their stratigraphic equivalents throughout north-central Utah and southeastern Idaho. This unit consists of dark-brown to black thin-bedded mudstone interbedded with dark-gray to black thin- to thick-bedded argillaceous carbonate rocks and thin, probably discontinuous, layers of grayish-brown and black pelletal phosphorites.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Detrital rocks in the western part of the miogeosynclinal belt probably were derived from a source area to the west or southwest and were deposited in euxenic or brackish conditions that began during interval A time. To the east they grade into carbonate rocks that were deposited under dominantly normal marine conditions. The presence of anhydrite beds in central and eastern Montana may indicate periodic restriction of circulation in the cratonic seas.

The Deseret-Brazer phosphatic shale was deposited in about mid-interval B time along the hinge line between the miogeosyncline and the shelf (fig. 50) with a northerly trending axis of deposition in north-central Utah and southeastern Idaho (fig. 52). The paleoequator ran northeastward through the southeastern part of Wyoming (Sheldon, 1964, p. C111). Thus, the Deseret-Brazer phosphatic shale was deposited at 3°–5° N. paleolatitude. Conditions for deposition of this phosphatic unit are as follows: (1) an adjacent deep sea; (2) low paleolatitudes; (3) upwelling of cold water to the surface; (4) adjacent lands that were low and not shedding large amounts of detrital material or volcanic debris into the sea; and (5) tectonism quiet enough to allow these conditions to prevail over a long period of time.

PALEOTECTONIC IMPLICATIONS

During interval B the Williston basin subsided but was connected to the miogeosynclinal belt to the west by an east-trending subsiding trough, the Central Montana trough (pl. 4-A; pl. 10, fig. 2). The trough continued to subside through the remainder of Mississippian time.

The greater thickness of interval B (pl. 4-A; pl. 10, fig. 2) than interval A (pl. 3-A; pl. 10, fig. 1) in the shelf area indicates either increased epeiric subsidence of the craton or increased eustatic rise of sea level during interval B.

INTERVAL C

FORMATIONS INCLUDED

SHELF AREA

Interval C consists of the upper member of the Mission Canyon Limestone near Three Forks in southwestern Montana (pl. 15, col. 103). This stratigraphic unit is present in southeastern Idaho, western Wyoming, and western southern Montana. The stratigraphic equivalent in central eastern Montana is the Charles Formation (pl. 15, col. 101), and in northwestern Montana it is the Sun River Member of the Castle Reef Dolomite (pl. 15, col. 102). In north-central Utah in the Crawford Mountains, interval C consists of member 3 of the type Brazer Dolomite or its stratigraphic equivalents (pl. 15, cols. 107–109), member B and the upper part of member A of the *Brazer Limestone* in the Bear River Range, and unit 2 and the upper part of unit 1 of the *Brazer Formation* in the Pisgah Hills.

MIOGEOSYNCLINAL AREA

In the Lost River Range of central Idaho the interval B–interval C boundary is placed at the base of a transitional sequence from the dark-gray shale and argillite of the McGowan Creek Formation to the carbonate sequence of the *White Knob Group.* This transitional unit, the Middle Canyon Formation (Huh, 1967, p. 37), consists of calcareous quartzose siltstone and limestone, and it contains a microfauna of Meramec age (Mamet and others, 1971). Above the Middle Canyon the Scott Peak Formation (Huh, 1967, p. 39–40) is assigned to interval C. The Scott Peak contains beds of Chester age at the top (Mamet and others, 1971). In south-central Idaho equivalent strata consist of the upper member of the Deep Creek Formation and the lower member of the Great Blue Limestone (Carr and Trimble, 1961, p. C181). In north-central Utah, near Morgan, interval C consists of the upper part of the Deseret Limestone and the Humbug Formation (Mullens and Laraway, 1967).

STRATIGRAPHIC RELATIONS

The upper member of the Mission Canyon Limestone or its equivalent, the Charles Formation, consists chiefly of massive finely crystalline cyclically bedded limestone, dolomitic limestone, calcitic dolomite, and dolomite interbedded with fine-grained clastics and evaporites or solution breccias of dolomite and dolomitic grayish-red siltstone. The evaporites are predominantly anhydrite but include lesser amounts of gypsum and salt. The cyclically bedded carbonate rocks of the Charles Formation are generally similar to those of the Mission Canyon except that the Charles rocks

are less fragmental. In western Montana and western Wyoming the siltstone beds are ferruginous and well laminated, and they weather to form indentations between the more resistant limestone beds (Roberts, 1966, p. B7). These siltstone beds are in the stratigraphic positions of the evaporite zones in the upper member of the Mission Canyon Limestone or in the Charles Formation. The red siltstone beds have been mistaken for matrix materials of karst-filled deposits, which they resemble in color, lithology, and grain size.

The laterally continuous solution breccia beds exposed in the surface sections correlate with the evaporite beds in the subsurface sections (Roberts, 1966). Some of the massive anhydrite beds are locally as much as 100 feet thick, and the correlation with persistent breccia beds and the distinction of these solution breccia beds from similar-appearing karst deposits in the Madison Group are important in the stratigraphic analysis of interval C.

In uplifted areas of central Montana, removal of soluble minerals from the basal anhydrite of the Charles Formation formed a laterally persistent breccia. In this area the rocks are included in the Mission Canyon Limestone, and this persistent breccia marks the base of the upper member of the Mission Canyon Limestone (Roberts, 1961, p. B295; 1966, p. B19).

The solution breccia at the base of the upper member of the Mission Canyon Limestone forms prominent cliffs in the northwestern part of Yellowstone National Park near Bannock Peak. The breccia is continuous throughout the area and is 60 feet thick and 310 feet below the top of the upper member — a markedly similar position to that in the Livingston, Mont., area. Similar stratigraphically continuous breccia beds in the Mission Canyon Limestone are exposed in Montana in the Pryor Mountains (Blackstone, 1940, p. 594; and Richards, 1955, p. 22), in the Bridger Range (McMannis, 1955, p. 1400), in the Little Rocky Mountains (Knechtel, 1959, p. 735), in the Elkhorn Mountains (Klepper, 1950), and in Milligan Canyon near Three Forks (Robinson and Barnett, 1963, p. 42).

The solution breccia beds probably formed after Pennsylvanian time, for they are restricted to areas of Late Cretaceous and early Tertiary uplift. In areas that were not uplifted, the evaporite zones in the Madison Group are generally unaltered. For example, about 25 miles northeast of the Fairy Lake section in the Bridger Range, Mont., three evaporite zones were penetrated in The California Co.'s Crowley well 1. Similar zones were found in the subsurface less than 10 miles from the Big Snowy Mountains. In some localities there may be a surface expression in younger rocks that would indicate the presence of underlying solution-breccia beds. Such an example of large sinks or depressions due to collapse over major solution zones was observed near Monarch and Riceville in the Little Belt Mountains, Mont., and was described by Severson (1952, p. 41). He noted that rocks as young as Colorado Shale of Cretaceous age were affected in this area and, accordingly, that solution activity at this locality occurred after deposition of the Colorado Shale. It is possible, however, that not all of the solution activity and consequent collapse took place at the same time but that each area of uplift had its own history.

The solution breccia deposits are composed of red clayey siltstone matrix in a heterogeneous mixture of small angular carbonate and chert fragments. Within the breccia deposit the number of fragments increase from the base upward into slightly fractured limestone and dolomite. The upper surface of the solution breccia is poorly defined; however, the lower surface of the solution breccia is generally sharp, well defined, and continuous. The solution breccia units are relatively continuous along the outcrop. Illite is the chief clay mineral in the matrix of the solution breccia (Roberts, 1966, p. B21).

UPPER BOUNDARY OF INTERVAL C

Uplift near the close of Meramec time in western Montana and western Wyoming was accompanied by erosion. The upper beds of the Mission Canyon Limestone or Charles Formation were truncated, and, locally, a karst topography was developed. Similarly, in north-central Utah, at the west end of the Uinta Mountains, a red clastic unit commonly occurs at the base of the Doughnut Formation (interval D of this study and equivalent to part of the Manning Canyon Shale and part of the Great Blue Limestone). Sadlick (1955, p. 54) interpreted this red-bed unit as a karst deposit that formed on the Humbug Formation while the Great Blue Limestone was being deposited in the Chester Sea to the west.

Deposition during the close of interval C and the beginning of interval D was continuous in the Central Montana trough of Montana and in the miogeosynclinal area of central Idaho and north-central Utah.

THICKNESS TRENDS

Interval C ranges in thickness from 500 feet to more than 700 feet in the Central Montana trough and thins northward to a zero edge in northern Montana and southward to a zero edge in northern Wyoming. In Idaho in two local basins in the miogeosynclinal area, interval C attains maximum thicknesses of more than 2,500 feet (pl. 5-A). Thickness of the upper member of the Mission Canyon Limestone or the Charles Forma-

tion may vary considerably in adjacent surface and subsurface sections. Comparison of thickness, lithology, and fabric on a regional basis suggests that interval C must have been reduced in near-surface and surface sections by internal solution during a period of gradual uplift after Mississippian time took place, probably during or after the Late Cretaceous and early Tertiary Laramide uplifts (Roberts, 1966, p. B21). The isopach map also reflects post-interval C or post-Mississippian erosion or both in parts of Montana, Wyoming, and north-central Utah. Post-Mississippian erosion and solution make meaningful depositional thickness trends difficult to identify.

LITHOFACIES TRENDS

Deposition during interval C produced the following generalized distribution of facies: limestone in the miogeosynclinal area of central Idaho; detrital sedimentary rocks, dolomite, and limestone in the miogeosynclinal area of south-central Idaho and north-central Utah; dolomite in the shelf areas of eastern and southeastern Idaho, western Montana, western Wyoming, and northeastern Utah; and dolomite interbedded with gypsum or anhydrite in the shelf areas of central and southern Montana and in northern Wyoming (pl. 5-B). Locally, abrupt lithologic changes occur in short distances. Also, in many places in the shelf areas the lithologic succession of limestone, dolomite, and gypsum or anhydrite is repeated vertically in cyclic fashion suggesting alternating periods of shallowing and deepening water.

The limestone of interval C is light to medium gray and light olive gray, fine to coarse grained, and generally in beds 1 foot or more thick. The limestones are dense and often argillaceous with meager or no fossil content. Brownish-gray chert, generally in nodules, is common. Many of the coarser grained limestones are bioclastic and contain brachiopods and corals. Dolomite and dolomitic limestone are light gray to yellowish gray and are the predominant lithologies of the interval. The dolomite is commonly argillaceous. Solution breccias, composed mostly of dolomite fragments, are conspicuous in outcrops of interval C, and beds of gypsum and anhydrite mark the stratigraphic position of the breccias in the subsurface. The anhydrite and gypsum beds range from pure to very argillaceous; silt and sand is disseminated locally.

The thickest and most widespread of the evaporite units (diagrammatically shown on pl. 9-F, secs. *O–O'*, *P–P'*) is at the base of interval C (Roberts, 1966). Uplift, probably during Late Cretaceous and early Tertiary time, caused solutions to permeate and subsequently leach surface and near-surface evaporite beds forming stratigrphic solution breccias. In subsurface areas where the gypsum unit at the base of interval C has not been leached out of the sequence, such as western Wyoming, there are large (commercial) quantities of helium. The helium is a decay product from uranium that occurs within small asphaltic pellets in the gypsum. To the east in uplifted areas, such as the Pryor Mountains in south-central Montana, the gypsum has been leached out, but the uranium has been concentrated in the matrix of the solution breccia. Throughout eastern Montana and northern Wyoming this evaporite unit or solution breccia emits gamma rays that form a very characteristic curve on the gamma ray–neutron log making it an excellent regional stratigraphic marker unit.

In the miogeosyncline, deposition of detritus took place during part of interval C time. The upper part of the Deseret Limestone and the Humbug Formation are predominantly sandstone. The Humbug is differentiated from other formations deposited in the thick miogeosynclinal sequence by the presence of numerous beds of sandstone or quartzite. Sandstone beds within the Humbug are medium to fine grained, lenticularly bedded, and in part calcareous and in part quartzitic. These massive, well-sorted sandstones are reworked sediments that become thinner and contain interbedded shale or siltstone toward the east. The grain size of the sands increases toward the west. The sandstone units are interbedded with light- to dark-gray, thin- to thick-bedded cherty limestone and dolomite. In central Idaho, deposition during interval C gradually changed from quartzose siltstone and very fine grained sandstone to a cyclic sequence of coarse- and fine-grained limestone.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

The Northern Rocky Mountains and Adjacent Plains region probably was between 5° N. and 10° N. paleolatitude, and the climate probably was warm (Sheldon, 1964, p. C111). Interval C for most of the region consists of carbonate and carbonate-evaporite sequences. These characteristic lithofacies are considered to have formed as a consequence of cyclic rising and lowering of sea level. Marine cycles are also indicated by alternating deposition of calcium and magnesium carbonate rocks as shown in figure 51. The sparsity of terrigenous detrital material within the carbonate and carbonate-evaporite sequences suggests that the shoreline was at a considerable distance and that land areas had very low relief (pl. 11, fig. 3; pl. 12, fig. 3). During interval C time the sea began a gradual regression from the shelf area of Montana and Wyoming, and carbonate deposition on much of the shelf area became more dolomitic.

Gypsum or anhydrite was deposited on a near-level bottom of a shallow epeiric sea, apparently with no physical barriers. Epeiric seas, according to Shaw (1964, p. 5), must have had very gentle bottom slopes that probably averaged about 0.1 foot per mile and movement of water as open ocean currents probably did not occur in epeiric seas of the past (Shaw, 1964, p. 7). Hypersaline waters could retain their position on the shelf primarily as a result of bottom friction, which would inhibit the development of downslope density currents (Scruton, 1953, p. 2503–2505). Shaw (1964, p. 10) concluded that the gentler the bottom slope in an epeiric sea the higher the salinities that could be expected to develop across a given distance.

The sedimentary cycles on the Mississippian shelf of the Northern Rocky Mountains and Adjacent Plains region and the cyclic pattern of Ca:Mg ratios (Roberts, 1961) may be explained by (1) epeirogenic cyclic crustal movements; (2) eustatic cyclic sea level changes; or (3) cyclic climatic changes. Each or a combination of some of these possibilities is reasonable; however, repeated widespread shelf evaporite sequences suggest sea level changes, probably due to periodic crustal movements in the hinge line between the miogeosyncline and the shelf.

In the miogeosynclinal area of Idaho and adjacent shelf area in western Montana and Wyoming and north-central Utah, the microfossil and algae assemblage of the carbonate and detrital rocks indicates deposition in shallow water. Detrital grains of microcline and plagioclase in the Humbug Formation (Gilluly, 1932, p 26–29) and equivalent strata in north-central Utah and southeastern Idaho and the increasing proportion of detrital rock to the west (pl. 5-B) indicate a source area to the west in Nevada (pl. 10, fig. 3). Livingston (1955, p. 10), on the basis of grain sorting and roundness, stated that the sands of the Humbug Formation were probably derived from preexisting sedimentary rock that had undergone more than one cycle of erosion. Lenticular or crossbedded and ripple-marked sand bodies in the Humbug Formation together with a coral fauna suggest deposition in shallow water on a bottom traversed by strong currents (Livingston, 1955, p. 33).

PALEOTECTONIC IMPLICATIONS

The Central Montana trough subsided gradually (pl. 10, fig. 3) during interval C time. In the miogeosynclinal area of central Idaho the substitution of limestone for terrigenous mud and sand as the dominant sediment during interval C time suggests a lowering of the land area to the west (pl. 10, fig. 3; pl. 12, fig. 3). However, late Osage or early Meramec uplift of older Paleozoic rocks in northeastern Nevada, presumably within the Antler orogenic belt (pl. 10, fig. 3), caused detritus to be shed into the late Deseret sea in the miogeosynclinal area of north-central Utah to southeastern Idaho. This uplift culminated during deposition of the Humbug Formation.

INTERVAL D

FORMATIONS INCLUDED

SHELF AREA

Interval D in the shelf area of Montana, eastern Idaho, and western Wyoming consists of the Big Snowy Formation or Group and equivalent beds (pl. 15). The Big Snowy Group consists of the Kibbey, Otter, and Heath Formations. The Big Snowy Group is restricted at the top to dark-gray to black marine shale and interbedded limestone beneath a Late Mississippian to Early Pennsylvanian regional unconformity by Maughan and Roberts (1967). The Big Snowy Group as thus restricted is mostly, if not entirely, of Late Mississippian (Chester) age.

In the shelf area of north-central Utah interval D includes units 3, 4, and part of 5 of the *Brazer Formation* of Williams (1948), member C of the *Brazer Limestone* of Mullens and Izett (1964), and the Doughnut Formation of Crittenden, Sharp, and Calkins (1952).

MIOGEOSYNCLINAL AREA

The miogeosynclinal rocks of interval D in south-central Idaho and extending into Utah include the middle and upper members of the Great Blue Limestone and the lower part of the Manning Canyon Shale (pl. 15, col. 105).

The Great Blue conformably overlies the Humbug Formation and is conformably overlain by the Manning Canyon Shale. At the type locality the Great Blue Limestone consists of a basal limestone unit, a medial black shale called the Long Trail Shale Member (Gilluly, 1932, p. 29), and an upper limestone unit. In the Deep Creek Mountains near Pocatello, Idaho (Carr and Trimble, 1961, p. C183), and near Provo, Utah (Baker and others, 1949, p. 1176), the Long Trail Shale Member retains a similar stratigraphic position, but to the southwest in the Tintic Mountains (Morris and Lovering, 1961, p. 112) the formation contains more mudstone and this fact introduces some uncertainty in the regional correlation of the Long Trail.

In central Idaho interval D consists of the South Creek Formation and the Surrett Canyon Formation.

STRATIGRAPHIC RELATIONS

The Big Snowy Group, except in one small area in the Pioneer Mountains, rests upon the Mississippian Charles Formation or the equivalent part of the Mission

Canyon Limestone where the Charles is not distinguished or is not present. This contact seems conformable in most of Montana, but erosion may have preceded deposition of the Kibbey in parts of western Montana. In the Three Forks area, Scott (1935, p. 1026) noted that the Kibbey rests unconformably on the Madison Group (pl. 15, col. 103). A karst surface described by Robinson and Barnett (1963, p. 43) in this area has local relief of as much as 100 feet along 300 feet of contact of the basal Kibbey. Walton (1946, p. 1297), Severson (1952, p. 19), and Miller (1959, p. 13) also recognized an erosional unconformity on the upper surface of the Madison in central and southern Montana. At a few places elsewhere in western Montana, Big Snowy rocks rest on markedly thinned Madison. In an isolated exposure in the Pioneer Mountains southwest of Butte, Mont., Big Snowy rests upon the Lodgepole Limestone (G. D. Fraser, oral commun., 1964) which indicates erosion and removal of part of the Madison Group prior to deposition of the Big Snowy Group.

Correlation of the Big Snowy Group becomes increasingly difficult west and southwest of the Big Snowy Mountains in central Montana. Thickness and lithologic changes and limited exposures of the stratigraphic sequence do not permit separation into the individual formations of the type Big Snowy Group. Thus, in parts of southwestern Montana the Big Snowy is reduced in rank to formation. The Big Snowy Formation is composed mostly of light weathering dark-gray limestone and dark-gray shale which very closely resemble the limestone and shale of the Heath Formation, and which have been termed the *Lombard facies* (Blake, 1959). Distribution of the Big Snowy Formation in southwestern Montana is limited by two factors (Sloss and Moritz, 1951, p. 2158). First, parts of the area lie near the depositional margin of the sequence, and the beds are lacking through nondeposition. Second, widespread erosion preceding deposition of the Pennsylvanian Amsden Formation has stripped sediments of the sequence from most of southwestern Montana, and pre-Jurassic erosion truncated the Big Snowy Group progressively from central to northern Montana (Perry, 1951, p. 57). In parts of southwestern Montana and western Wyoming, the Big Snowy Formation is composed of strata equivalent to the Kibbey Formation and is a "red bed sequence" similar to the Amsden Formation. Perry (1949, p. 17) found beds of gypsum 10–15 feet thick near the base of the Big Snowy Formation on Little Sheep Creek, southwest of Lima, Mont., and in the Snowcrest Range (fig. 49) the Big Snowy sequence contains impure beds of gypsum interbedded with shales and siltstones.

Equivalents or partial equivalents of the Big Snowy Group have been mapped or described in parts of southwestern Montana where Kibbey and Heath lithologies are apparent (Sloss and Moritz, 1951; Gealy, 1953; McMannis, 1955; Scholten and others, 1955; Blake, 1959; Hadley, 1960; Robinson and Barnett, 1963). Big Snowy rocks have been recognized on the north flank of the Beartooth Range east of Boulder Canyon in Sweet Grass County in south-central Montana. In western Wyoming, sandstone and red shale in this stratigraphic position (Rubey, 1958) are similar to those in southwestern Montana; however, 50 miles to the west equivalent strata in the Chesterfield Range of southeastern Idaho are cherty limestone of the upper part of the Monroe Canyon Limestone (Dutro and Sando, 1963, p. 1983–1984).

Ruppel (1968) measured and described 850 feet of mudstone, shale, and limestone in the Beaverhead Mountains near Leadore, Idaho, which he tentatively correlated with the Big Snowy Group. This is the westernmost extent of this sequence reported to date, and, in the vicinity of Mackay, Idaho, equivalent strata are carbonates of the *White Knob Group* (Huh, 1967, p. 46).

The Amsden Group overlies the Big Snowy Group throughout central Montana, except for a narrow zone immediately adjacent to the northern flanks of the Little Belt and Big Snowy Mountains, where pre-Jurassic erosion removed the Amsden, but not all of the Big Snowy, prior to Jurassic deposition (pl. 3; Perry and Sloss, 1943, p. 1292). It is not possible to determine the original former northern extent of the Big Snowy Group, because a short distance north of the truncated edge of the Amsden the Big Snowy Group is truncated also and the Jurassic Ellis Group rests upon Madison strata (Perry, 1951, p. 57). The Big Snowy Group changes little in lithology northward in central and eastern Montana, although the thickness varies considerably (Willis, 1959, fig. 3, p. 1945), owing to relief of the Late Mississippian to Early Pennsylvanian unconformity.

Rocks of the Big Snowy Group were deeply eroded or removed completely in Late Mississippian time in south-central Montana south of Musselshell, Golden Valley, Sweet Grass, and Park Counties (fig. 49). Near Livingston, Park County, the upper member of the Mission Canyon Limestone (interval C) is overlain by the Pennsylvanian Amsden Formation (Roberts, 1964). At least 550 feet of interstratified dark-gray shale and limestone of the Heath Formation (restricted) and a total of 1,189 feet of the Big Snowy Group had been deposited in central Montana (beds 13–50 of Potter Creek Dome section (Easton, 1962, p. 118–119)) prior

to this period of erosion. No remnants of the Big Snowy Group are known to be preserved in south-central Montana to indicate how far south these rocks originally may have extended; instead, they are abruptly truncated, and the lithology of the Big Snowy Group now preserved immediately to the north does not suggest a shore facies.

In the miogeosyncline, the thickness of the Mississippian part of the Manning Canyon Shale increases from the Oquirrh Mountains (Gilluly, 1932, p. 32) eastward to Provo, Utah (Baker and others, 1949, p. 1179). The thickness of the Manning Canyon Shale in the Deep Creek Mountains of south-central Idaho is very similar to that of the Oquirrh Mountains (D. E. Trimble, 1965, oral commun.). Both the lower and upper contacts are gradational, which may account for some east-west differences in thickness. The apparent lateral gradation from limestone to shale in the Great Blue Limestone and the lithologic similarity of the shale and limestone in the two formations (Baker and others, 1949, p. 1179) suggest that the contact between them is a facies contact which may be at different stratigraphic positions in different areas.

UPPER BOUNDARY OF INTERVAL D

The Mississippian-Pennsylvanian boundary is a regional unconformity between the Big Snowy and Amsden Groups. It is believed (Maughan and Roberts, 1967) that this unconformity represents latest Mississippian and earliest Pennsylvanian time and that it is nearly contemporaneous with the unconformity (Chamberlin and Salisbury, 1906; Cheney and others, 1945), separating the two systems in the Mississippi Valley.

The length of hiatus represented by the Late Mississippian to Early Pennsylvanian unconformity is not known, but probably it was relatively short, although enough time elapsed prior to deposition of the overlying sediments for the entire Big Snowy Group to be stripped from some areas. Youngest rocks of the restricted Heath Formation of Maughan and Roberts (1967) are of Chester age (Easton, 1962, p. 23) while oldest rocks above the unconformity are either very latest Mississippian or earliest Pennsylvanian. If the overlying rocks are latest Mississippian in age, they cannot be separated lithologically from succeeding beds of definite Pennsylvanian age, and the unconformity is the only place to put the systemic boundary that is satisfactory for mapping.

Differences in rock types between the Heath Formation and the Stonehouse Canyon Member of the Tyler Formation (basal formation of the Amsden Group) are generally slight. The limestone and shale in the Heath are thin bedded, very fine grained, and well sorted, indicating deposition in relatively quiet water. The detrital and carbonate beds of the Stonehouse Canyon are thick bedded, lenticular, very fine to coarse grained, and generally poorly sorted. At most places where there is sandstone at the base of the Stonehouse Canyon, it is conglomeratic. More turbulent conditions must have existed during deposition of the Stonehouse Canyon. These differences in rock types indicate an abrupt change in transporting energies.

Although the contact is poorly exposed, Douglass (1954) mapped it as his Amsden–Big Snowy boundary in the southwestern part of the Big Snowy Mountains; and the same contact has been extensively mapped on the northeast flank of the Little Belt Mountains by Vine (1956). The Heath-Tyler unconformity is present throughout eastern Montana and western North Dakota (Beekly, 1955; Mundt, 1956; Foster, 1956; Willis, 1959; Maughan and Roberts, 1967, pls. 1, 3).

In the miogeosynclincal area in central and southeastern Idaho (pl. 15, cols. 104, 105) the upper boundary of interval D seems conformable. Unnamed Pennsylvanian sandstone rests on the Surrett Canyon Formation in central Idaho. The system boundary lies within the Manning Canyon Shale in southeastern Idaho.

In the hinge area of western Wyoming and north-central Utah, Early Pennsylvanian or pre-Pennsylvanian erosion in places has removed some or all of interval D, and the systemic boundary is marked by a disconformity. In Brazer Canyon (pl. 15, col. 108) in the Wasatch Range in north-central Utah, the Wells Formation of Pennsylvanian and Permian age rests on beds of Meramec age, and beds of Chester age probably are missing. In the western flank of the Wasatch Range, however, the magnitude of the disconformity is less or absent. Beds of Chester age are recognized in parts of the area and the upper boundary of interval D is conformable, as in the Pisgah Hills (pl. 15, col. 107) where the system boundary falls within unit 5 of the *Brazer Formation.*

THICKNESS TRENDS

Interval D is the least widespread of any of the Mississippian intervals. Deposition was greatest in the miogeosyncline and thicknesses of interval D greater than 4,000 feet are shown in central Idaho and north-central Utah (pl. 6-A). In much of the miogeosyncline, deposition continued from interval D into the Pennsylvanian, and lateral changes in thickness are gradual. In the Central Montana trough, however, thicknesses change rapidly. Erosion at the end of Mississippian or beginning of Pennsylvanian time ac-

counts for the rapid thinning of interval D along both the northern and southern margins of the Central Montana trough (pl. 9-F, sec. *P–P'*). Similarly, in north-central Utah, Early Pennsylvanian erosion truncates the top of the Brazer progressively lower in stratigraphic section from the Pisgah Hills eastward to the Crawford Mountains (pl. 15), a thinning of 1,000 feet or more. In the Central Montana trough interval D attains a maximum thickness of almost 1,200 feet and is more than 600 feet thick along most of its length.

LITHOFACIES TRENDS

SHELF AREA

In central and western Montana the Big Snowy Group or its stratigraphic equivalents consist of variable amounts of sandstone, siltstone, shale, carbonate, and evaporite. These lithologic units range in thickness from a few inches to about 20 feet.

The Kibbey Formation in most of central Montana consists of grayish-red siltstone, sandstone, and shale interbedded with limestone, dolomite, and anhydrite. In eastern Montana the formation can be divided into three lithologic units based on recognition of the medial limestone. This unit generally consists of limestone or dolomite although locally it consists of sandstone or anhydrite (Nordquist, 1953, p. 81). In southern Montana the basal sandstone or the Kibbey Formation contains fragments of Madison Limestone and fills caverns and solution channels in the Madison (Sloss and Hamblin, 1942, p. 309). These karst deposits are commonly colored red or maroon by infiltration from the overlying siltstone of the Kibbey Formation. In the Bridger Range the karst deposits occur as deep sinkholelike channels carved into and filled with reworked materials of Mission Canyon Limestone (Laudon, 1948, p. 295).

The Otter Formation consists predominantly of greenish-gray shale. Locally, it also contains gray, purple, and black shale and interbeds of yellowish-gray argillaceous limestone, dolomite, and gypsum. The contact between the Otter and overlying Heath Formations and the contact with the underlying Kibbey Formation are commonly gradational, and the three formations intertongue.

The Heath Formation consists of interbedded dark-gray to black marine petroliferous limestone and shale. The shale is calcareous and silty and commonly contains fairly abundant heavy minerals. Limestone is increasingly abundant from central and western Montana to central Idaho; locally, beds of gypsum are present.

Sandstone beds make up the thickest lithologic units. Some of the sandstones are laminted light olive gray and grayish black, and some contain light patches cemented with carbonate in a darker matrix cemented by silica. Cross-stratification is common. Organic matter is common, and calcite veinlets and stylolites are also present. The intergranular porosity is poor.

The carbonate rocks include both limestone and dolomite in thin beds interbedded with siltstone and sandstone. The carbonates themselves are commonly sandy or silty. They range from medium gray to dark gray, and from olive gray to olive black. The crystal size is mostly very fine to microcrystalline. Pyrite and organic matter are common, and hydrocarbon residues coat the stylolitic surfaces. Many of the limestones contain microfossils and echinoid spines and have a spergenitic texture. Porosity is poor to fair.

The Heath was first considered by Freeman (1922, p. 826) as the source bed for petroleum in adjacent strata, particularly in the overlying Pennsylvanian sandstone reservoirs of the Stonehouse Canyon Member of the Tyler Formation which are oil producers in parts of central Montana.

MIOGEOSYNCLINAL AREA

In north-central Utah and south-central Idaho the Great Blue Limestone consists of a sequence of massive- to thin-bedded, medium-blue-gray to dark-gray limestone which weathers light gray. The formation is generally fine grained and in places is silty and cherty. The Great Blue also contains some beds of dark-brown to black shale and a few thin beds of calcareous and quartzitic sandstone. The Manning Canyon Shale consists predominantly of dark-gray to black shale with interbedded limestone and quartzose sandstone or quartzite and a few beds of siltstone and mudstone. The limestone is generally massive to thin bedded, fine grained, and medium gray to black. The quartzose sandstone or quartzite is generally thin to medium bedded, fine to medium grained, and olive gray to brown. Both the lower and upper contacts of the Manning Canyon are gradational, which may account for some differences in thickness from one locality to another. Baker, Huddle, and Kinney (1949, p. 1179) pointed out the apparent lateral gradation from limestone to shale in the Great Blue Limestone and the lithologic similarity of the shale and limestone in the two formations, which suggests that the contact between them is a facies contact which may be at different stratigraphic positions in different areas.

In central and east-central Idaho, interval D is comprised of the South Creek Formation and the Surrett Canyon Formation. The South Creek conformably overlies the Scott Peak Formation and consists of light- to dark-gray medium- to thin-bedded or platy silty cherty limestone. Conformably overlying the South Creek Formation is the Surrett Canyon Formation comprised of

dark-gray to black massive- to thin-bedded limestone and cherty limestone with a few intercalated calcareous sandstone and siltstone beds.

SOURCES OF SEDIMENT AND ENVIRONMENTS OF DEPOSITION

Near the close of the Meramec, extensive erosion took place along the margins of the Central Montana trough. (Compare pl. 6-A and pl. 10, fig. 4.) Basal beds of the Big Snowy Group include terrestrial "red-bed" clastics and marine dolomite in western Montana and Wyoming, interbedded nonmarine and brackish-water clastics and evaporites in central Montana; and brackish-water clastics, evaporites, and marine shale in eastern Montana. The basal coarse-grained clastics and associated sediments of the Kibbey pass upward into the shallow marine shale of the Otter Formation, which was deposited approximately during the middle of interval D time (pl. 11, fig. 4). Overlying the Otter are the dark-gray shales of the Heath Formation whose fauna suggests a slightly deeper marine environment. Basal beds of the Big Snowy Formation or its equivalent (the upper part of the Scott Peak Formation) west of the hinge line and in the miogeosyncline included marine carbonates and shale that contain algae, crinoids, bryozoa, and microfossils that represent a normal shallow-water marine environment (Mamet and others, 1971).

The location of bordering lands that were source areas for these sediments is uncertain. Only in central North Dakota do the facies suggest nearshore deposition (pl. 6-B). The Kibbey here, as elsewhere, is beveled, and it is not certain how far east the sea may have extended at this Late Mississippian time. Nevertheless, the Transcontinental arch, or Siouxia land area, probably was not far east of this beveled edge in central or eastern North Dakota (pl. 10, fig. 4; pl. 11, fig. 4*A*). Earlier in Mississippian time the shore probably was farther to the east.

Sediments were shed into the miogeosyncline (pl. 10, fig. 4) from land areas in southern Wyoming and Colorado. These sediments were spread widely by marine currents. Much of the sediment undoubtedly came from preexisting Paleozoic sedimentary rocks. Thickness of the Manning Canyon Shale increases toward the west, and the grain size and amount of clastic sediments also increases toward the west indicating a source area in northeastern Nevada and western Idaho.

The environment of the Manning Canyon Shale — as indicated by its fine grain size, its invertebrate and plant fossils, and its abundance of organic material and iron — was brackish lagoonal or tidal flat to shallow marine (Moyle, 1958, p. 35). The water was warm, and reducing conditions allowed large amounts of terrestrial organic material to be preserved. The fossiliferous limestone suggests intermittent shallow open marine conditions. Quartzose sandstone or quartzite in the Manning Canyon consist of reworked well-sorted subangular to subrounded grains. The thickness of these beds is uniform laterally with no evidence of channeling or lensing that would suggest deposition by active currents offshore.

PALEOTECTONIC IMPLICATIONS

A regional uplift, which occurred at the close of the Meramec or the beginning of the Chester, affected the type of sediment deposited, the extent of deposits, and depositional patterns of interval D. In western Montana and eastern Idaho, areas emerged that had been inundated during older intervals of the Mississippian (Compare pl. 10, fig. 4, with pl. 10, figs. 1–3.) Uplifted carbonate beds of the Mission Canyon in western Montana and western Wyoming were eroded forming karst deposits. Sandstone that was spread widely over much of the region, including sandstone in the basal Big Snowy Group in Montana, the Darwin Sandstone Member of the Amsden in Wyoming, and sandstones of the lower member of the Doughnut Formation of Utah, was probably derived by erosion of older Mississippian and Paleozoic rocks on uplifted areas.

The gradation in the Big Snowy Group from dominantly sandstone in the lower part through green and red shale into black shale and limestone in the upper part indicates renewed downwarping and gradual deepening of the sea during interval D time. At the end of Chester time the sea was gradually restricted to the Central Montana trough, and land areas made up of limestone of the Madison Group north and south of the trough were gradually enlarged and exposed to weathering and erosion. Regional uplift of the craton ended deposition of interval D.

Numerous linear boundaries between areas of markedly different thicknesses of Chester rocks or between Chester and Pennsylvanian rocks were interpreted by Maughan and Roberts (1967) as pre-Pennsylvanian faults. These faults bound Chester rocks on the south flank of the Central Montana trough. In southwestern Montana the eastern limit of the Chester rocks is also interpreted as a fault (Maughan and Roberts, 1967). The displacement of this fault is at least 800 feet as shown by 800 feet of the Big Snowy Formation preserved on the west side of the fault compared to none preserved on the east. Probable faults that bound the north flank of the Central Montana trough are post-Pennsylvanian (Maughan and Roberts, 1967). Pennsylvanian rocks which overlap the probably faulted and beveled Mississippian rocks along the south flank of the Central Montana trough are themselves

truncated along the fault system on the north flank of the trough and do not overlap farther north beyond Chester rocks.

In the miogeosyncline, rocks of the Manning Canyon Shale have a crude cyclic alternation of lithologic units that suggests two transgressions and three regressions of the sea (Moyle, 1958, p. 36–39), probably due to a combination of episodic sinking of the geosyncline and rising of the surrounding landmasses.

Conglomerates and related clastic sedimentary rocks of probable Chester age in central Idaho probably reflect recurrent tectonic movement along the Antler orogenic belt, which probably extended from Nevada northward into central Idaho (fig. 50; pl. 10, fig.4; pl. 12, fig. 4*A*). This belt was first uplifted during the Late Devonian–Early Mississippian Antler orogeny (Smith and Ketner, 1968, pl. 1).

TOTAL THICKNESS OF MISSISSIPPIAN SYSTEM

THICKNESS TRENDS

The Mississippian System attains a thickness of as much as 2,800 feet in the east-trending Central Montana trough (pl. 7; pl. 9-F, secs. *O–O'*, *P–P'*). To the north it thins to a minimum of about 700 feet along the central part of the northern Montana border where it has not been affected by post-Pennsylvanian erosion, and to the south it thins to about 800 or 900 feet along the Wyoming-Montana State line. To the west in central Idaho, miogeosynclinal rocks attain a maximum thickness of nearly 9,000 feet.

PALEOTECTONIC IMPLICATIONS

Much of the region was a stable continental shelf or a moderately submerged segment of the craton during Kinderhook, Osage, and Meramec time. Overlap of Lower Mississippian strata progressively southward indicates gradually sinking land areas in south-central Wyoming and farther south in Colorado that probably included the Ancestral Rocky Mountains (Maughan, 1963, p. C26). A regional uplift, the Milk River uplift, may have been slightly positive in Early Mississippian time north of Montana (Maughan and Roberts, 1967, p. B23). This uplift was probably more active later in the Paleozoic and early Mesozoic when it was centered in Alberta and Saskatchewan, but it may have formed a shoal area in Meramec time that intermittently restricted free circulation northward within the Madison Sea in the Williston basin and caused deposition of evaporites and dolomites in the Charles Formation.

Regional uplift occurred near the end of Meramec and the end of Chester time, intervals C and D, respectively. During the interval C uplift, only small amounts of Mississippian rocks were stripped away; during the interval D uplift, however, which was more prolonged and widespread, large amounts of Upper Mississippian rocks were stripped from some areas.

In the miogeosyncline, downwarping was continuous during the Mississippian, although the rate fluctuated from place to place and from time to time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

Pennsylvanian rocks unconformably overlie the Mississippian in most areas of the Northern Rocky Mountains and Adjacent Plains region (pl. 8). The Amsden Group overlies the Big Snowy Group throughout central Montana, except for a narrow zone immediately adjacent to the northern flanks of the Little Belt and Big Snowy Mountains, where pre-Jurassic erosion removed the Amsden, but not all of the Big Snowy Group, prior to Jurassic deposition (pl. 3-A; Perry and Sloss, 1943, p. 1292). The original former northward extent of the Big Snowy Group is unknown. A short distance farther north of the truncation by the Amsden by pre-Jurassic erosion the Big Snowy Group is truncated also and the Jurassic Ellis Group rests upon Madison strata (Perry, 1951, p. 57).

Rocks of the Big Snowy Group were deeply eroded or removed completely in Late Mississippian time in south-central Montana south of Musselshell, Golden Valley, Sweet Grass, and Park Counties. In areas where the Big Snowy Group has been removed, formations of the Amsden Group rest unconformably on the upper member of the Mission Canyon Limestone.

In western Wyoming, rocks equivalent to the Big Snowy Group were partially or completely truncated during Late Mississippian to Early Pennsylvanian time. In this area the Amsden Formation rests unconformably on remnants of the Big Snowy or the upper member of the Mission Canyon Limestone.

The Big Snowy Group extends as far west as the Beaverhead Mountains near Leadore, Idaho (Ruppel, 1968). At this locality quartzitic sandstone, dolomite, and limestone assigned to the lower unit of the Quadrant Formation overlie the Big Snowy sequence. This lower unit probably correlates eastward with the Tyler Formation of the Amsden Group. West of Leadore, in the vicinity of Challis, Idaho, strata equivalent to the Big Snowy Group form a carbonate sequence, assigned to the upper part of the *White Knob Group*, which is overlain by an unnamed Pennsylvanian sandstone that probably correlates eastward with the Tyler Formation.

In southeastern Idaho, near Pocatello, the uppermost stratigraphic unit in interval D is the lower part of the Manning Canyon Shale. The upper part includes beds of Early Pennsylvanian age (Carr and Trimble, 1961, p. C183). In the Chesterfield Range, near Soda Springs, Idaho, the cherty limestone member of the Monroe Canyon Limestone is unconformably overlain by the Wells Formation of Pennsylvanian and Permian age (Dutro and Sando, 1963).

Late Mississippian to Early Pennsylvanian erosion removed parts of Mississippian intervals C and D in north-central Utah (pl. 15, cols. 107, 108). In the Crawford Mountains the Wells Formation rests unconformably on member 3 of the type Brazer Dolomite (Sando and others, 1959, p. 2761). West of the Crawford Mountains in the Bear River Range the Oquirrh Formation of Pennsylvanian and Permian age unconformably overlies the *Brazer Limestone* (Mullens and Izett, 1964). Near Morgan, Utah, the upper member of the Doughnut Formation is conformably overlain by the Wells Formation (T. E. Mullens, written commun., 1970).

BASAL PENNSYLVANIAN KARST DEPOSITS IN MONTANA

The Lower Pennsylvanian basal part of the Amsden Group consists of widespread residual karst deposits which are composed of red or maroon clayey siltstone matrix that contains chaotically distributed carbonate and chert fragments. The deposits, whose upper and lower surfaces are poorly defined, are laterally discontinuous.

In south-central Montana collapsed caverns, or sinkholes, formed along the joints and bedding planes in the upper part of the Madison prior to the deposition of the overlying Pennsylvanian Amsden Formation (Thom, 1923, p. 42; Thom and others, 1935, p. 35; Richards, 1955, p. 5). Along the north flank of the Beartooth Range a deeply weathered surface and a red residual soil occurs on the top of the Madison (Knappen and Moulton, 1930, p. 11). Henbest (1958) briefly described the regional significance of the karst terrane in Upper Mississippian and Lower Pennsylvanian rocks in the Rocky Mountain region.

In south-central Montana the karst deposits are lithologically very similar to the solution breccia beds. Both are unstratified fragmental deposits that consist primarily of carbonate rocks and lesser amounts of chert in a matrix that weathers reddish or yellowish gray. The sorting is poor, and the range in size of the fragments is wide. The matrix is composed of indurated clayey siltstone with limestone fragments and calcite and smaller amounts of quartz. The fragments in both the karst deposits and the solution breccia beds are mostly angular to subangular, although some carbonate fragments are rounded; these rounded fragments may have been shaped by the action of solution. The chert in both types of deposits is sharply angular.

Kaolinite is the chief clay mineral in the insoluble residues from the karst deposits, and illite is the chief clay mineral in the insoluble residues from the bedded red siltstones and from the basal solution breccia of the upper member of the Mission Canyon Limestone (Roberts, 1966, p. B21). Illite is also the chief clay mineral in samples collected from the lower, middle, and upper parts of the overlying transgressive Amsden Formation. Thus, though the Amsden sediments and karst deposits are lithologically similar, the predominance of illite in the Amsden and the predominance of kaolinite in the karst deposits indicate different sources and origins. Kaolinite in the karst deposits may have formed during prolonged weathering of an ancient soil that was developed on the uplifted surface of the Mission Canyon. The illite in the siltstone and solution breccia beds was deposited in marine conditions and was not changed during uplift or brecciation.

PALEOTECTONIC IMPLICATIONS

Gradual regional subsidence initiated Pennsylvanian deposition in central and eastern Montana. A shallow Pennsylvanian Amsden sea spread into the Central Montana trough and from there southward, and it inundated southern Montana and most of Wyoming. At first, red detrital sediments were deposited in this sea. As the sea spread, deposition of red beds gave way to deposition of carbonate. The Amsden Group comprises a sedimentary cycle of sandstone, shale, and limestone that closely resembles the Big Snowy Group sedimentary cycle of interval D (Maughan and Roberts, 1967).

Uplift and erosion in north-central Montana during early Mesozoic time truncated and consequently thinned the Mississippian rocks northward before deposition of the Jurassic. Local uplifts between Late Mississippian and early Tertiary time brought evaporite beds of the Madison Group near the surface in western and southern Montana and western and northern Wyoming. The evaporite beds were leached, and the Madison was reduced in thickness in these areas.

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Williston Basin Region

By RICHARD P. SHELDON *and* M. DEVEREUX CARTER

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–O

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

WILLISTON BASIN REGION

By RICHARD P. SHELDON and M. DEVEREUX CARTER

ABSTRACT

Mississippian rocks of the Williston Basin region of this chapter are in the North Dakota, Montana, and South Dakota parts of the Willison basin and Central Montana trough and on the craton south and east of the basin. The Mississippian rocks are underlain by the Bakken and Englewood Formations within most of the basin and by older Devonian, Silurian, Ordovician, Cambrian, and Precambrian rocks near the margins. The lower boundary of Mississippian interval A is the contact between pre-Mississippian rocks and basal glauconitic limestone of the Lodgepole Limestone or a correlative surface. Conformity within the basin and disconformity near the margins indicate development of a structural basin before Mississippian deposition.

Mississippian intervals A, B, C, and D in this region are essentially marker-bed units (para-time-rock units) that approximate the Kinderhook, Osage, Meramec, and Chester Provinicial Series. The terms Madison Group and Big Snowy Group of the Montana outcrop belt are also applied in the Williston basin. Formations in the Madison Group intertongue in the basin.

Interval A consists of the lower parts of the Lodgepole Limestone and, in the Black Hills area, the lower part of the Pahasapa Limestone. Deposition occurred in a shallow clastic-starved basin; subsidence exceeded deposition, causing stagnation below normal marine surface waters. Interval thickness is more than 100 feet on the shelves and less than 100 feet in the central basin. Relatively pure carbonate rock on the shelf grades through shaly limestone and spicular chert on the basin flanks to calcareous shale containing organic matter in the basin.

Interval B consists of the intertonguing part of the Lodgepole Limestone, the lower part of the Mission Canyon Limestone, and an anhydrite tongue of the Charles Formation. Time-parallel shale marker beds cut across the time-transgressive lithologic units. Facies distribution is similar to that of interval A, but lagoonal evaporitic and algal carbonate mud formed behind the oolitic and pellitic banks on the shelves. In contrast to A, interval B is thinner on the margins, is 800–1,200 feet thick in the basin and the Central Montana trough, and contains shallow-water carbonate rock at the top of the interval across the basin. Mild tectonism in the region was restricted to basin subsidence and, possibly, to folding on the Cedar Creek and Nesson anticlines.

Interval C consists of two units: (1) a lower unit of basinal shallow-water limestone of the Mission Canyon intertonguing with marginal lagoonal evaporite and dolomite of the Charles Formation, and (2) an upper unit of Charles evaporitic rocks, including basinal halite, and thin beds of red sandstone and mudstone. Widespread marker beds are present. The upper boundary of the interval is a probable disconformity below the Kibbey Formation. Thickness of the interval is more than 1,300 feet in the Williston basin and as much as 600 feet in the Central Montana trough, which connected the semirestricted sea with the normal marine ocean to the west. Activity along the Cedar Creek trend may have reduced normal marine flow into the basin. Subsidence was generally slower than sedimentation, and at the end of interval C time the region was mildly positive.

Three widespread and lithologically fairly uniform units — the Kibbey, Otter, and Heath Formations (in ascending order) of the Big Snowy Group — make up interval D, which contains more detrital rock than does interval C and only minor evaporitic material. The interval is as much as 500 feet thick in the central basin and thickens to more than 1,000 feet in the Central Montana trough. The Big Snowy Group represents a single episode of deposition, preceded and followed by subaerial erosion. Low-amplitude epeirogenic movements caused continued basin subsidence and probably northward and eastward extensive transgression; the Siouxia arch probably was a barrier to the south and the Cedar Creek anticline was a shoal within the sea.

The Mississippian System has a simple pattern of maximum thickness in the central Williston basin (almost 2,800 feet) and in the Central Montana trough, and minimum thickness on the shelves, enhanced by marginal erosional beveling, except on the south. The post-Mississippian unconformity has only minor relief. Pennsylvanian rocks overlie Mississippian in the central basin area and are overlapped on the north and east by Jurassic and Cretaceous rocks.

Following the usage of Sandberg (1962), the southwestern and western limits of the Williston basin (fig. 53) are placed at the northeast flank of the Black Hills uplift, the northeast flank of an unnamed arch that connects the Black Hills and Central Montana uplifts, the north flank of the Central Montana uplift, and the east flanks of the Little Rocky and Bearpaw Mountains. The western limit of the area studied in this chapter was drawn along county lines approximating this tectonic boundary of the basin. Sandberg (1962, fig. 1) placed the eastern limit between the –500-foot and –1,000-foot structure contours drawn at the base of Mississippian rocks. However, for our investigation the Mississippian rocks are studied to their eastern and southern subcrop, which extends east and south of the basin onto the craton. With the minor exception of the small area of outcrop on the northeastern flank of the Black Hills, the Mississippian rocks of this region lie wholly in the subsurface and are known only from well logs and cores.

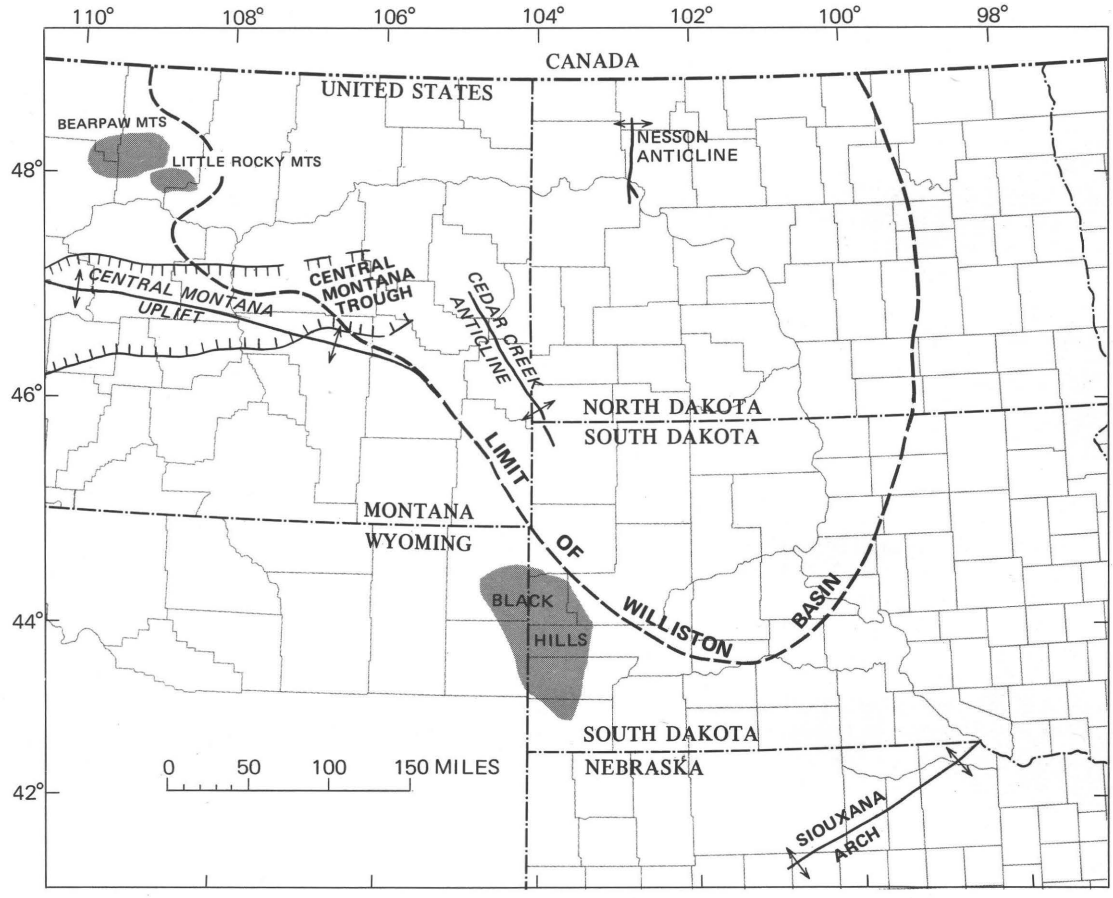


FIGURE 53. — Localities and geographic features in the Williston Basin region referred to in text. Pre-Mississippian tectonic features in italics, Mississippian tectonic features in bold type, and post-Mississippian tectonic features in other type.

REGION DEFINED

The Williston basin is a structural intracratonic basin covering about 200,000 square miles in the high plains of the United States and Canada. Only the United States portion of the basin, which lies in the States of North Dakota, Montana, and South Dakota, is dealt with in this chapter.

PALEOGEOLOGY

The geologic map of units directly beneath the base of the Mississippian System (pl. 2) for the Williston basin was constructed from data supplied by C. A. Sandberg. It was prepared by combining information shown on published maps (Sandberg, 1961, pl. 6; Sandberg and Mapel, 1967, figs. 4, 5, and 10) and more recent subsurface data (C. A. Sandberg, written commun., 1970). The base of interval A and the base of the Mississippian System as dealt with in this chapter are drawn at the contact between the Lodgepole Limestone and the underlying Bakken and Englewood Formations. The Bakken and Englewood are of Mississippian and Devonian age on the basis of conodont evidence; they were discussed by Sandberg (1963), Sandberg and Klapper (1967), Sandberg and Mapel (1967), and Macqueen and Sandberg (1970).

UNITS UNDERLYING MISSISSIPPIAN

Except for a few small areas in the Williston basin, interval A of the Mississippian is underlain by the Bakken or Engelwood Formations. These two formations are probably equivalent in age but were deposited in two basins of deposition (Sandberg, 1963). There are no fossil data concerning the age of these formations in the Williston basin, but Sandberg has made lithologic and mechanical log correlations of beds of these formations to areas of their outcrop and to the subsurface of Alberta, where Mississippian and Devonian conodonts have been found. These correlations show overlap of the beds of the Bakken and Englewood Formations

toward the edges of their respective basins of deposition, and so it is likely that only beds of Mississippian age of the Bakken and Englewood directly underlie interval A.

In a narrow belt, 10–25 miles wide, encircling the deeper parts of the eastern and southern parts of Williston basin and in a large area in the southwestern part of the basin, in the vicinity of the Cedar Creek anticline and the Central Montana uplift, interval A is underlain by Devonian rocks older than the Bakken and Englewood and in small part by Silurian, Ordovician, or Cambrian rocks (pl. 2). These older Devonian rocks comprise, from older to younger, the Dawson Bay, Souris River, Duperow, Birdbear, and Three Forks Formations. In general, the younger formations subcrop toward the center of the Williston basin and the older rocks subcrop adjacent to or on the ancient positive areas marginal to the basin. Their distribution on the pre-interval A paleogeologic map is partly masked by the overlapping younger Devonian Bakken and Englewood Formations, but Sandberg's map (1961, pl. 6) of the pre-Bakken and pre-Englewood paleogeology shows that undivided Silurian-Ordovician-Cambrian rocks subcrop at the crest of the Cedar Creek anticline in southeastern Montana in the positive areas of southeastern North Dakota and south-central and northwestern South Dakota. Along the eastern and southern edges of the Williston basin Sandberg has delimited areas where Ordovician and Precambrian rocks subcrop beneath Mississippian rocks (pl. 2).

LOWER BOUNDARY OF MISSISSIPPIAN

The lower boundary of interval A is drawn at the base of a thin glauconitic limestone (Sandberg and Klapper, 1967, p. B43, B67) that occurs at the base of the Lodgepole Limestone throughout the Williston basin. In most of the area, this boundary is easy to recognize in the mechanical logs of oil-well tests. Also, where the Bakken and Englewood Formations underlie the Lodgepole Limestone, the high radioactivity of both the Bakken and Englewood serves to differentiate them from the Lodgepole. Where the Lodgepole is underlain by rocks older than the Bakken and Engelwood, differences of lithology and electric log characteristics are generally distinctive. However, in the south-central part of the Williston basin, the lower part of the Lodgepole becomes shaly, slightly radioactive, and dark in color; thus, it is distinguished with difficulty from the Bakken Formation. Physical stratigraphic correlations into this area using mechinical logs show the facies change of the lower part of the Lodgepole Limestone into shale above the thin glauconitic limestone, whereas the Bakken Formation remains about constant in lithology and thickness.

Despite the minor difficulties, the lower boundary of interval A is easy to place in the Williston basin and represents a relatively consistent stratigraphic boundary. Along the flanks of the Williston basin it probably represents a minor disconformity at which the thin Bakken and Englewood Formations were beveled. However, in the center of the basin the Lodgepole and Bakken show conformable relations, although the contact is quite sharp. It is possible that the boundary between them represents a time gap, but this question must remain open until fossil data are collected to resolve it. In outcrops throughout Montana the lower glauconitic limestone of the Lodgepole and the underlying black shale — called Bakken Formation in the eastern part and Cottonwood Canyon Member of the Lodgepole in the central part — represent the same or adjacent conodont zones (Sandberg and Klapper, 1967).

PALEOTECTONIC IMPLICATIONS

The map of units underlying the Mississippian System (pl. 2) shows that the Williston basin was already structurally well developed before the sediments of interval A were deposited. The pre-Bakken beveling of the Three Forks Formation and older rocks clearly shows the differential uplift and erosion of the flanks of the basin relative to the central part; that is, the younger rocks are preserved in the central parts of the basin, whereas only the older rocks are preserved on the flanks. The Bakken Formation is now preserved only in the central part of the basin, reflecting both original deposition and minor post-Bakken uplift and erosion, particularly on the flanks of the basin. The Englewood Formation was deposited marginally to the basin, and it seems likely that the circumferential belt separating the Englewood from the Bakken was tectonically controlled, both during and after Bakken and Englewood deposition. The ancestral Cedar Creek anticline in southeastern Montana merges with this circumferential high at its northern end, but to the south it appears to be a part of a linear positive structure extending at least into northwestern South Dakota.

At the end of Bakken and Englewood deposition, minor erosion, particularly on the shelf areas, produced the flat surface upon which the sediments of interval A were deposited. The only evidence for relief on this surface is in the southeastern part of the basin in South Dakota where Pennsylvanian rocks lie directly on the Precambrian. Sandberg (1961, p. 110–111) postulated that this may represent a small island or monadnock on which Paleozoic rocks older than Pennsylvanian were not deposited.

INTERVAL A

STRATIGRAPHIC NOMENCLATURE OF MADISON GROUP

With the discovery of commercial quantities of oil in Mississippian rocks in the Day field in Manitoba in 1951 and soon thereafter in the Beaver Lodge field on the Nesson anticline in North Dakota, oil-well tests began to penetrate the Mississippian rocks in all parts of the Williston basin. Stratigraphic control of the Mississippian rocks now is sufficient over the basin to set up a fairly accurate physical stratigraphic framework, mainly through the use of mechanical logs. Paleontologic data in the basin are still scarce (Sando, 1960; Brindle, 1960; Ziebarth, 1962) so most of the age assignments are based on correlations from the subsurface to Mississippian outcrops in central Montana, where the rocks are better dated.

The nomenclature of Mississippian rocks in the Williston basin subsurface was borrowed, in the early stages of petroleum exploration, from outcrops in central Montana. The Madison Group (Peale, 1893) was recognized and subdivided into the Lodgepole and Mission Canyon Limestones (Collier and Cathcart, 1922). A third formation, the Charles Formation (Seager, 1942), was recognized in the subsurface and was placed in the Madison Group (Perry and Sloss, 1943).

However, in 1954 a different type of nomenclature replaced the use of the Madison Group in the Williston basin. By using thin shale marker-bed correlations, the rocks were divided in approximate equivalents (provincial series) (Thomas, 1954; Fuller, 1956; Harrison and Flood, 1956; Fish and Kinard, 1959; North Dakota Geological Society, 1959; McCabe, 1959; Fuzesy, 1960; Ballard, 1963). The nomenclatural justification and concept of genesis of these marker-bed units is given by Cumming, Fuller, and Porter (1959), who described them as para-time-rock units. These marker-bed units in different parts have been called beds, intervals, or formats, and have been given different geographic or alphabetical–numerical names. The correlation of the marker-bed units allowed the unraveling of the intertonguing relations between the Lodgepole and Mission Canyon Limestones and between the Mission Canyon Limestone and the Charles Formation. When it was realized that the Lodgepole, Mission Canyon, and Charles were not blanket formations representing the same time interval over the basin, they were dropped as a part of the nomenclature. A facies nomenclature was preferred by some (Saskatchewan Geological Society, Mississippian Names and Correlations Committee, 1956; Cumming and others, 1959; Smith, 1960; Ballard, 1963), who then substituted marker-bed terminology. Others (McCabe, 1959) adapted both systems, using the Lodgepole, Mission Canyon, and Charles as formations (in the classical sense as did Powell (1882, p. XLV–XLVI) and Walcott (1903, p. 23) and the marker-bed units as a separate, superimposed nomenclature.

In the present compilation, both types of nomenclature are used. Intervals A, B, and C are essentially the marker-bed units that most closely approximate the Kinderhook, Osage, and Meramec Provincial Series, respectively, as they are known from the preliminary paleontologic studies in outcrops of central Montana (W. J. Sando and J. T. Dutro, Jr., written commun., 1962). The Lodgepole and Mission Canyon Formations, as defined at their type sections, and the Charles Formation, as defined in its type well, are intertonguing lithologic units. The relations between intervals A, B, and C, the marker-bed units, and the formations in the Williston basin are shown in figure 54 and associated table 4, figure 55, and the cross sections on plate 9-F (secs. *O–O'*, *P–P'*, *Q–Q'*).

AGE OF INTERVALS

A discrepancy in ages of the para-time-rock units arises when age assignments of Sando and Dutro in central Montana are compared with Brindle's (1960) age assignments in southeastern Saskatchewan by means of physical correlations based on marker beds (fig. 54). The stratigraphic horizons in central Montana picked by Sando and Dutro as the Kinderhook-Osage and Osage-Meramec boundaries correlate on a physical stratigraphic basis lower in the section than the horizons in southeastern Saskatchewan picked by Brindle as these boundaries. Moreover, the same faunal zones can be recognized in both areas, and Sando and Dutro are in essential agreement with the positions of series boundaries recognized by Brindle in the Saskatchewan sequence except that they would place the Kinderhook-Osage boundary at or near the base of the *Tilston beds* instead of at the top of the *Tilston*. There is no clear solution to this dilemma. A part of the discrepancy disappears if the fossil collections themselves are plotted and compared instead of arbitrarily making the series boundaries coincide with lithologic boundaries between either marker-bed or lithostratigraphic units. Whether such adjustments will account for all the discrepancies is questionable. The final solution of the problem must await further studies, but for the purposes of this study the age designations of Sando and Dutro are accepted and the physical stratigraphic correlations shown in figure 54 and on plate 9-F are presumed correct.

FORMATIONS INCLUDED IN INTERVAL A

Interval A includes the lower beds of the Lodgepole Limestone that are equivalent to the Paine Member of

TABLE 4 — *Drill hole identification and (or) source of logs shown in figure 54, with total depth to top of Madison Group*

[Abbreviations are defined in chapter W. †, indicates a locality number listed in chapter W and shown on the plates]

Drill hole identification and (or) source	Depth to top of Madison Group (feet)
Montana	
No. 302†	0
No. 303†	150
No. 326†	2,130
No. 330†	327
No. 312†	0
No. 471†	0
No. 472†	3,664
No. 373†	3,347
No. 381†	3,163
Amerada 1 Lester Pestol, 17–18N–24E, Fergus County. AmStrat., 1963	4,525
No. 36†	2,864
No. 25†	2,522
Amerada-Kootenai 1, 18–16N–26E, Petroleum County. Northwest Geol. Serv.	2,615
Cobb-Northern Pacific 1, 27–25N–30E, Garfield County	3,495
No. 26†	5,994
Amerada 1 NPRR "B", 1–14N–32E, Garfield County, AmStrat., 1963	5,900
Murphy Corp. 1 NPRR, 23–15N–34E, Garfield County. AmStrat., 1962	6,117
No. 30†	5,874
No. 27†	5,768
Murphy Corp. 1 Govt., 22–15N–37E, Garfield County. AmStrat., 1962	5,833
General Petroleum 5–25–P, 25–16N–38E, Garfield County. AmStrat., 1963	5,786
Shell Oil Williamson 14–34, 34–16N–42E, Garfield County. Northwest Geol. Serv.	7,010
No. 39†	7,015
Shell 44–2 Fraday, 2–18N–43E, Garfield County. AmStrat., 1963	6,747
Stanolind Oil and Gas–Amerada Petroleum NPRR 1, 5–20N–45E, McCone County. Northwest Geol. Serv.	6,195
No. 46†	6,815
Shell 32–33 "B" NP S. W. Richey, 33–22n–48E, McCone County. AmStrat., 1962	6,845
Shell 12–29 "C" NP Richey, 29–22N–49E, McCone County. AmStrat., 1962	6,867
Shell Oil 11–3 "C" NP Richey, 3–22N–49E, McCone County. AmStrat., 1962	6,885
Oceanic Oil 1 Casterline, 4–22N–50E, Dawson County. AmStrat., 1962	6,795
Texota 2 Arthur E. Schock, 26–23N–49E, McCone County. AmStrat., 1963	6,764
Regent Drlg. 1 Art Schock, 22–23N–49E, McCone County. AmStrat., 1963	6,704
No. 54†	6,662
Shell Oil and others 23–9 Govt., 9–24N–48E, McCone County. AmStrat., 1963	6,354
Pure Oil 1 Leuenberger, 30–24N–49E, McCone County, AmStrat., 1961	6,030
Ohio Oil 1 G. L. Jacobs, 27–25N–48E, McCone County, AmStrat., 1963	6,125
Marigold-Farham 1, 14–25N–48E, McCone County. Northwest Geol. Serv.	6,146
No. 61†	5,940
Shell-Etzell 22–29, 29–26N–51E, Richland County. AmStrat., 1952	6,085
D. E. L. Byers 1 Jerome, 8–27N–51E, Roosevelt County. AmStrat., 1963	5,315
Carter 1 R. W. Lowe, 13–28N–50E, Roosevelt County. AmStrat., 1963	5,380
Murphy, C. H., Jr., East Poplar Unit 1, 2–28N–51E, Roosevelt County. Northwest Geol. Serv.	5,250
Murphy-Sohio-D'Orsey 1 Tribal, 15–29N–51E, Roosevelt County. AmStrat., 1963	5,474
Ajax 1 McGowan, 10–29N–50E, Roosevelt County, AmStrat., 1963	5,765
Richfield 1 Stanolind-Govt., 21–30N–51E, Roosevelt County. AmStrat., 1963	6,053
Signal Drlg. 1 Nesbit, 4–31N–51E, Roosevelt County. AmStrat., 1962	6,568
No. 95†	6,807
No. 88†	6,917
Zach Brooks 1 Larson, 2–32N–55E, Sheridan County. AmStrat., 1963	6,675
No. 99†	6,720
Phillips 1–A Brinkman, 14–32N–58E, Sheridan County. AmStrat., 1963	6,965
Calvert Drlg. 1 Siggard Fed., 2–32N–58E, Sheridan County. AmStrat., 1963	6,857
I. J. Wilhite 1 Haaven, 26–34N–58E, Sheridan County. AmStrat., 1963	6,580
I. J. Wilhite 1 Erickson, 24–33N–58E, Sheridan County. AmStrat., 1961	6,802
North Dakota	
Hunt 1 Clifford Price, 29–159N–103W, Williams County. AmStrat., 1963	6,981
No. 178†	6,638
Skelly 1 O. M. Olsen, 33–162N–100W, Divide County. AmStrat., 1962	6,796
Signal Drlg. 1 Lund, 15–161N–100W, Divide County. AmStrat., 1963	6,857
No. 177†	6,116
Signal 1 Knute Unhjem, 5–162N–98W, Divide County. AmStrat., 1963	6,538
Signal Drlg. 1 Joyce, 13–161N–98W, Divide County. AmStrat., 1962	6,744
Signal Drlg. & Explor. 1 Elmar, 11–162N–96W, Divide County. AmStrat., 1962	6,224
Phillips-Ballard 1 P. Braathen, 29–162N–95W, Divide County. AmStrat., 1962	6,490
Kerr-McGee 1 Wifstad Unit, 13–163N–95W, Divide County. AmStrat., 1962	5,790
Calvert-Kelsch and Donlin 1 Anderson, 21–162N–94W, Burke County. AmStrat., 1961	6,332
Luling O & G 1 Amos Peterson, 15–163N–94W, Burke County. AmStrat., 1963	5,812
Texota Oil 1 Dward Ely, 32–163N–93W, Burke County. AmStrat., 1963	5,817
Texota Oil 1 Sarah Swenson, 22–163N–93W, Burke County. AmStrat., 1963	5,730
Northwest Oil 1 Durick Bros., 31–164N–93W, Burke County. AmStrat., 1963	5,552
Saskatchewan	
Fuzesy (1960, pl. 6, well 32)	5,132
Fuzesy (1960, pl. 6, well 33)	4,562
Fuzesy (1960, pl. 6, well 34)	4,650
Fuzesy (1960, pl. 6, well 35)	4,243
Fuzesy (1960, pl. 6, well 36)	4,008
Fuzesy (1960, pl. 6, well 37)	3,853
Fuzesy (1960, pl. 6, well 38)	3,904

the Lodgepole in central Montana (pl. 15) and to the lower part of the *Bottineau interval* in eastern North Dakota (Ballard, 1963, p. 19–24) or to the lower part of the *Souris Valley beds* in Saskatchewan (fig. 55; Fuzesy, 1960). The rocks of this interval are not markedly different from the overlying rocks of the Lodgepole but were differentiated because they appear best to represent the Kinderhook Series as determined in outcrop in central Montana.

In the Mississippian outcrops on the flanks of the Black Hills and in the adjacent subsurface, interval A is made up of the lower part of the Pahasapa Limestone, a unit largely composed of dolomite.

UPPER BOUNDARY OF INTERVAL A

The upper boundary of interval A is placed at the Paine-Woodhurst contact in central Montana and is traced by mechanical logs into the subsurface of the Williston basin. This boundary is not clear cut and is traced with difficulty, largely on shale kicks on gamma-ray logs. In many areas purely arbitrary decisions as to the correlation had to be made. W. J. Sando (oral commun., 1962) placed the Kinderhook-Osage boundary at the base of his C1 coral zone for the purposes of this study, and in three wells in the Montana portion of the Williston basin his series boundary roughly corresponds to the interval boundary picked by the mechanical log correlation, giving some confidence to the correlations.

THICKNESS TRENDS

The rocks of interval A show a thickness configuration (pl. 3-A) that is related crudely to the Williston basin. Around the edges of the basin the interval is more than 100 feet thick, whereas in most of the central parts of the basin it thins to less than 100 feet. This relation was also reported by H. R. McCabe (1959, fig. 11) for the lower Lodgepole Limestone of Manitoba. To the east the interval is beveled at the post-Mississippian unconformity; so, the interval is absent in eastern North Dakota and eastern and southern South Dakota. West of the Williston basin in central Montana, the interval thickens to more than 200 feet. These thickness changes are rather irregular and in part probably are due to minor miscorrelations resulting from the arbitrary choice of the contact of interval A with interval B. However, the regional thickness trends probably represent depositional differences.

LITHOFACIES TRENDS

The lithofacies of interval A (pl. 3-B) also have a concentric pattern related to the configuration of the Williston basin. In the center of the basin in western

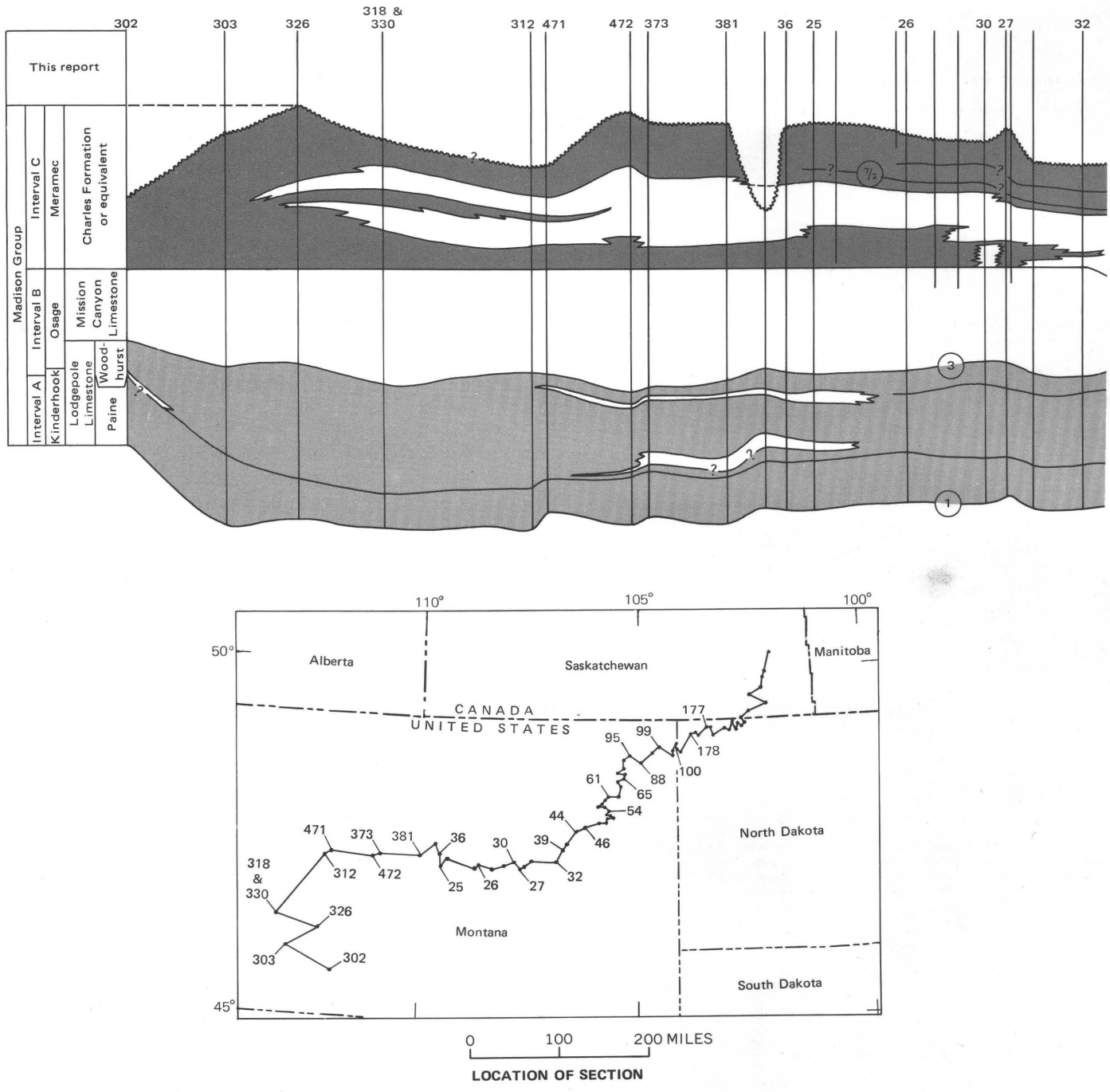
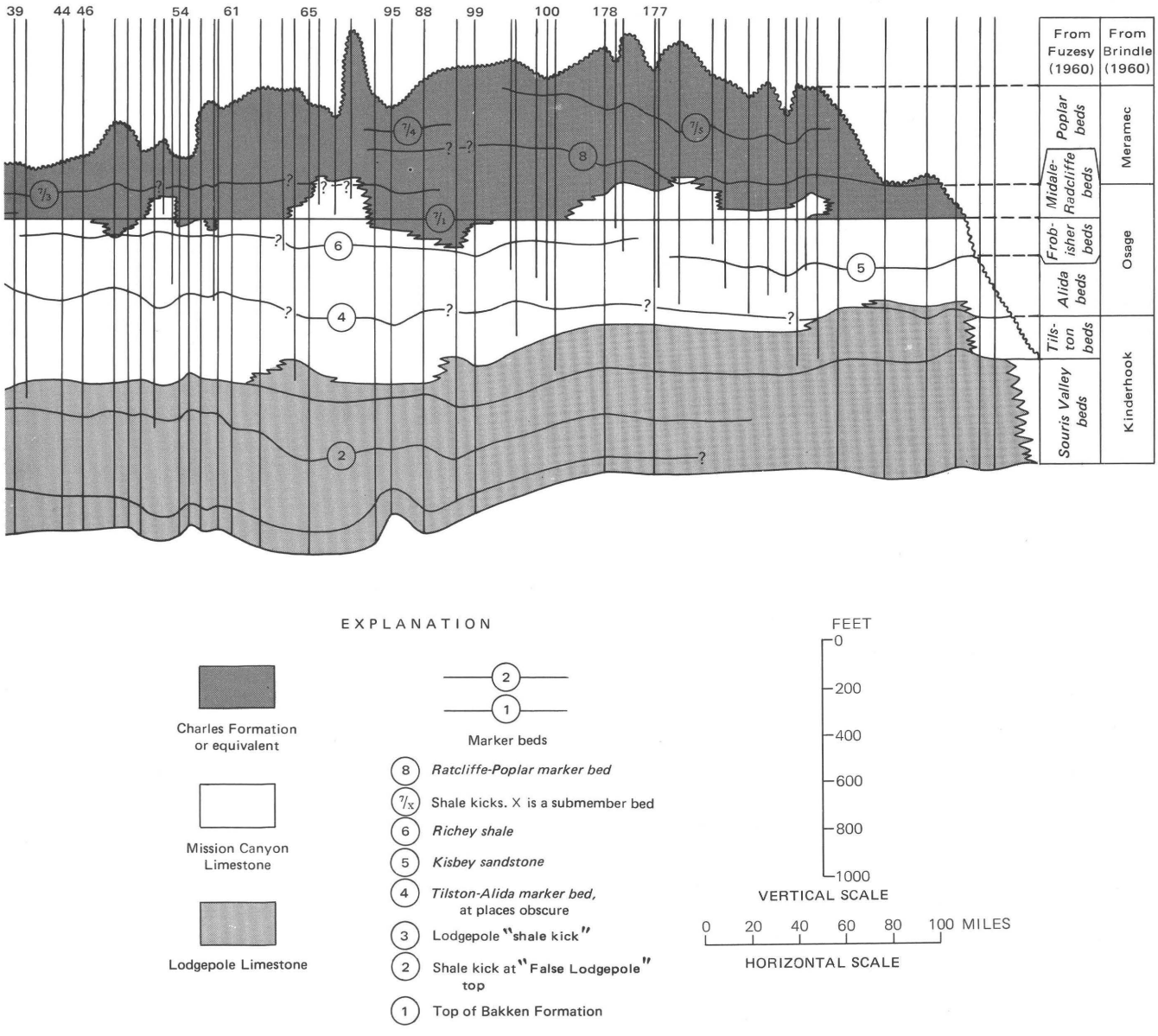


FIGURE 54. — Detailed correlation of beds of the Madison Group between central Montana and southeastern Saskatchewan.

North Dakota, the interval is made up of calcareous shale. Surrounding this facies is a shaly limestone facies that covers most of the rest of the basin. On the shelf areas outside the basin, the interval is made up of relatively pure carbonate rock. The lithofacies and thickness of the interval roughly are related in that the more shaly facies are thinner.

The dolomite percentage of the carbonate rock varies in a regular way also (fig. 56). In the central parts of the basin the carbonate is nearly pure limestone, whereas toward the edges of the basin and, particularly, on the shelf areas to the south of the basin the dolomite percentage increases. Thus, in northwestern South Dakota the carbonate rock in the inter-



(See table 4 for localities.) The shale marker beds are correlated from recorded deflections (shale kicks) in gamma-ray logs.

val contains 10–50 percent dolomite, and in the vicinity of the Black Hills it is predominantly dolomite.

In northwestern South Dakota, where the carbonate becomes dolomitic, some of the carbonate becomes oolitic. The total thickness of oolitic carbonate rock in the interval is shown in the isopach map of figure 57. The lenticular configuration of the oolitic carbonate rock, when visualized in cross section, is suggestive of oolite banks similar to those of the present-day Bahama Banks. However, to further substantiate this suggestion, it would be necessary to show that single oolite bodies are also lenses.

The total thicknesses of chert and nondetrital silica in the interval are shown in figure 58. The silica is ab-

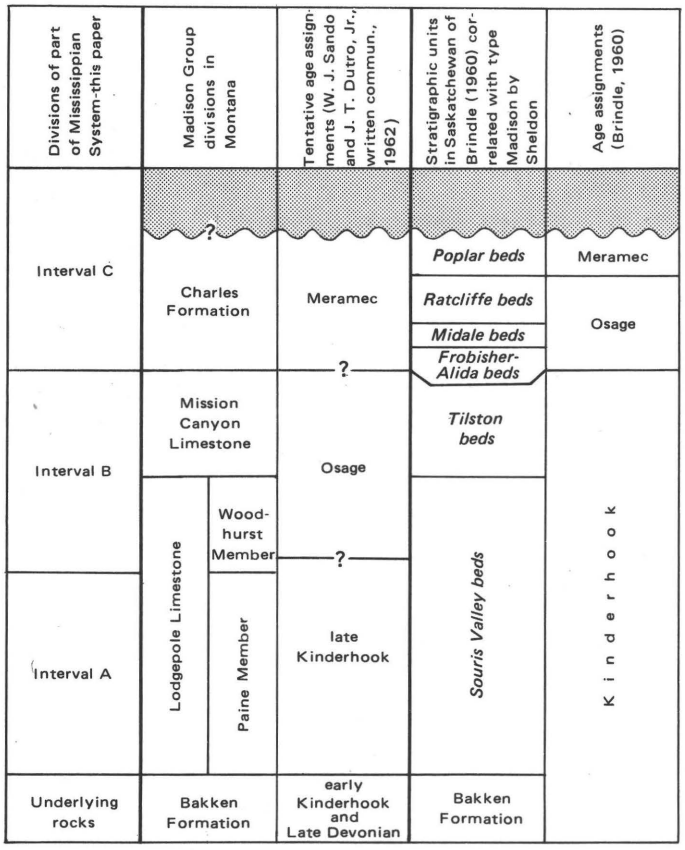


FIGURE 55. — Comparison of age assignments of correlative pre-Chester Mississippian strata in Montana and Saskatchewan.

sent in the center of the basin and on the shelf area outside the basin; it occurs as a basin-flank sediment. Thus, the maximum accumulations of chert and silica occur more basinward than the maximum dolomite and carbonate oolite accumulations.

SEDIMENTATION

In interval A time the Williston basin area can best be interpreted as the site of starved basin deposition under normal marine conditions (pl. 11, fig. 1). A mode of sedimentation in interval A is shown in figure 59. If the physical stratigraphic correlations are correct, it would mean that very little sediment was deposited in the central parts of the basin, and subsidence must have exceeded deposition. Thus, one can infer that the water was relatively deep in comparison to later stages of sedimentation.

In the central parts of the basin, silt, clay, and organic matter were the main sediments deposited. On the flanks of the basin, probably in shallower water, silica was also deposited. MacDonald (1956) and Edie (1958) reported sponge spicules from the cherty rocks of the Lodgepole Limestone; thus, the silica was formed at least in part from siliceous sponge spicules, and its distribution was dependent on the paleoecology of the sponges. It seems plausible that a reducing environment of the deeper water was not favorable for sponge colonies; whereas an oxidizing environment on the better aerated flanks of the basin was favorable. The facies relation between deeper water organic-rich shale and shelf spiculite has been reported in other areas, notably the Permian of the middle Rocky Mountains (Cressman, 1955, p. 25; Sheldon, 1963, p. 84; Cressman and Swanson, 1964, p. 369).

Shelfward from the cherty limestone and shale, carbonate rock was deposited with a greater rate of sedimentation, judging from the greater thickness of the more carbonate-rich facies. Probably, most of the carbonate was deposited as thin beds of biogenic carbonate grains, but some of it was deposited as reefoid or biohermal bodies (Michelson, 1956; Andrichuck, 1958; McCabe, 1959; Lewis, 1960). On the shelf and basin flank, calcitic pellets or oolites were deposited. Also on the basin flanks, dolomite was deposited either as a primary sediment or as a replacement of calcite, owing to increased salinity and the Mg/Ca ratio of the water. It seems likely that the carbonate deposition was in response to a rising depositional pH and salinity in shoaling water. Biogenic calcitic sand was deposited in the basin and on its edges, calcitic pellets or oolities were deposited in banks on or near the strand at the basin edge, and either primary dolomite mud or secondary dolomite was deposited landward of the calcitic pellets or oolites.

PALEOTECTONIC IMPLICATIONS

The sedimentation of interval A was generally related to the paleotectonics (pl. 10, fig. 1). The basin was subsiding relative to the shelf areas and was starved of clastics, so the effect of the subsidence was to deepen the water and cause stagnation. Thus, the basinal sediments are thin. Lack of aeration of the deeper parts of the basin and the sedimentation of organic matter caused a reducing environment. This stage of the development of the Williston basin can be compared with Pettijohn's (1957, p. 636) euxinic stage of geosynclinal development. In a sense, it was a continuation of the Late Devonian–Early Mississippian stage of development represented by the Bakken and equivalent formations, even though the two units may be separated by a disconformity. On the flanks of the basin and the adjacent shelf, the carbonate deposits, even though thicker than the basinal sediments, are relatively thin and show that the shelf was tectonically only slightly negative during interval A time. The area surrounding the Williston basin and adjoining shelf was topographically low (pl. 12, fig. 1), as indicated by the small quantity of clastic material supplied to the basin.

Thus, tectonism in the area was mild during interval

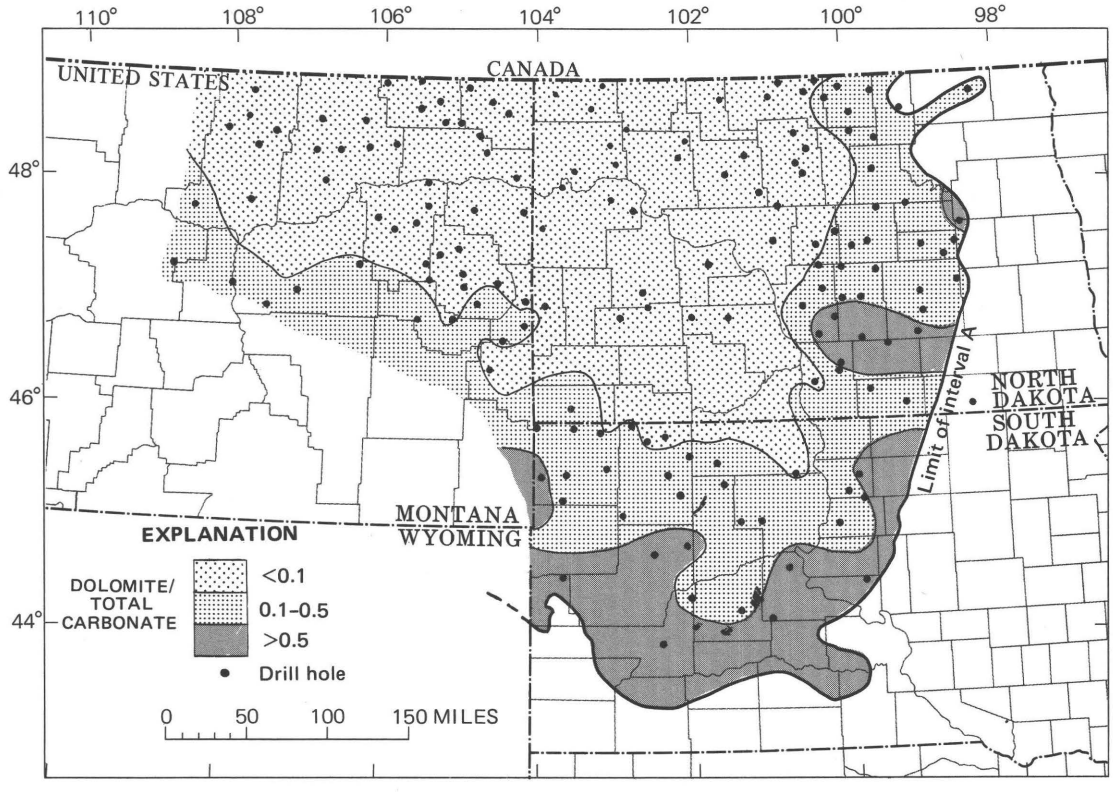


FIGURE 56. — Dolomite to total carbonate ratio distribution in interval A in the Williston basin.

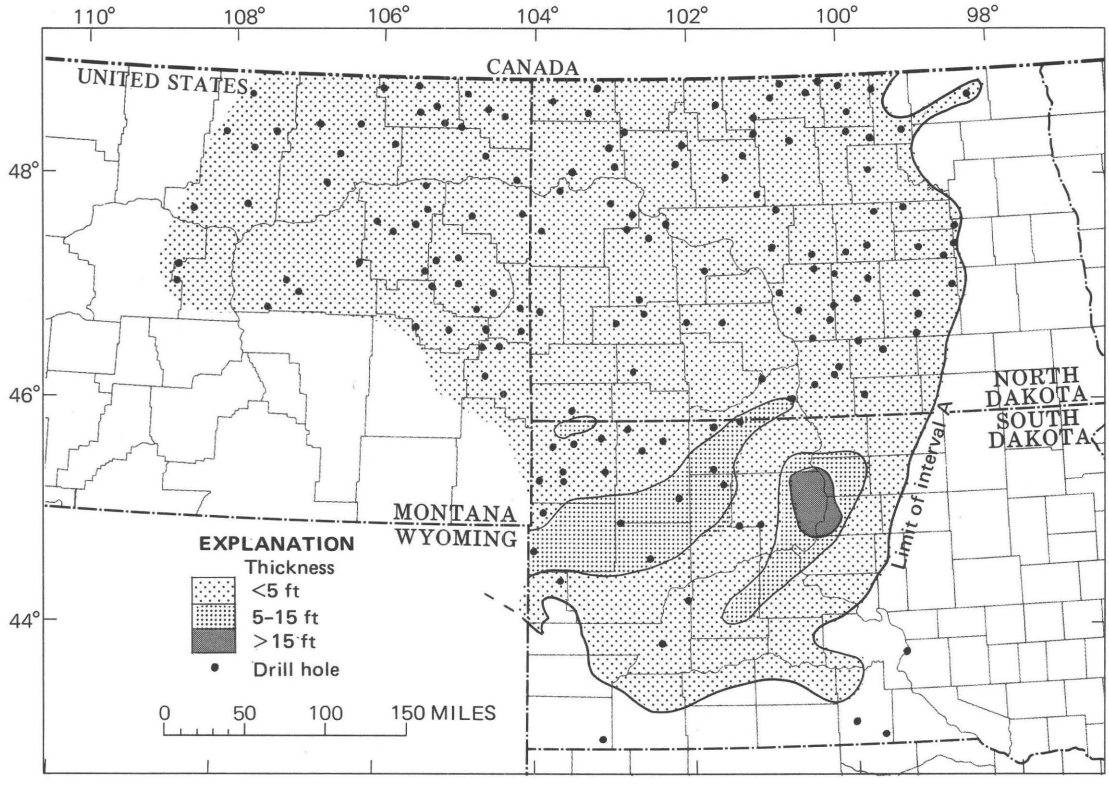


FIGURE 57. — Cumulative thickness of carbonate oolite beds in interval A in the Williston basin.

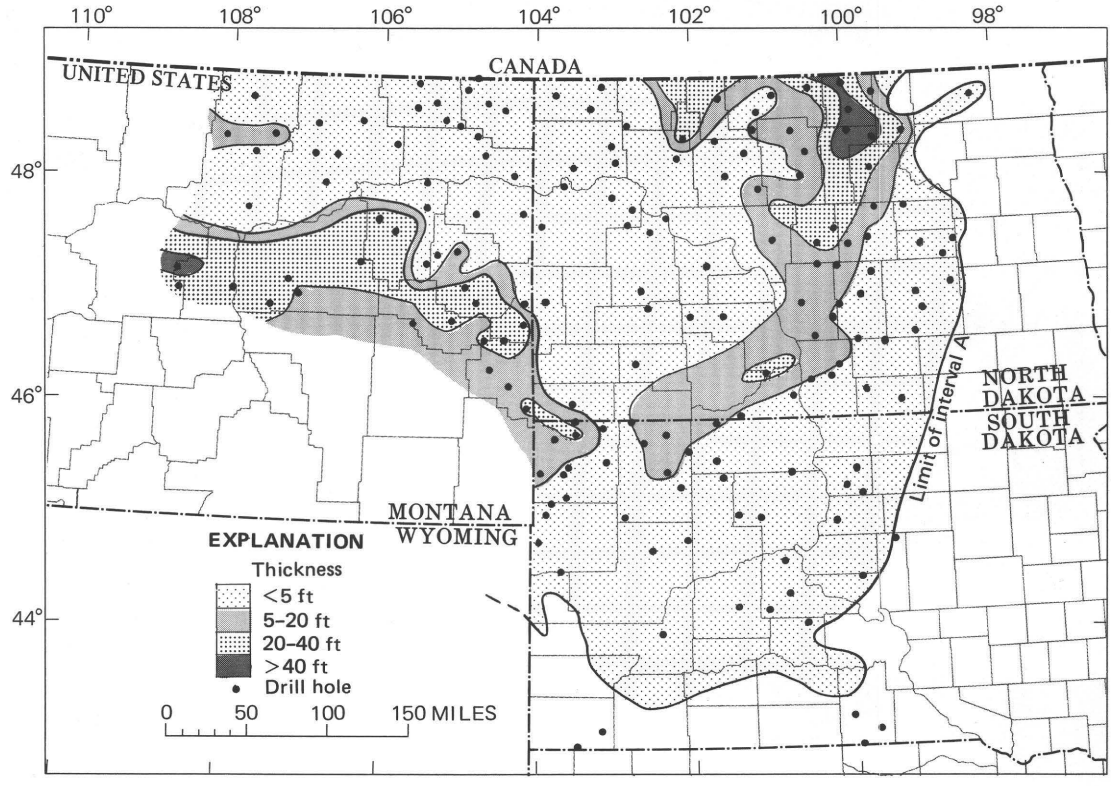


FIGURE 58. — Cumulative thickness of chert and silica in interval A in the Williston basin.

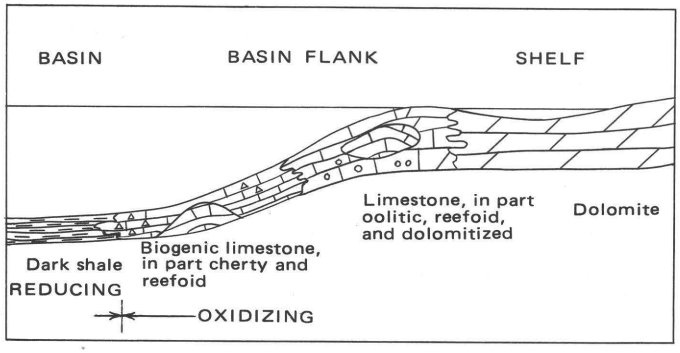


FIGURE 59. — Model of sedimentation in interval A time.

A time. The ancestral Williston basin was subsiding a little more than the adjoining shelf areas, but the total subsidence was small.

INTERVAL B

FORMATIONS INCLUDED

The rocks of interval B are included wholly within the Madison Group. The general treatment of the stratigraphic nomenclature of the Madison Group in the Williston basin was given in the section on interval A and will not be further treated here.

Interval B includes all of the Lodgepole Limestone not in interval A and includes the lower part of the Mission Canyon Limestone (pl. 15, cols. 98–101; pl 9-F; secs. *O–O'*, *Q–Q'*). Within the Mission Canyon Limestone assigned to interval B is a bed of anhydrite (fig. 60) as much as 70 feet thick. This bed joins with the main body of the Charles in central North Dakota (pl. 9-F, sec. *O–O'*); and therefore, it seems more properly regarded as a tongue of the Charles Formation than a member of the Mission Canyon. Interval B includes this tongue of the Charles Formation.

In the outcrops of Mississippian age in the northern Black Hills of South Dakota, interval B is made up of the upper part of the Pahasapa Limestone. The Pahasapa is largely dolomite, and the contact between intervals A and B is poorly defined and is determined somewhat arbitrarily.

STRATIGRAPHIC RELATIONS

Interval B is made up of parts of three intertonguing formations, the Lodgepole, Mission Canyon, and the Charles. Generally, the Lodgepole occurs at the base, the Mission Canyon in the middle, and the Charles at the top. The Lodgepole is a dark siliceous shaley limestone, which is a basinal facies. The Mission Canyon is a light-colored bioclastic or oolitic limestone and

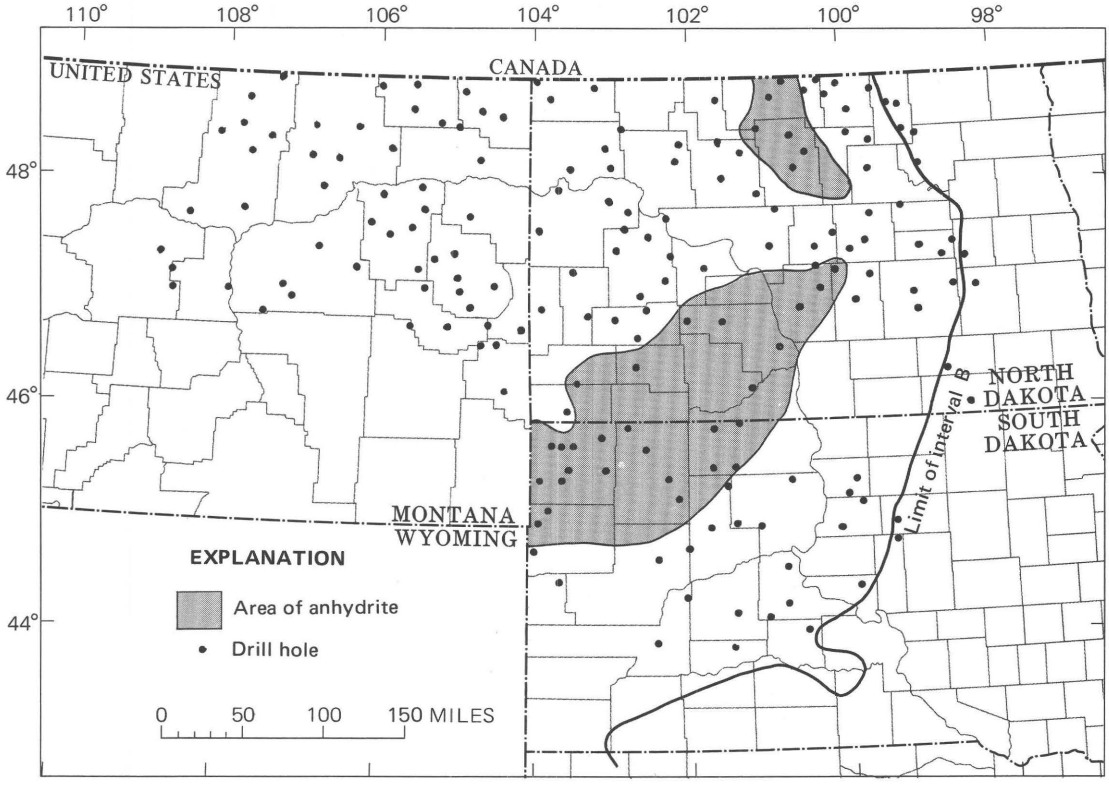


FIGURE 60. — Distribution of anhydrite bed of interval B in the Williston basin.

dolomite, which is a shallower water facies of the Lodgepole; and the tongue of the Charles is an evaporitic facies containing minor red shales, which in interval B time was a shelf facies of the upper part of the Mission Canyon. Their stratigraphic relations are shown in the cross sections of plate 9-F and in figure 54. The upper beds of the Lodgepole grade shelfward into Mission Canyon, and the upper beds of the Mission Canyon grade shelfward into Charles.

Intercalated in these formations are widespread thin shale beds that have been used as marker beds, which are para-time-rock units of Cumming, Fuller, and Porter (1959) and are nearly in accord with the time-stratigraphic unit boundaries. These marker beds are not basinwide in their extent, and beds that are prominent in the northeastern Williston basin in part are difficult to trace to other parts of the basin (fig. 54). Thus, even though the marker beds help prove the intertonguing relations of the Lodgepole, Mission Canyon, and Charles Formations, they can be used only with difficulty in establishing basinwide correlations with the time-stratigraphic units.

UPPER BOUNDARY OF INTERVAL B

Roberts (chap. N, p. 233) has placed the top of interval B in central Montana at the base of a widespread anhydrite bed (fig. 61) that correlates with the anhydrite bed lying at the base of the Charles Formation at the type well (Seager, 1942). This anhydrite occurs on the southern and eastern flanks of the basin (pl. 9-F, secs. *O–O'*, *Q–Q'*) and is equivalent to the *Bottineau evaporite* (Fish and Kinard, 1959, p. 127) on the eastern side of the basin. In the central and northwestern parts of the basin this anhydrite is absent, and the top of interval B falls within the Mission Canyon Limestone. This horizon has been traced with difficulty throughout this area. Around the edges of the basin, interval B is truncated by Pennsylvanian, Jurassic, or Cretaceous rocks, and the top of interval B in these areas lies at the post-Mississippian unconformity.

THICKNESS TRENDS

The isopach map of interval B is relatively simple (pl. 4-A). The interval reaches 1,200 feet in thickness in the central part of the basin in western North Dakota and thins fairly regularly shelfward. The Central Montana trough which extends westward from the center of the basin across central Montana contains interval B rocks more than 800 feet thick in comparison to the thinner rocks on the shelves to the north and south of the trough. In eastern North and South Dakota the interval is beveled beneath the post-Mississippian unconformity, and the interval thins and pinches out eastward

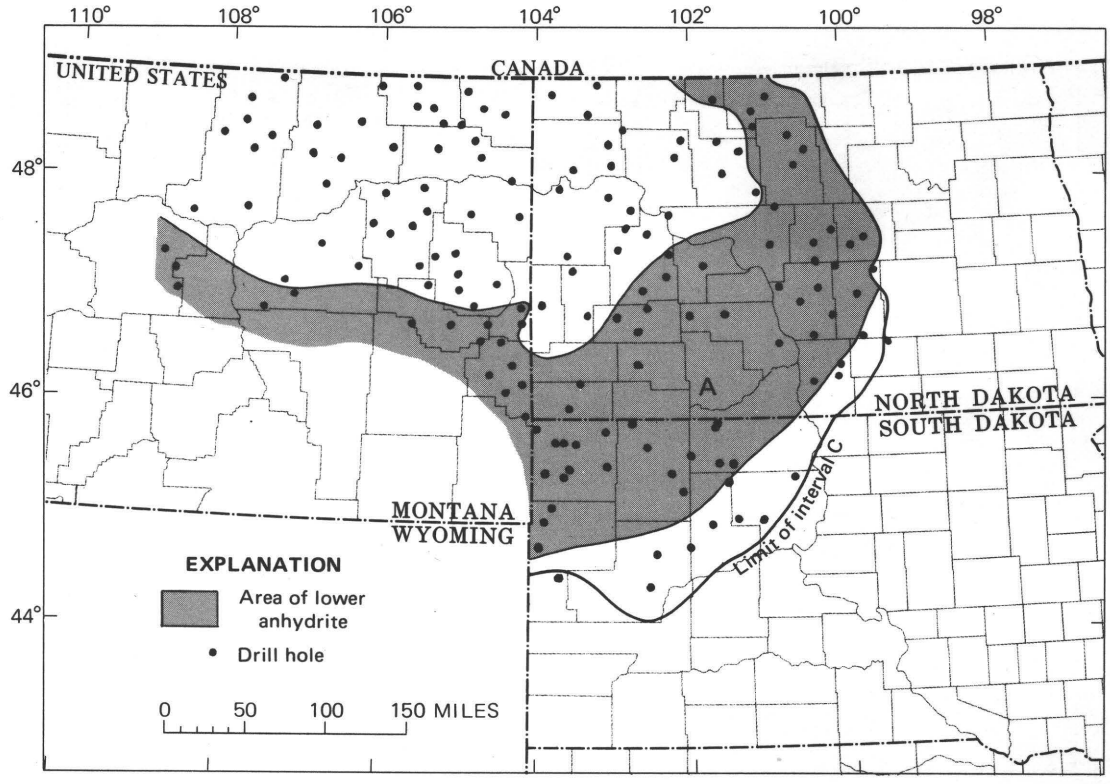


FIGURE 61. — Distribution of anhydrite bed at base of interval C in the Williston basin.

and southeastward. In central South Dakota the interval thins markedly around and is absent in a small area near the regional beveled edge of the interval. This area was probably a small island in the Mississippian Sea (Sandberg, 1961, p. 110–111) where the interval was not deposited.

Even though the beveling of the interval on the edges of the basin and on the surrounding shelf is due to post-Mississippian erosion, in the central part of the basin interval B is conformably overlain by interval C, and the shelfward thinning in the central part of the basin is sedimentational, not erosional.

LITHOFACIES TRENDS

The lithofacies trends of interval B (pl. 4-B) correlate fairly well with the thickness trends. In the central parts of the basin, where the interval is thick, it is shaly limestone, but on the flanks of the basin and the shelf areas it is relatively pure carbonate rock.

On the shelf areas, the percentage of dolomite increases in the interval (fig. 62); near the Black Hills the interval is more than half dolomite and is more than a quarter dolomite on most of the eastern and southern shelf. A small area of 10–30 percent dolomite occurs on the crest of the modern Nesson anticline in the center of the basin.

More or less coincident with the increase of dolomite is the presence of carbonate oolite (fig. 63). The oolite occurs in a roughly circular belt on the flanks of the basin and is absent both in the central parts of the basin and on the shelf areas away from the basin flanks. However, a small area of oolite occurs on the crest of the modern Nesson anticline in the same position as the dolomite mentioned previously.

Silica and chert occur with a similar distribution (fig. 64) but are stratigraphically lower than the oolite. However, the silica is more abundant in the northern part of the area and the oolite, in the southern. The modern Nesson anticline also has an anomalous concentration of silica, as much as 60 feet thick in interval B.

The anhydrite bed discussed previously is marginal to the basin on the eastern and southern sides (fig. 60).

SEDIMENTATION

Sedimentation of interval A continued into interval B time (pl. 11, fig. 2) without break. Interval B is about 10 times thicker than interval A, and if the two were at all comparable in terms of time intervals, then the rate of carbonate deposition markedly increased. The lithic character of the carbonate and the distribution of rock

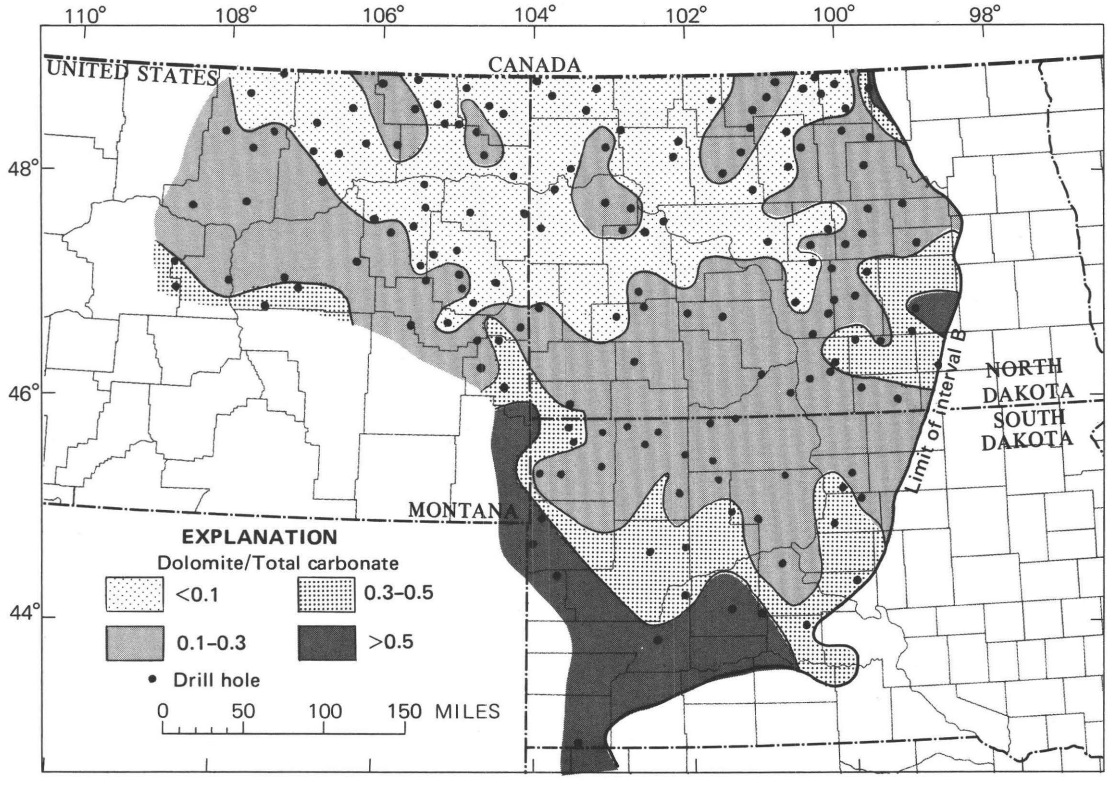


FIGURE 62. — Dolomite to total carbonate ratio distribution in interval B in the Williston basin.

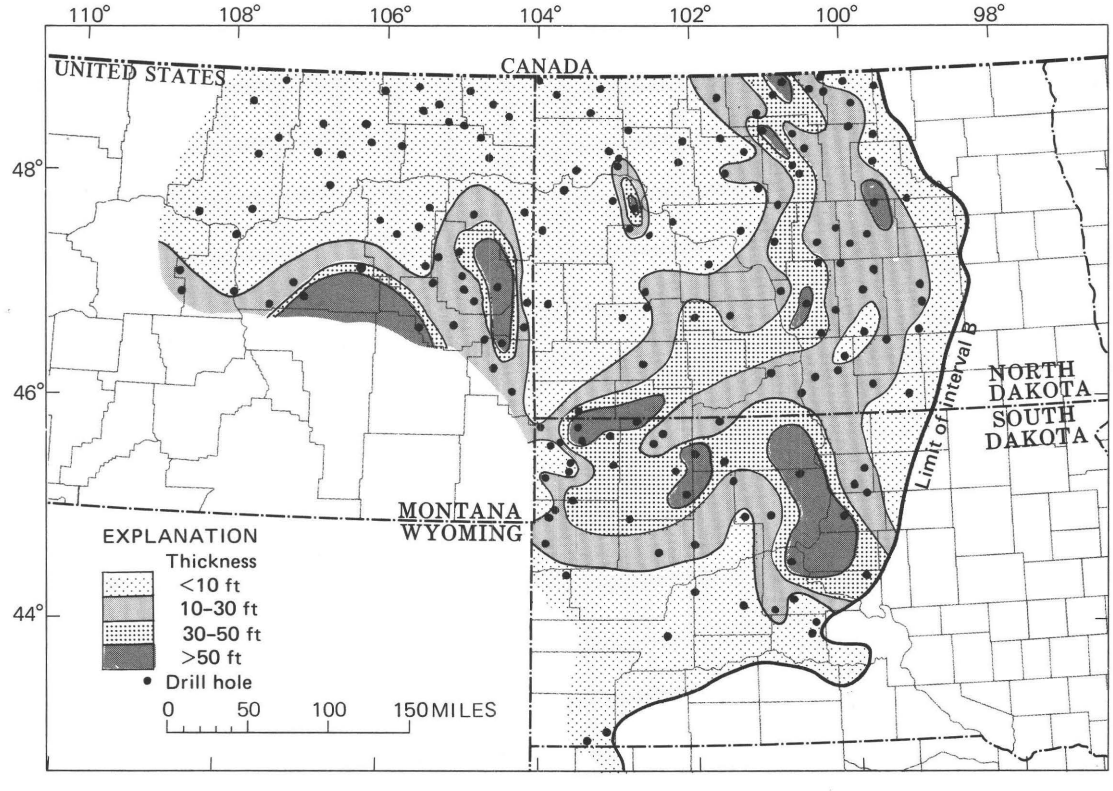


FIGURE 63. — Cumulative thickness of carbonate oolite beds in interval B in the Williston basin.

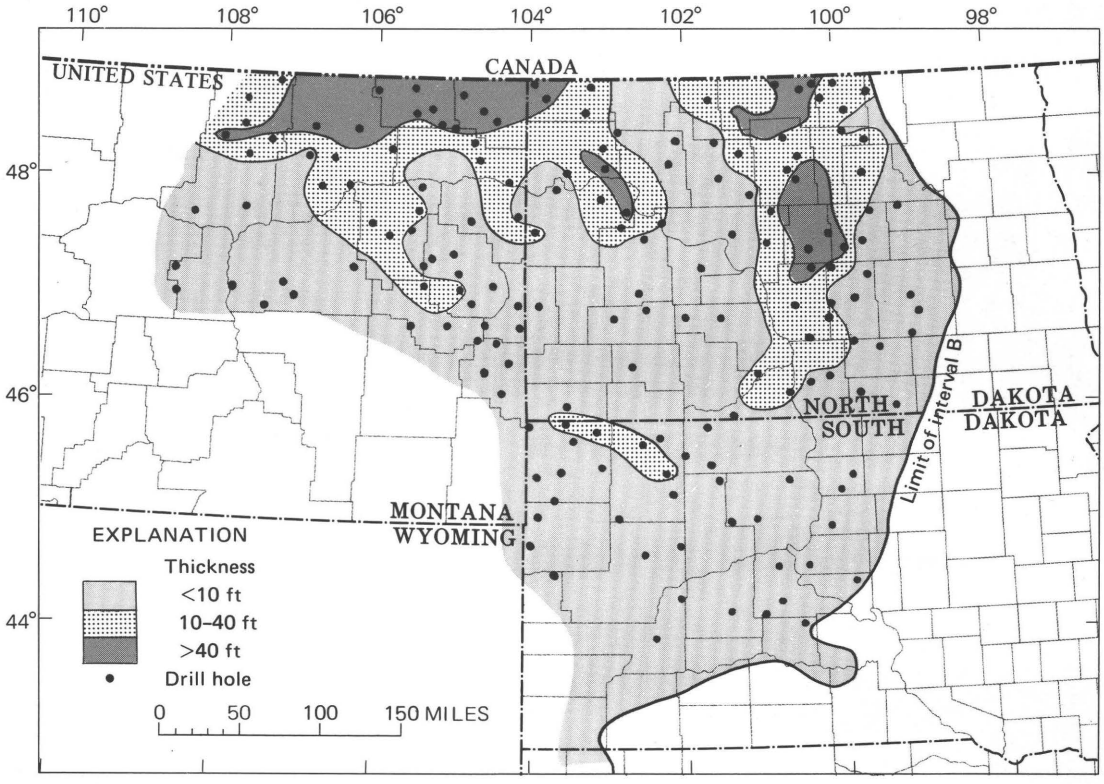


FIGURE 64. — Cumulative thickness of chert and silica in interval B in the Williston basin.

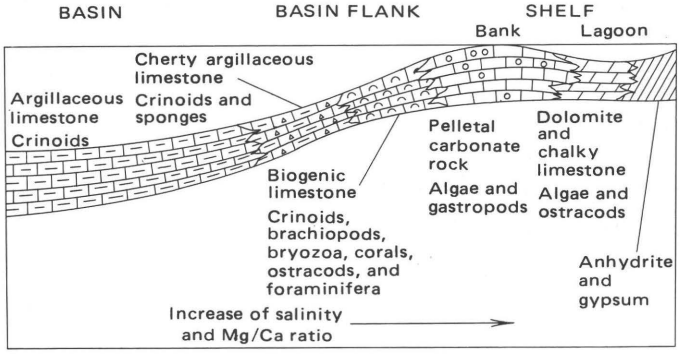


FIGURE 65. — Model of sedimentation in interval B time (from Edie, 1958).

varieties of the two intervals are similar. (Compare figs. 59 and 65.) In the basin, argillaceous and calcitic mud was deposited; whereas on the flanks of the basin cherty argillaceous and calcitic mud and siliceous sponge spicules were deposited. Farther shelfward beds of light-colored bioclastic calcitic sand and banks of calcitic pellets and oolites formed. Calcium sulfate and evaporitic and algal calcitic and dolomitic mud formed in the lagoons behind the banks of the calcitic pellets and oolites.

The sedimentation and environments of deposition were discussed by Edie (1958), who subdivided the environments into the basin, shelf and bank margin, bank, and lagoon. He concluded that sponges and crinoids were most abundant in the basin environment. Crinoids, brachiopods, bryozoa, corals, and ostracods were most abundant in the shelf and bank margin, a conclusion regarding the corals substantiated by Sando (1960) who found the corals most abundant in the marginal facies. The banks contain precipitated limestone and algae and gastropods are the most abundant forms. The chalky limestone of the lagoon contains ostracods.

The increase of dolomite on the shelf was probably caused by increasing salinity. Precipitation of anhydrite or gypsum in the lagoons would increase the Mg/Ca ratio of the brines, which in turn could react with calcium carbonate sediments of the bank to form dolomite (Adams and Rhodes, 1960; Deffeyes and others, 1964).

The rocks of interval B show facies relations at various horizons in a manner illustrated by figure 65. A similar sequence occurs vertically within interval B, showing that the basin was becoming shallower; the shallower water sediments progressively overlie the deeper water sediments. McCabe (1959, p. 58) advanced the concept of a single large cycle of basin development, starting with a deeper water widespread Lodgepole and gradual contracting and shallowing of the sea until a basinal evaporitic restricted sea remained in Charles time. Superimposed on this general regression were

numerous minor oscillations of the environments giving minor intertonguing of the genetic rock units and a vertical cyclic sedimentation.

The shale marker beds of interval B present a different problem of sedimentation. They cross the facies boundaries defined by the intertonguing formational boundaries. Cumming, Fuller, and Porter (1959, p. 730) gave evidence that these shale beds are due to diastrophism and a rapid influx of sediment into the basin, but they pointed out that their origin has not been settled yet. It seems possible that some of the shale beds may mark temporary basinwide cessation of chemical sedimentation, due perhaps to freshening of the water or a lowering of pH, so that fine-grained clastic material which had been entering the basin in small amounts all the time produced a relatively pure shale bed in much the same fashion as during the time of interval A. Despite the unknown origin of these beds, they are time parallel in contrast to the other sedimentary beds, which are time transgressive.

PALEOTECTONIC IMPLICATIONS

The Williston basin in interval B time continued subsiding, allowing the deposition of more than 1,200 feet of carbonate rock (pl. 10, fig. 2). At the outset, the basin probably was fairly deep, owing to being starved of both chemical and clastic sediments. However, as time progressed, chemical sedimentation caught up with subsidence and the basin filled, so shallow water carbonate sediments were being deposited in the center of the basin at the end of interval B time.

The shelf areas were also subsiding during interval B time, although not as much as the basin. Similarly, the seas became shallower on the shelf; thus, at the end of interval B time, shallow-water evaporites were being deposited on the shelf.

The Cedar Creek anticline marked the southwest edge of the basin during interval B time, and, apparently, it acted as a hinge line for basinal subsidence. There appears to be no evidence from thickness relations that it acted as a positive area relative to the shelf area to the southwest, and no anticlinal uplift of any magnitude occurred. However, the increase in dolomite percentage of the carbonate rock (fig. 62) and the presence of carbonate oolite bodies (fig. 63) in the vicinity of the Cedar Creek anticline suggest that uplift on the anticline affected water depth and thereby sedimentary environments. It is possible, though, that this relationship is accidental.

The anomalous presence of increased dolomite, oolite, and chert in an elongate area coinciding with the Nesson anticline suggests that this anticline also may have been active during the deposition of interval B sediments, as proposed by McCabe (1954) and Towse (1957) for the Beaver Lodge anticline, a part of the Nesson anticline.

The lack of appreciable amounts of terrigenous sediments in the Williston basin and surrounding shelf indicates that the craton to the east was low and supplied little material to the area. However, if Cumming, Fuller, and Porter (1959) are correct in their interpretation of the marker beds representing clastic influxes to the basin, perhaps mildly positive movements occurred on the craton to the east.

INTERVAL C

FORMATIONS INCLUDED

The rocks of interval C make up the upper part of the Madison Group (pl. 15, cols. 98–101). The stratigraphic nomenclature of the Madison Group in the Williston basin was included in the discussion of interval A.

Interval C in general comprises the Charles Formation and an underlying upper part of the Mission Canyon Limestone. Only at the edges of the basin does the evaporite facies of the Charles Formation appear in beds older than interval C (fig. 60). The interval includes that part of the Mission Canyon that is equivalent to or younger than the lower Charles evaporite at the type well of the Charles Formation.

Interval C contains widespread marker beds similar to those in interval B. It is impossible to trace these marker beds over all the basin, and the para-time-rock intervals that are useful in the northeastern part of the Williston basin are not identified in the southwestern part of the basin. However, in the eastern part of the basin, interval C comprises the *Frobisher-Alida*, *Ratcliffe*, and *Poplar intervals* as used in North Dakota by Ballard (1963). Ballard (1963, p. 25) treated the *Midale* as a lower subinterval of the *Ratcliffe interval.*

STRATIGRAPHIC RELATIONS

Interval C is made up of two para-time-rock units that have different internal stratigraphic relations. The lower unit comprises the Mission Canyon marine shallow-water limestones in the central part of the basin and the Charles lagoonal evaporites and dolomite on the margins of the basin. These two formations intertongue with the younger Charles evaporite tongues extending generally more basinward than the older ones. The upper unit, which includes all the *Poplar interval* and possibly the upper part of the *Ratcliffe interval*, consists wholly of evaporitic rocks of the Charles Formation, including extensive basinal halite deposits. These relations are illustrated by cross section Q–Q' of plate 9-F.

UPPER BOUNDARY OF INTERVAL C

The upper boundary of interval C is placed at the base of the Kibbey Formation, except around the edges of the basin where interval C is truncated beneath Pennsylvanian or Jurassic rocks. Many geologists have recognized a disconformity between the Madison Group and overlying beds of Late Mississippian or Pennsylvanian age in parts of the Northern Rocky Mountains region of the United States. The time span of the post-Madison hiatus was discussed by Sando, Mamet, and Dutro (1969, p. E20). According to W. J. Sando (oral commun., 1970), fossils younger than early Meramec have not been found in the Madison and fossils older than Chester have not been found in the Big Snowy Group. Although few fossils have been recovered from the Charles Formation and equivalent beds in the Williston basin (Sando, 1960; Brindle, 1960), these represent the same age as those found outside the basin, where faunas are more abundant. According to McCabe (1954), the top of the Charles is a disconformity in the Williston basin. On the other hand, Maughan and Roberts (1967, p. B5, pl. 2) recognized local intertonguing of the Kibbey and Charles in the subsurface of northeastern Montana. They concluded that the Kibbey rests conformably on the Madison throughout much of Montana except at some localities near the margin of the Kibbey deposits. If the contact is conformable and the Kibbey Formation is of Chester age, then a time interval of considerable magnitude representing the middle and late Meramec should be represented in the upper part of the Charles Formation and equivalent beds.

We favor the interpretation that the upper boundary of interval C is a disconformity in the Williston basin. This interpretation differs from that presented in chapter N (p. 241) for the area just to the west, where deposition is regarded as continuous from interval C into interval D time. In the Williston basin a few tens of feet of erosional relief is noted on the upper surface of the Charles Formation (fig. 54). No regional drainage pattern can be detected to account for this relief using the isopach map of the overlying sandstone of the lower part of the Kibbey Formation (Rawson, 1969, fig. 4).

THICKNESS TRENDS

Interval C has a simple thickness distribution (pl. 5-A). The interval is a lens-shaped rock body that is more than 1,300 feet thick in the center of the basin and thins regularly to the east and south; the Central Montana trough extends westward to the Cordilleran geosyncline and within this region of study contains interval C rocks as much as 600 feet thick. The interval thins both to the north and to the south of the trough.

Throughout the central part of the Williston basin, interval C is overlain by the Kibbey Formation of interval D; little of interval C has been lost by post-Mississippian erosion, and the thickness changes reflect largely the original thickness. On the edges of the basin where the interval is overlain by Pennsylvanian rocks and on the northern and eastern parts of the basin, by Jurassic rocks, considerable amounts of the interval have been removed by post-Mississippian erosion.

LITHOFACIES TRENDS

Interval C has generally simple lithofacies trends in the Williston basin (pl. 5-B). The interval is made up largely of chemical sediments. It tends to be shaly and contains the largest amounts of gypsum and halite in the central parts of the basin. These facies coincide with the thickening of the interval in the central parts of the basin, and this relation between thickness and facies is sedimentational. Only along the subcrop of interval C below post-Mississippian rocks (fig. 68) have the original lithofacies trends been influenced by removal of the upper part of the interval during postdepositional erosion.

In detail, the lithofacies trends are fairly complex. As noted previously, the interval is made up of two rock units that behave independently, the lower showing a marginal evaporite facies and the upper a basinal evaporite facies. As a result, the lithofacies maps cannot be simply interpreted, because they are a composite of two superposed patterns. Evaporite rocks of the interval are found mainly in the central part of the basin and in the Central Montana trough, but also local concentrations occur peripherally to the basin. Similarly, shale occurs prominently both in the central parts of the basin and in the trough, as well as peripherally. The peripheral habit of some evaporite is well shown by distribution of the lower Charles evaporite bed of the type well throughout the Williston basin (fig. 61). The basinal habit of other evaporite is well shown by the isopach map of the cumulative thickness of halite in interval C (fig. 66). The dolomite to total carbonate ratio (fig. 67) likewise shows a complex relation. The lower rock unit consists of Mission Canyon calcitic beds in the central parts of the basin interbedded with dolomitic beds of the Charles Formation on the edges of the basin. The upper rock unit is made up entirely of Charles beds, whose carbonates are largely dolomitic. Figure 67 superposes the one distribution pattern on the other, giving a resultant pattern of basinal limestone and peripheral dolomite, but with considerable irregularity.

SEDIMENTATION

Sedimentation of the lower rock unit of interval C

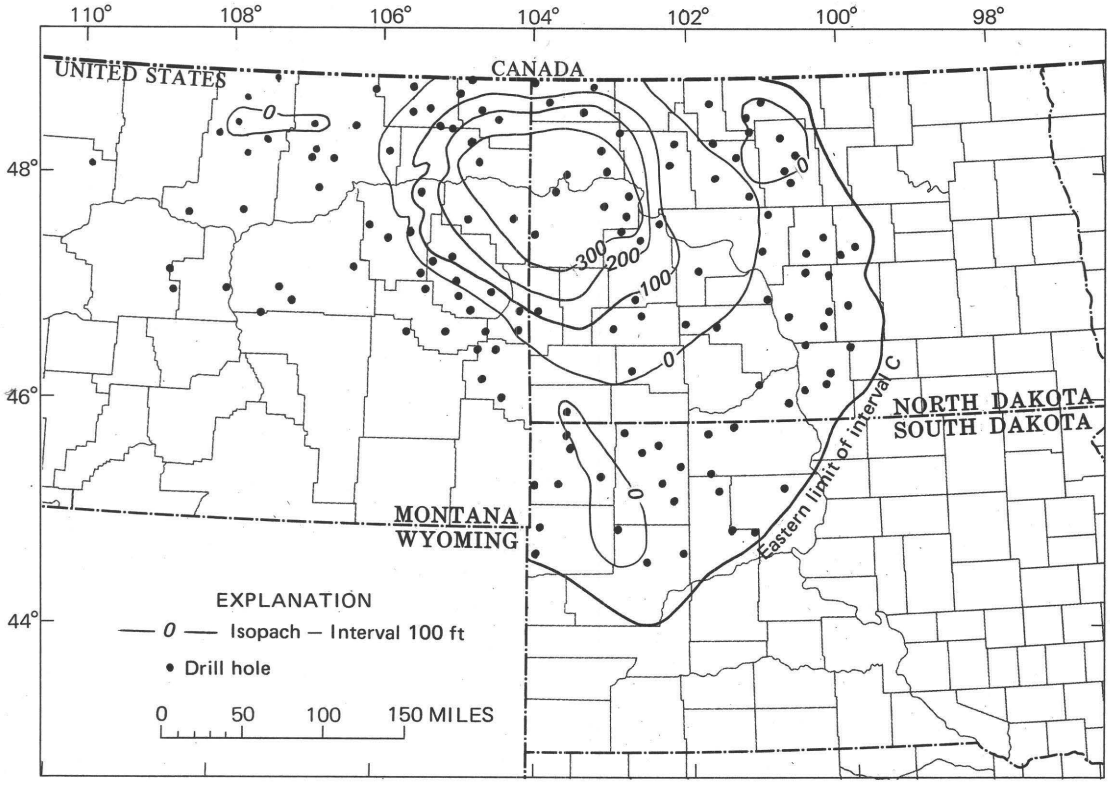


FIGURE 66. — Cumulative thickness of halite in interval C in the Williston basin. In part from Anderson and Hansen (1957, pls. 1, 3) and C. A. Sandberg (written commun., 1962).

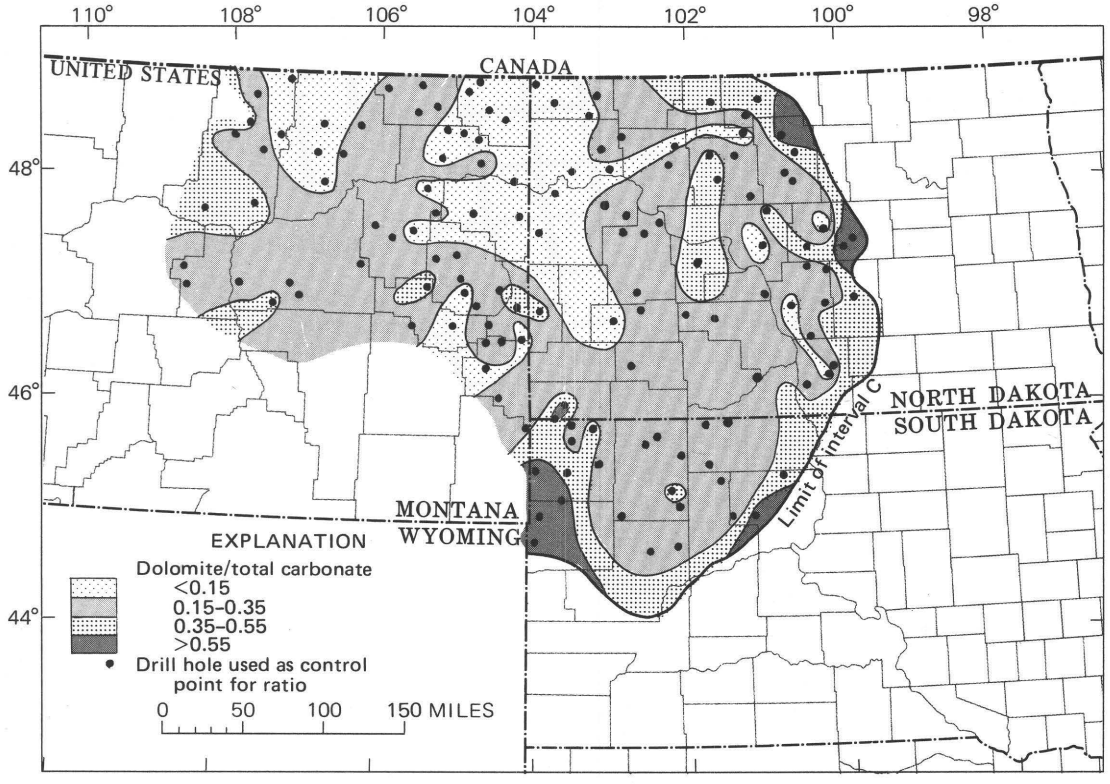


FIGURE 67. — Dolomite to total carbonate ratio distribution in interval C in the Williston basin.

(pl. 11, fig. 3) was similar to the sedimentation of interval B. The sedimentation of the upper rock unit of interval C was quite different. Evaporitic precipitates of calcium sulfate and halite were formed alternately with evaporitic and algal, calcitic, and dolomitic muds. Red muds and sands also were deposited as thin beds in the basin.

The source of the detritus is difficult to pinpoint inasmuch as the beds occur as widespread thin units. There has not been evidence to show that the Siouxia arch had emerged and was shedding sediments in interval C time; however, the detailed work that would be necessary to show mild uplift and erosion has not been carried out. Immediately following interval C deposition, mild uplift of the Siouxia arch was in progress and the time at which this began should be further investigated.

The Williston basin and Central Montana trough were the site of a semirestricted sea which opened westward to the ocean and which toward the end of interval C time, formed a large salt pan in which the evaporites were deposited along with carbonaceous and argillaceous muds.

PALEOTECTONIC IMPLICATIONS

Interval C time represented the closing stages of the single large Mississippian sedimentary cycle during which the Williston basin was filled. The basin remainedly mildly negative until the end of interval C time (pl. 10, fig. 3), although sedimentation was more rapid than subsidence. This excess of sedimentation resulted in the gradual encroachment into the basin of marginal calcitic pellet banks and their shelfward lagoons, until the entire basin became restricted and evaporitic sedimentation completed the filling of the basin. From time to time the waters in the basin returned to more or less normal marine salinities, probably owing to minor subsidence, resulting in cyclic sedimentation.

During deposition of the Charles Formation, renewed intermittent activity along an extended Cedar Creek trend may have temporarily stopped or at least reduced the flow of water into the basin. This positive structural activity periodically produced a highly restricted marine environment in the basin in which halite was precipitated (Sandberg, 1961, p. 63–64).

At the end of interval C time the area of the Williston basin was relatively positive, allowing widespread minor erosion before additional subsidence and deposition of the sediments of interval D.

INTERVAL D

FORMATIONS INCLUDED

Interval D in the Williston basin is made up of the Big Snowy Group, which comprises from base to top the Kibbey, Otter, and Heath Formations (pl. 15). All this nomenclature originates in the outcrop area of central Montana and is discussed by Roberts in chapter N (p. 237). The extrapolation of central Montana surface nomenclature into the subsurface of the Williston basin is relatively easy because the stratigraphic units are widespread and fairly regular in their lithology.

The Kibbey Formation is made up of a few hundred feet of sandstone, shale, and minor carbonate and evaporite. It is easily traced in the subsurface of the basin and contains a distinctive thin marker bed of limestone. This bed has been named the *Ray Member* by Rawson (1968) and is underlain by the lower member of the Kibbey and overlain by the upper member of the Kibbey. The lower member of the Kibbey is mainly shale, and the upper member is mainly sandstone (Rawson, 1968, figs. 4, 7). In central Montana the Otter Formation consists mainly of green shale, whereas the Heath Formation is mainly black limestone and shale. The distinction between the two is based on the color difference. Toward the east in Montana, the distinction becomes more difficult, and in most of North Dakota and South Dakota, the Heath is not recognized.

The Big Snowy Group is Chester in age in the Williston basin. This designation is based wholly on the subsurface physical correlations of the rocks to the outcrops in central Montana, inasmuch as the paleontology of the Big Snowy rocks in the basin has not been studied.

STRATIGRAPHIC RELATIONS

The Kibbey Formation lies disconformably on the Charles Formation. The Otter lies conformably on the Kibbey, and the Heath, where it is recognized, lies conformably on the Otter. The Heath Formation is shown in figure 68 to extend eastward from Montana into North Dakota. It is not known whether the Heath is absent in North Dakota as a result of erosion or whether it changes facies into rocks that are similar to the Otter. This problem requires more study for resolution.

UPPER BOUNDARY OF INTERVAL D

The Big Snowy Group extends over the Williston basin and Central Montana trough and is truncated on the basin and trough edges by post-Mississippian erosion. The group is overlain unconformably in most

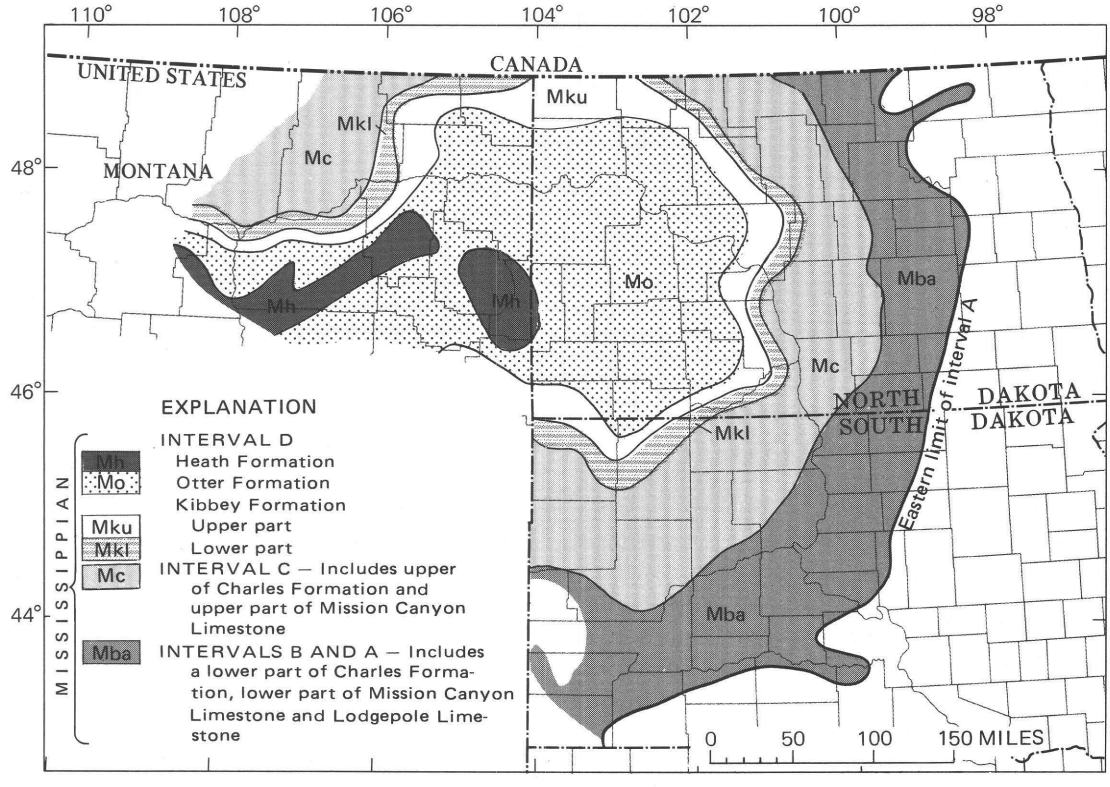


FIGURE 68. — Interval subcrops on the erosional upper surface of the Mississippian System. In part from Maughan and Roberts (1967, pl. 3).

places by Pennsylvanian rocks. Only in the northern area of its extent is it overlain by younger rocks of Jurassic age. The subcrops of the formations making up interval D on the erosional upper surface of the Mississippian System are shown in figure 68.

THICKNESS TRENDS

Interval D is a lens-shaped body of rock in the Williston basin; in the central parts of the basin it is as much as 500 feet thick (pl. 6-A). The interval is present in the Central Montana trough, where it gradually thickens westward, becoming more than 1,000 feet thick in central Montana.

The variation of thickness of interval D over the area is due largely to postdepositional erosion. Where stratigraphic units are not affected by post-Mississippian erosion, they have a relatively uniform thickness distribution that shows at best a minor tectonic or topographic influence on thickness.

LITHOFACIES TRENDS

Lithofacies of the interval are primarily controlled by the depth of post-Mississippian beveling. The interval becomes more shaly toward the center of the Williston basin and the Central Montana trough (pl. 6-B), reflecting the preservation from erosion of the younger and more shaly beds in these areas. The interval everywhere has 20–50 percent combined carbonate and anhydrite, and the ratio of carbonate to anhydrite is everywhere greater than 4; therefore, no appreciable variations of the chemical fraction of the interval shows up on the lithofacies map (pl. 6-B).

Rawson (1968) made a detailed study of the Kibbey Formation in which sedimentationally significant lithofacies and thickness variations were demonstrated. The lower member of the Kibbey (Rawson, 1968, fig. 4) ranges in thickness from 60 to 140 feet throughout the area, where it is conformably overlain by the ***Ray Member*** and has sand to shale ratios that range from $<1/4$ to >1. The thickness variations show no relation to the configuration of the Williston basin or Central Montana trough; however, the lithofacies are related in that the sand to shale ratio distribution, though somewhat complex, is lower in the central part of the basin than on its edges. The ***Ray Member*** (Rawson, 1968, figs. 5, 6) ranges in thickness from <20 to 50 feet and is generally thicker in the central parts of the basin. The member is dominantly limestone over nearly all of the Williston basin, but, as discussed under "Sedimentation," the distribution of limestone sub-

facies relates to the structural configuration of the Williston basin. The upper member of the Kibbey Formation (Rawson, 1968, fig. 7), where it is conformably overlain by the Otter, is 50–150 feet thick and tends to be somewhat thicker in the central parts of the basin and in the center of the Central Montana trough. The sand to shale ratio of the upper member of the Kibbey ranges from <1 to >4 and is generally lower in the area where the unit is thicker.

SEDIMENTATION

The Big Snowy Group represents a single episode of deposition beginning and ending with subaerial erosion (pl. 11, fig. 4). To the west in Montana, Maughan and Roberts (1967, p. B13, B23–B24) considered that the Big Snowy Group represents a cycle of deposition from red sandstone through green and red shale to dark shale and limestone, suggesting a transgression followed by gradual deepening of the sea as the region slowly subsided. They also suggested tectonically elevated surrounding areas to account for increased detritus in the Williston basin region, compared to the detritus supplied in earlier Mississippian time.

In the Williston basin, the Big Snowy Group represents the same cycle as in central Montana except that the limestone is more closely associated with the green shale than the dark shale. Willis (1959, p. 1947) indicated that in Heath time the Williston basin was covered by brackish to fresh water. A sea of normal salinity covered the area farther to the west in the Central Montana trough. Possibly, the eastern limit of the limestone in the Heath crudely marks the eastern limit of normal marine sedimentation in Heath time.

The source of detrital sediments during interval D time in the Williston basin was generally to the east. Using as evidence an entropy map of the Tyler-Heath interval, Willis (1959, p. 1947) postulated that considerable sorting and movement of the sand occurred within the Williston basin and that the sands were "deposited in a pattern approximating that of offshore bars." This reasonable postulation makes it difficult to identify the direction of the source of the detritus. Ballard (1965), however, pointed to a largely southern source for the detritus in the Kibbey Formation and stated that the southern edge of the Kibbey is essentially depositional.

The *Ray Member* of the Kibbey Formation was studied in detail by Rawson (1968), who showed that on the edges of the Williston basin the member consists mainly of almost pure carbonate micrite and in the center of the basin it consists of impure or argillaceous carbonate micrite. A narrow somewhat discontinuous belt of carbonate oolite separates the two micrite facies. Rawson interpreted the impure micrite to have been deposited in a low-energy basin area, the oolite on a high-energy bank margin, and the pure micrite on a low-energy bank.

PALEOTECTONIC IMPLICATIONS

The rocks of interval D record a single major transgressive-regressive cycle of deposition caused, presumably, by very low amplitude epeirogenic movements of the craton. Rocks of the interval were originally more widespread than at present, although their present southern edge probably is close to the original depositional edge (pl. 10, fig. 4). It is apparent, then, that the area to the north and east over which the sea transgressed extended beyond the Williston basin, though how far beyond is difficult to determine. The Siouxia arch was probably active and served as a barrier to the sea on the south, but the eastern and northern limits of the sea cannot be determined.

The Williston basin was more tectonically negative than the surrounding shelf during interval D time, as shown by the thickness and facies relations of the Kibbey Formation. The Cedar Creek anticline was positive and was a shoal that localized the deposition of carbonate oolite during the deposition of the *Ray Member* of the Kibbey.

The Central Montana trough also was more negative than surrounding shelf areas during interval D time.

At the end of deposition of the sediments of interval D, the shelf was more strongly uplifted than the basin, and subsequent erosion removed the rocks of the interval everywhere in the area, except in the Williston basin and Central Montana trough.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

The thickness pattern of the total Mississippian rocks in the Williston basin (pl. 7) reflects a fairly simple tectonic history. The rocks are thin on the shelf and are thick in the basin and in the Central Montana trough. In the center of the basin they attain a maximum thickness of almost 2,800 feet. The rate of thickening of the rocks is fairly even over all the region, except on the north side of the Central Montana trough where they thicken abruptly into the trough. The present margins of Mississippian rocks, except possibly on the south, are the limits after post-Mississippian erosion. Mississippian rocks probably extended considerably beyond their present area of distribution at the end of Mississippian time.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN

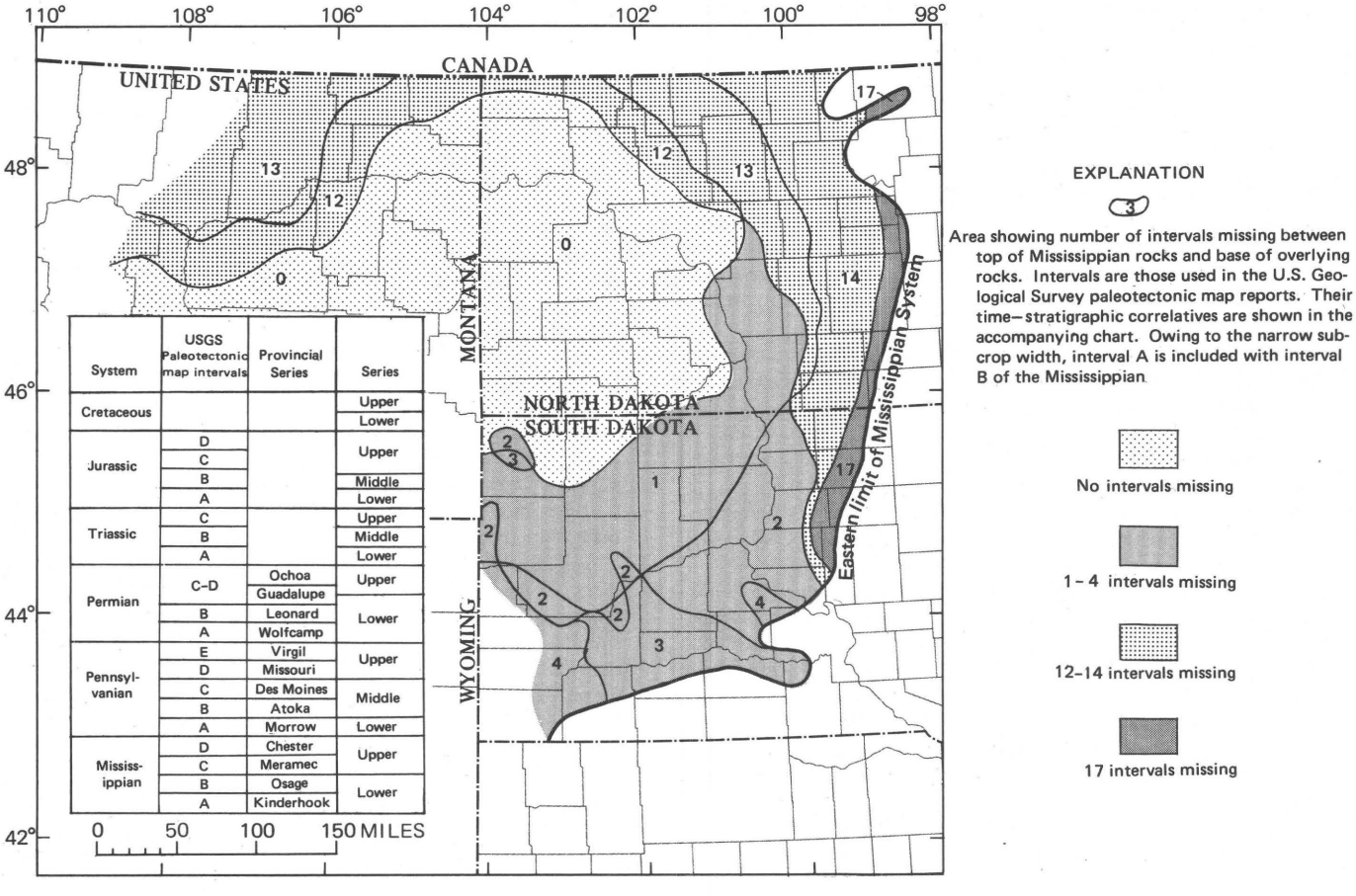
The Mississippian rocks are overlain (pl. 8) over most of the Williston basin by the Amsden Group of Pennsylvanian age (Maughan and Roberts, 1967, p. B11). In the area of the Black Hills, the Mississippian rocks are overlain by the Minnelusa Formation of Pennsylvanian age and the partial equivalent of the Amsden Group. The basal unit of the Amsden Group is the Tyler Formation of Morrow age; this formation is the most extensive unit of the Amsden in the Williston basin and is overlapped by younger Pennsylvanian rocks in only a few areas. However, in western South Dakota several windows alined in a belt having a northwest trend appear in the Tyler, and rocks of Atoka or Des Moines age lie directly on the Mississippian.

The Pennsylvanian rocks are overlapped by the *Saude Formation* of probable Middle Jurassic age (Ziegler, 1956; Sandberg, 1962) in the northern and eastern parts of the Williston basin. In eastern North Dakota and South Dakota the *Saude Formation* is overlapped by Lower Cretaceous rocks.

PALEOTECTONIC IMPLICATIONS

The disconformity at the top of Mississippian rocks has relatively minor relief and is generally marked by a near concordance of strata in the Williston basin, but the time value of the disconformity has a large range. This time value is illustrated for different parts of the basin in figure 69.

In the central part of the Williston basin and in the Central Montana trough, the time value is at a minimum inasmuch as the youngest Mississippian rocks are overlain by the oldest Pennsylvanian rocks, and no time intervals are missing. Away from these areas, the time value gets progressively larger until at the outer edge of Mississippian rocks 17 intervals, as defined in figure 69, are missing at the disconformity. A general shelfward increase in time value occurs, with the largest jumps in the time value of the disconformity at the points were Pennsylvanian and then Jurassic rocks are overlapped. This progressive shelfward in-



System	USGS Paleotectonic map intervals	Provincial Series	Series
Cretaceous			Upper
			Lower
Jurassic	D		Upper
	C		
	B		Middle
	A		Lower
Triassic	C		Upper
	B		Middle
	A		Lower
Permian	C–D	Ochoa	Upper
		Guadalupe	
	B	Leonard	Lower
	A	Wolfcamp	
Pennsylvanian	E	Virgil	Upper
	D	Missouri	
	C	Des Moines	Middle
	B	Atoka	
	A	Morrow	Lower
Mississippian	D	Chester	Upper
	C	Meramec	
	B	Osage	Lower
	A	Kinderhook	

FIGURE 69. — Time value of the disconformity between Mississippian and younger rocks.

crease in time value of the disconformity is the result of the more positive tendency of the shelf areas relative to the Williston basin and Central Montana trough during post-Mississippian uplift and erosion.

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Great Basin Region

By BETTY SKIPP

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–P

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

GREAT BASIN REGION

By BETTY SKIPP

ABSTRACT

Late Devonian crustal instability ended the geosynclinal sedimentation patterns that dominated the Great Basin region from late Precambrian to Devonian time. The Antler orogenic belt formed on the western edge of the continent between the eastern carbonate (miogeosynclinal and transitional) and western siliceous and volcanic (eugeosynclinal) lithosomes of earlier periods. West-to-east thrusting telescoped the diverse geosynclinal facies across a wide belt through central Nevada in latest Devonian and, possibly, earliest Mississippian time. The Antler highland controlled Great Basin clastic sedimentation during the Mississippian Period. Thick aprons of fine- to coarse-grained detrital sediments (flysch) filled the troughs that developed along the eastern edge of the highland, and in Late Mississippian time, the detrital sediments spread across the central Great Basin region into western Utah.

The Antler highland stood relatively low in Early Mississippian time (intervals A and B). During and after deposition of the widespread Pilot Shale, which is excluded from interval A in this report, shale and siltstone of the well-dated Webb and Chainman Formations were deposited in the northern part of the Chainman—Diamond Peak trough along the northeast edge of the highland. Undated fine-grained detrital sediments of the Chainman Shale in the Eureka area and the Eleana Formation in southern Nevada and equivalent strata in California may have accumulated simultaneously in the southern part of the Chainman–Diamond Peak trough and the Eleana trough, respectively, along the eastern edges of the highland. Sediment patterns of intervals A, B, and C suggest that the highland may have been divided into two segments by a large embayment or possibly a strait. East of the troughs, thin carbonate lithosomes of the Gardison Limestone and Fitchville Formation, the Joana Limestone, the Monte Cristo Limestone (lower part), and the Tin Mountain Limestone were laid down across the Great Basin from northern Utah to southern California on an erosional surface developed after deposition of the Pilot Shale. Shallow marine waters covered the relatively stable shelf area between the emergent craton to the east and the troughs to the west during most of intervals A and B time. The Wendover highland in northern Utah was intermittently emergent during interval B time.

In Late Mississippian time (intervals C and D) crustal stability of the central and eastern Great Basin area came to an end. The Antler highland became a major mountain chain, and steep gradient streams flowing eastward into the Chainman—Diamond Peak and Eleana troughs deposited a minimum of 5,500 and a maximum of 10,000 feet of coarse to fine detritus adjacent to the highland. In interval C time, the Ely arch emerged over a large part of eastern Nevada and western Utah, and parts of the Lower Mississippian carbonate lithosome were eroded from it before it was submerged again during interval D and buried under 1,500–2,000 feet of siltstone and shale. The northeast-trending elongate troughs of the Oquirrh basin in northern Utah began to subside and in Late Mississippian time received a minimum of 6,000 feet of impure carbonate rocks including the Deseret Limestone, the Humbug Formation, and the Great Blue Limestone. The small, but active Wendover highland in northwestern Utah northeast of the Ely arch was emergent during most or all of Late Mississippian time, and fine to coarse detritus from it accumulated in a clastic apron which graded southeastward into the thick impure carbonates of the Oquirrh basin.

Uplift of the Antler highland was sporadic. Major pulses recorded by extensive conglomerate wedges of the Diamond Peak and Chainman Formations of northern Nevada occur through interval C into late interval D (late Chester) time. In later Mississippian time, tectonic activity subsided, and the impure carbonate of the lower part of the Ely Limestone was laid down over a large part of eastern Nevada and western Utah.

No hiatus is recognized in the Ely Limestone of the eastern Great Basin area at or near the Mississippian-Pennsylvanian systemic boundary. Limestones of the lower part of the Ely contain a rather complete faunal record of latest Mississippian and earliest Pennsylvanian time. Similar faunas are present in central Idaho. Elsewhere in the Great Basin region, however, along the eastern margin of the Antler highland and in southeastern California, major unconformities occur at or near the Mississippian-Pennsylvanian systemic boundary.

West of the Antler highland, Mississippian rocks are sparse. Early Mississippian volcanism and eugeosynclinal sedimentation are recorded in the shales and cherts of the Peale Formation in the Taylorsville area of northern California. Upper Mississippian altered andesite and basalt of the Goughs Canyon Formation in north-central Nevada have been thrust onto the western edge of the Antler highland in north-central Nevada, but were probably deposited west of the continental margin.

Upper Mississippian andesite and basalt of the Nelson and Schoonover Formations are also present east of the northeast edge of the Antler highland in north-central Nevada. The Schoonover is known to be allochthonous, and may have been deposited west of the Antler highland on oceanic crust. In this report, however, volcanism is shown as having taken place along the eastern edge of the highland.

The record of Mississippian sedimentation in the Great Basin is complicated throughout the region by post-Mississippian thrusts, basin-and-range extension faults, oroflexural bends, and major right-lateral shears. The effects of these structures have been considered to some degree in this chapter.

REGION DEFINED

The Great Basin region of this report (fig. 70) includes Nevada, the southern two-thirds of California from a line just north of the Fortieth parallel to Baja California, and a part of western Utah. The part of Utah included in this region lies generally west of the Wasatch line (Stille, 1936; Kay, 1951, p. 14), though only the southern part of the Wasatch Range is treated here; the northern part of the Wasatch Range is included in the discussion of the Northern Rocky Mountains region (Roberts, chap. N).

The Great Basin region includes Mississippian rocks of dominantly marine shelf and delta, and volcanic facies and their metamorphosed equivalents. The region was affected by several periods of deformation in late Paleozoic and Mesozoic time. In addition, Cenozoic basin-and-range faults have complicated the Paleozoic and Mesozoic records, and much of the area is covered by Cenozoic volcanic and sedimentary rocks. Holocene and older strike-slip faults and oroflexural bends in the Walker Lane structural zone and Furnace Creek–Death Valley shear zones are thought to have effected, in large part, the anomalous distribution of Mississippian rocks along the southeastern California-Nevada border.

Three positive areas — the Antler highland (or Antler orogenic belt) in central Nevada, its corollary, the Wendover highland, in northwestern Utah, and the Ely arch in northeastern Nevada and west-central Utah — were major structural elements in the region during Mississippian time. The Chainman–Diamond Peak trough and the related Eleana trough east of the Antler highland were the dominant negative contemporaneous structural features. In Late Mississippian time, the Oquirrh basin began to subside in northern Utah, and during Pennsylvanian and Permian time this depression became a major negative structural feature (Roberts and others, 1965).

STRUCTURAL SETTING

DEVONIAN STRUCTURAL FEATURES

In Late Devonian time, the geosynclinal pattern — eugeosyncline to the west, miogeosyncline and craton to the east — which had dominated early and middle Paleozoic deposition in the Great Basin was interrupted by widespread crustal movement (Roberts and others, 1958, p. 2813; Poole and others, 1967, p. 879). Eugeosynclinal and transitional rocks were deformed and uplifted to form the Antler orogenic belt (fig. 71*A*) in Nevada which became the source for the important,

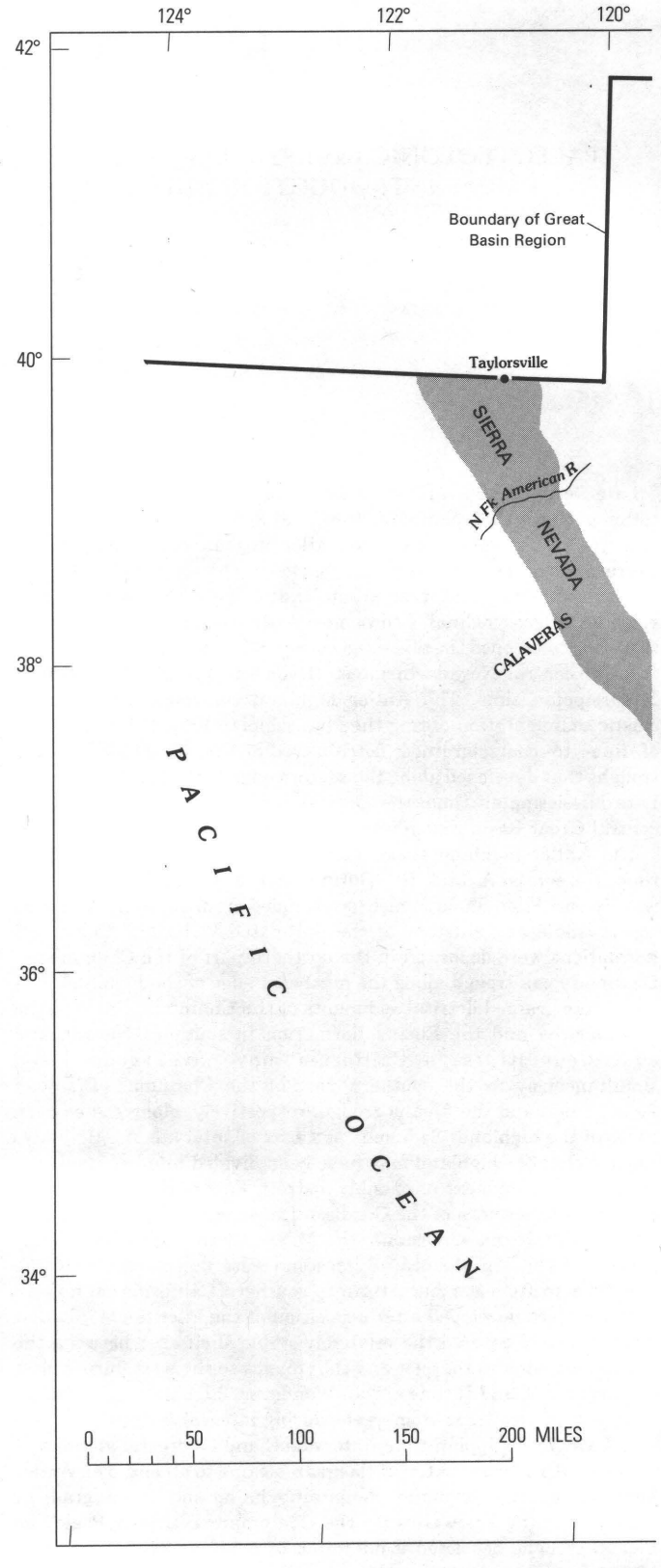
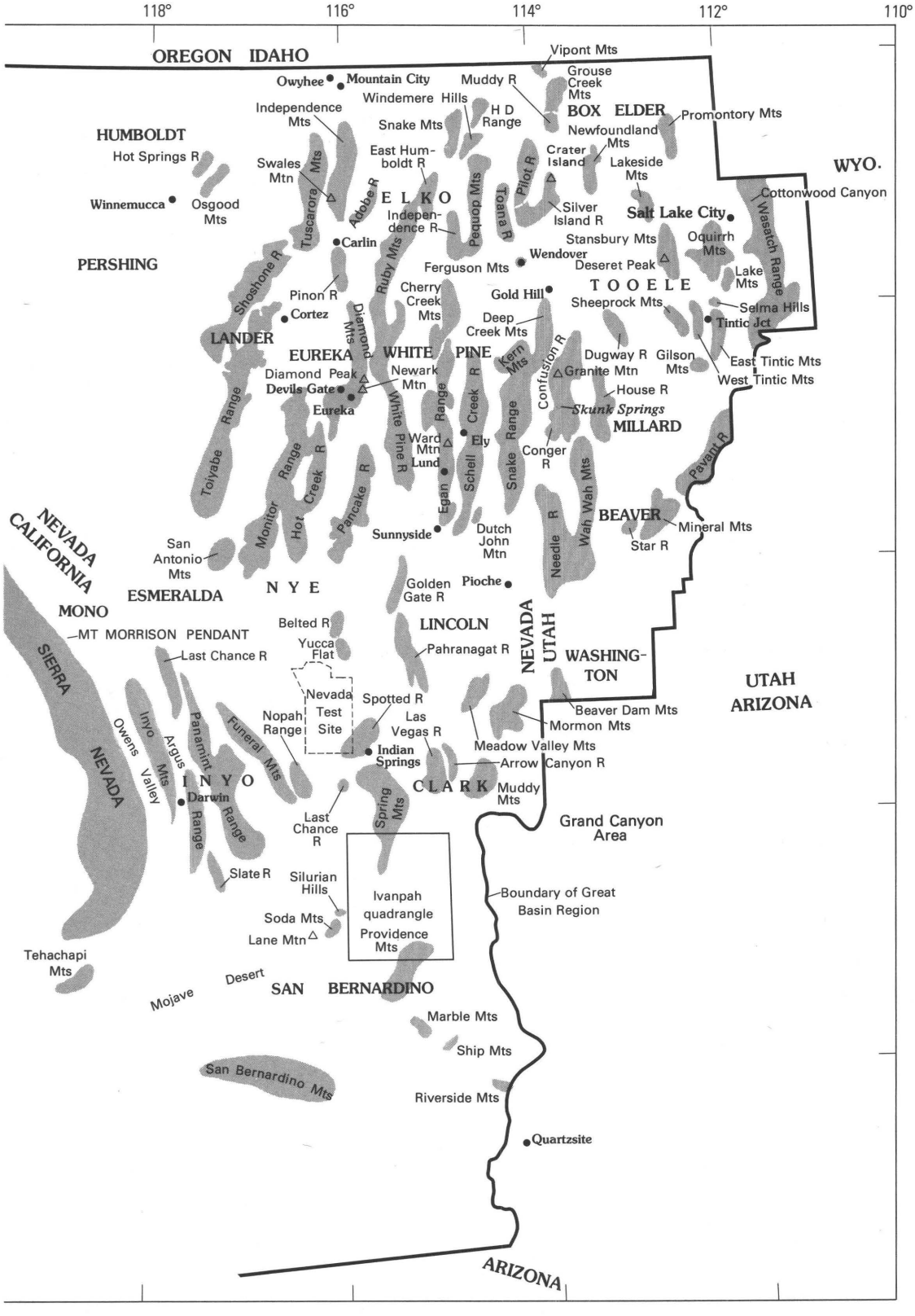
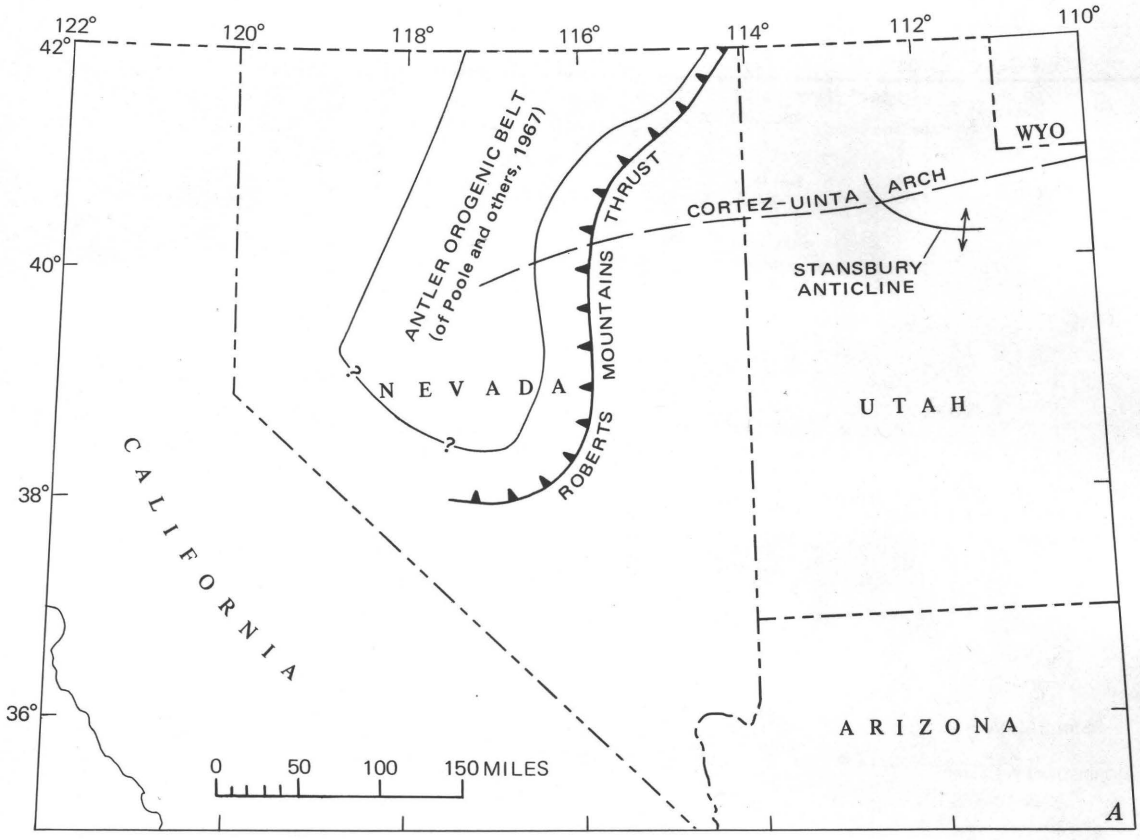
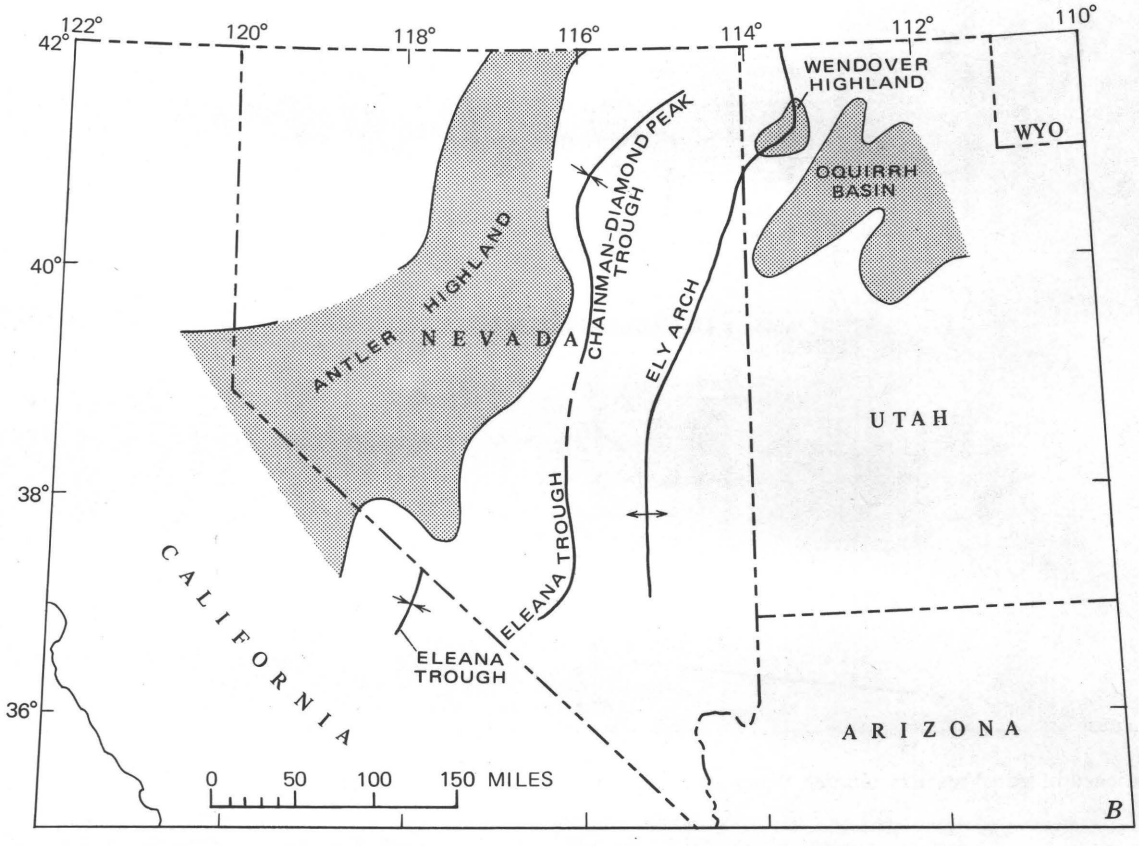


FIGURE 70. — Limit of Great Basin region and geographic



features mentioned in text. Vipont Mountains, Selma Hills, and HD Range are informal names.


122°
120°
118°
116°
114°
112°
110°
42°
40°
38°
36°
ANTLER OROGENIC BELT
(of Poole and others, 1967)
THRUST
ROBERTS MOUNTAINS
CORTEZ-UINTA ARCH
STANSBURY ANTICLINE
WYO
NEVADA
UTAH
CALIFORNIA
ARIZONA
0 50 100 150 MILES
A


122°
120°
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38°
36°
WENDOVER HIGHLAND
WYO
OQUIRRH BASIN
CHAINMAN-DIAMOND PEAK TROUGH
ELY ARCH
ANTLER HIGHLAND
NEVADA
UTAH
ELEANA TROUGH
ELEANA TROUGH
CALIFORNIA
ARIZONA
0 50 100 150 MILES
B

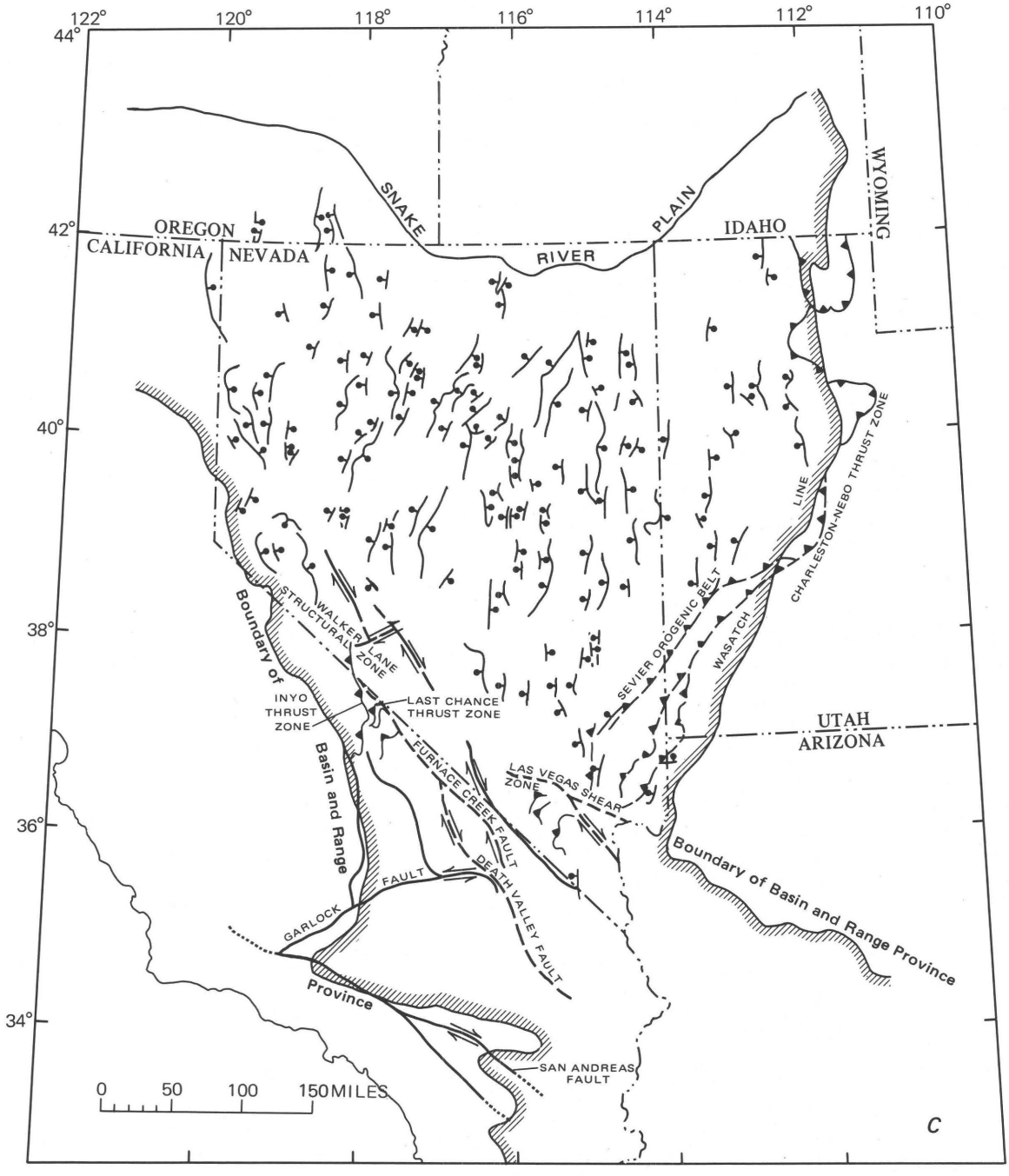


FIGURE 71 (above and facing page). — Structural features. *A*, Devonian and earliest Mississippian modified from Poole and others (1967, fig. 1), *B*, Mississippian, *C*, post-Mississippian; teeth indicate upper plate of thrust fault; hachures indicate downthrown side of normal fault. Basin-and-range faults modified from Stewart (1971, fig. 1).

largely fine-grained, terrigenous detrital facies on the shelf or miogeosyncline (Poole, 1973). These same eugeosynclinal and transitional rocks were then thrust eastward, on the Roberts Mountains allochthon, over autochthonous miogeosynclinal rocks in latest Devonian (Smith and Ketner, 1968) and earliest Mississippian (Ketner, 1970b) time. A major unconformity within the Devonian part of the Pilot Shale in the Pahranagat Range (fig. 70), recently determined from paleontologic evidence, tends to verify major local tec-

tonism in the miogeosyncline in latest Devonian time (Sandberg and Poole, 1970). Early Mississippian shales overlap the Roberts Mountains thrust which involved Late Devonian rocks in the Pinon Range of north-central Nevada (Smith and Ketner, 1968).

Local uplifts such as the Stansbury anticline (fig. 71*A*) and the Wendover highland (fig. 71*B*) were activated in northern Utah in Late Devonian time. These local uplifts were subjected to short periods of moderately intense erosion which stripped older rocks from emergent parts. Thus, the succeeding major Early Mississippian transgression overlapped rocks ranging in age from earliest Mississippian to Cambrian and ended Late Devonian epeirogeny along the eastern margin of the geosyncline (Rigby, 1959a; Poole and others, 1967, p. 907).

MISSISSIPPIAN STRUCTURAL FEATURES

The extensive, north-northeast-trending Antler highland originated in Late Devonian time and remained emergent during most or all of Mississippian time (fig. 71*B*). Varying amounts of fine to coarse detritus were eroded from the highland to form the clastic wedges in the adjacent Chainman–Diamond Peak and Eleana troughs. Still farther east, sporadic uplift of the Ely arch and Wendover highland accompanied the filling of the troughs. In northern Utah, the first subsidence of the Oquirrh basin took place. Intermittently, andesitic to basaltic volcanic rocks were extruded west and possibly, locally, east or north of the Antler highland. Some volcanic debris derived from these areas is recognized in the shale of the Chainman and Eleana Formations (F. G. Poole, oral commun., 1973). The allochthonous volcanic rocks now present on and along the western edge of the Antler highland may have been originated in a relatively distant island arc system which was moved eastward in post-Mississippian time. Much the same history has been proposed for the Pennsylvanian and Permian Havallah Formation (Silberling and Roberts, 1962). Local allochthonous basaltic rocks in northeastern Nevada on the northeastern edge of the highland probably originated west of their present position (Fagan, 1962) but possibly still east or north of the highland.

POST-MISSISSIPPIAN STRUCTURES

Post-Mississippian structures have greatly modified the presumed original depositional positions of Mississippian rocks in the Great Basin. Eastward thrusting and gravity sliding of late Paleozoic (Roberts, 1968) through Tertiary (Willden and others, 1967) age have produced shallow crustal shortening which has brought together Mississippian rocks of dissimilar facies in many areas of eastern Nevada and western Utah. A few obvious examples are present in Nevada in the Diamond Mountains (Brew, 1961b), the Ruby Mountains (Sharp, 1942; Willden and others, 1967), the Independence Mountains (Fagan, 1962; Churkin and Kay, 1967; Ketner, 1970b), the Windermere Hills (Oversby, 1969, 1972), the Pequop Mountains (Sadlick, 1965; Thorman, 1962, 1970), and the Snake Range (Nelson, 1966), and in western Utah in the Oquirrh Mountains (R. J. Roberts, written commun., 1970), southern Wasatch Range (Baker and Crittenden, 1961), and Wah Wah Mountains (Miller, 1966) (fig. 70). In areas, such as the northern Snake Mountains (Nev. loc. 23) in northern Elko County, Nev., a Mississippian zero isopach is placed where allochthonous Pennsylvanian and Permian rocks override allochthonous lower Paleozoic rocks (Gardner, 1968), even though the underlying unpenetrated autochthon very likely contains a thick sequence of Mississippian rocks. Direct evidence for such a thick sequence exists in the HD Range north of the Windermere Hills where a window in a complex of thrust plates exposes a minimum of 300 feet of Chainman Shale in fault contact with Diamond Peak, but beneath adjacent plates in which Middle Pennsylvanian rocks unconformably overlie Ordovician and Silurian rocks (Riva, 1970). Complex thrusting and facies relationships in Nye County, Nev., are evident in many areas in the northern (Kleinhampl and Ziony, 1967) and southern (Poole and others, 1961) parts of the county.

Significant thrusting is also present in northern Inyo County of southeastern California, where most of the Inyo Mountains and the White Mountains to the north may have moved eastward along the Inyo and Last Chance thrust zones offsetting Mississippian facies (Stewart and others, 1966, p. D31; Stevens and Olson, 1972, p. 3765).

The Sevier orogenic belt of Cretaceous age (Armstrong, 1964, 1968a), which has telescoped facies along the eastern edge of the Great Basin from western Utah through southern Nevada and into southeastern California, is shown as a continuous fault system (fig. 71*C*; pls. 2–8) on the basis of correlation of local fault segments in northern Utah (Crittenden, 1959, 1961), in western Utah (Armstrong, 1964, 1968a; Poole and others, 1967; and Woodward, 1970), and in southern Nevada and adjacent California (Poole and others, 1967). Armstrong (1968a, p. 441) stated "that total shortening of 40 miles across the Sevier belt is a minimum value and 60 miles is adequate to accommodate comfortably any of the estimates." Immediately west of the Sevier belt, fault patterns in southern Nevada and southern California are based on compilations by F. G. Poole (written commun., 1966).

Several right-lateral strike-slip fault or shear zones,

active in Mesozoic and Cenozoic time, have displaced Mississippian rocks and older fault structures in the southern parts of Nevada and southern California. Poole (in Stewart and others, 1968) showed that offset of the southeastern limit of Upper Mississippian (upper Chester or interval D) shale units indicates 30–40 miles of right-lateral displacement along the Las Vegas shear zone, and 60–80 miles along the Death Valley – Furnace Creek fault system. In addition, offset of the western limit of Lower Mississippian limestone units indicates 50–60 miles on the Death Valley–Furnace Creek system. These relations are shown in figure 72, which is modified from Stewart, Albers, and Poole (1968, fig. 2). Offset of Mississippian total and interval isopachs on plates 3–7 support these estimates. In addition, the sigmoidal or oroflexural bending described by Albers (1967) and suggested by the eastern limit of Permian, Triassic, and Jurassic volcanic rocks and the eastern limit of Middle and Upper Triassic sedimentary rocks (fig. 72) is similar to the bending of zero Mississippian isopachs shown in this report (pl. 7) along the eastern edge of the Antler orogenic belt in southern Nevada and adjacent California.

In addition to regional shear zones, Eocene (Gilluly, 1932, 1963) to Holocene basin-and-range extension faulting and concurrent volcanism modify and complicate the older thrust patterns throughout the Great Basin region. Hamilton and Myers (1966) suggested that total Cenozoic crustal extension across northern Nevada and Utah may be at least 60 miles and Stewart (1971) estimated 45 miles. The San Andreas strike-slip fault system, active since the early Tertiary (Crowell, 1962), has a total right-lateral displacement of at least 160 miles and possibly as much as 350 miles (Hill and Dibblee, 1953; Crowell, 1962). Left-lateral displacement of 40 miles since middle Mesozoic time on the east-west-oriented Garlock fault system has been suggested by Smith (1962). A combined paleogeologic and palinspastic map of the western United States at the beginning of Tertiary time (Hamilton and Myers, 1966, p. 541) shows restored basin-and-range tensional displacements. This palinspastic reconstruction forms the basis for figure 73 on which the major recognized Mississippian structural features of the Great Basin on the total isopach map (pl. 7) are superimposed.

Geophysical evidence indicates that the eastern part of the Snake River Plain, which bounds the northeastern portion of the Great Basin of this report, is a region of continental crust thinned by tension but not rifted beneath the thick Cenozoic volcanic rocks (D. R. Mabey, oral commun., 1970; Hamilton and Myers, 1966, p 540). The possibility of some lateral movement between the north and south sides of the plain has been proposed by Sandberg and Mapel (1967).

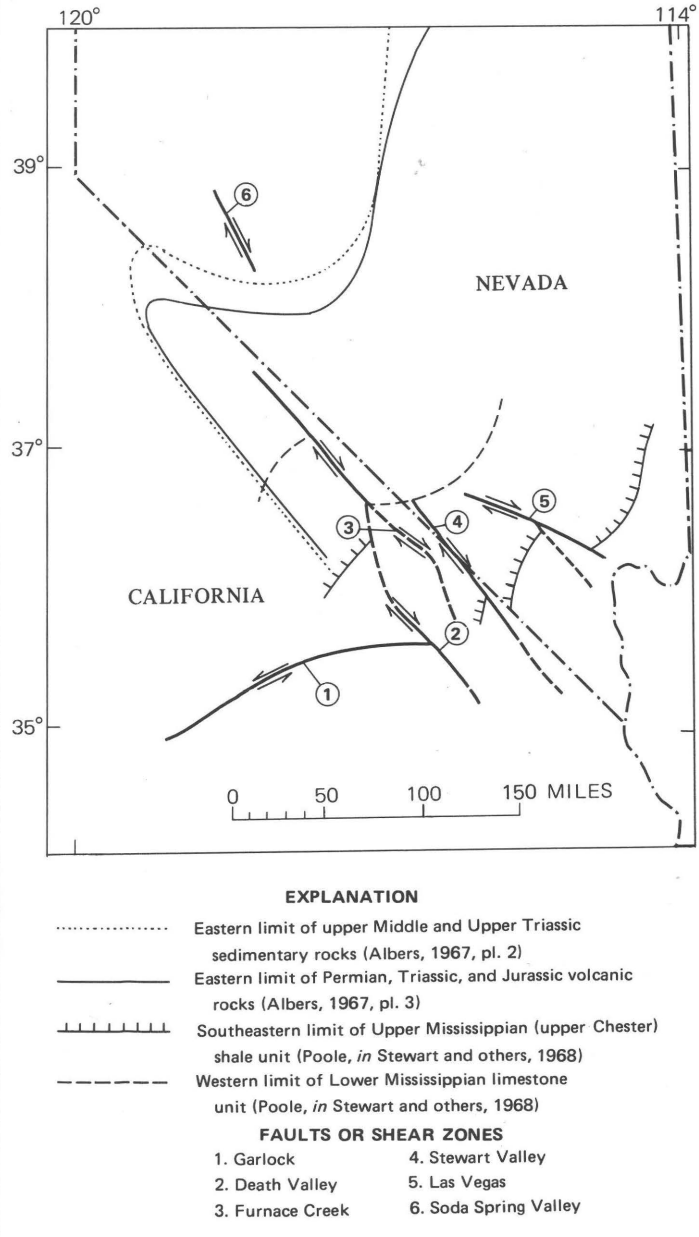


FIGURE 72. — Mississippian to Mesozoic isopach and facies trends in the Great Basin. Modified from Stewart, Albers, and Poole (1968, fig. 2).

PALEOGEOLOGY

UNITS UNDERLYING MISSISSIPPIAN AND LOWER BOUNDARY OF INTERVAL A

Little is known of the units underlying the Mississippian System in the area west of the Antler highland in western Nevada and eastern California. Near Taylorsville in northern California, a calcareous mudstone lens in the upper part of the Peale Formation contains Early Mississippian faunas and is underlain by the Taylor Metaandesite of Mississippian age. The Taylor Metaandesite overlies the *Sierra Buttes Forma-*

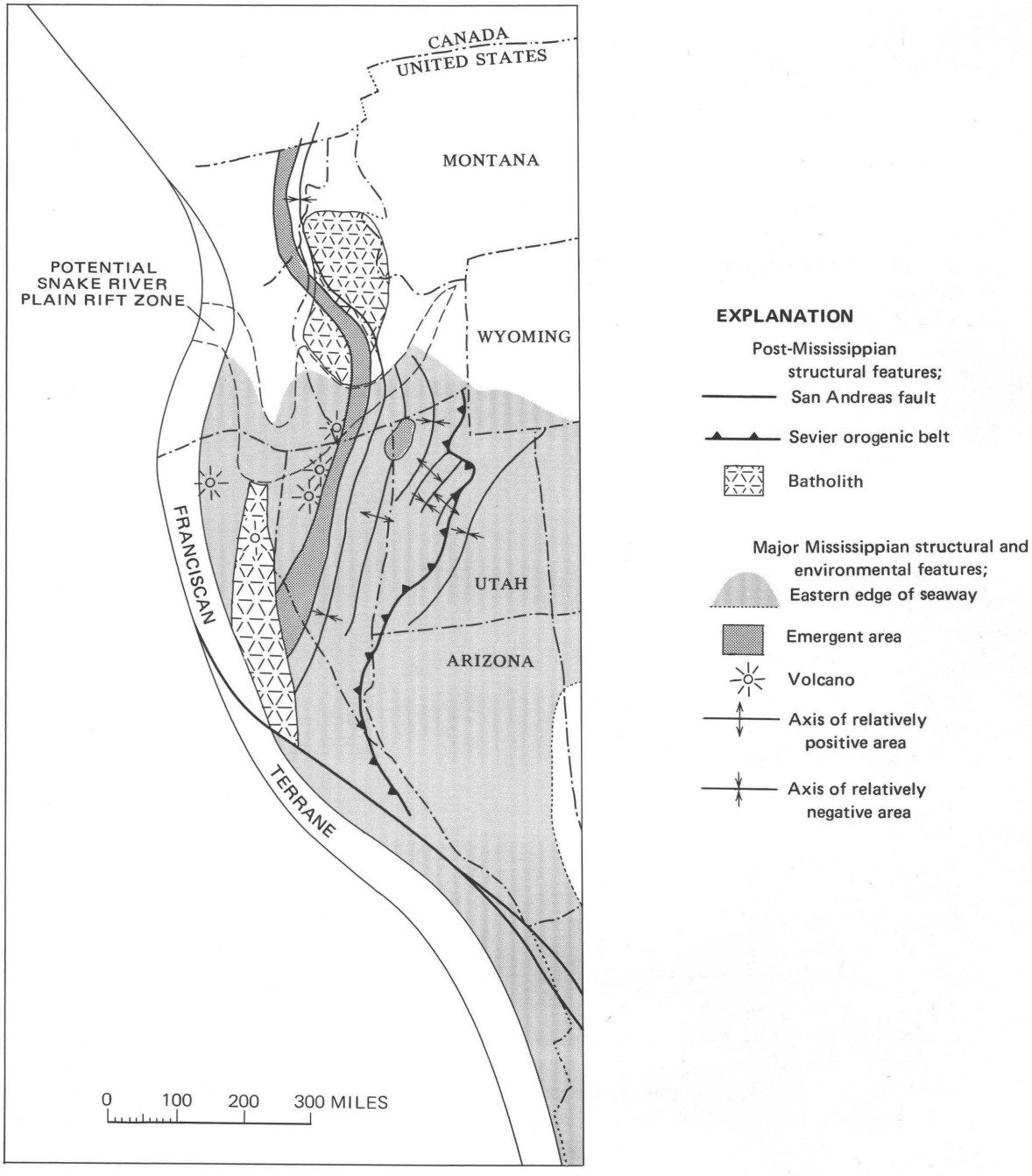


FIGURE 73. — Major Mississippian structures of the Great Basin and adjacent regions, taken from the total isopach map and plotted on a palinspastic reconstruction of the Western United States at the beginning of Tertiary time, prior to development of basin-and-range extension fault system. Modified from Hamilton and Myers (1966, fig. 9).

tion of McMath (1966, p. 176–180), for which a Devonian age is possible. In the Hot Springs Range and Osgood Mountains of eastern Humboldt County, Nev., the upper and lower contacts of the Goughs Canyon Formation are thrust faults, and stratigraphic relations with older and younger units are unknown (Willden, 1964; Hotz and Willden, 1964, p. 25–26). In the Hot Springs Range, the Goughs Canyon overrides rocks as old as Jurassic and is overlain unconformably by Tertiary andesite and basalt. Devonian rocks are present, however, in the nearby Shoshone Range in Lander County on the upper plate of the Roberts Mountains thrust (Roberts and others, 1958), and Willden (1964, p. 21) stated that "a large part of Humboldt County may once have been covered by [Silurian and Devonian] rocks." To the southwest, in Pershing County, Nev., the

Inskip Formation is underlain by the *Leach Formation* of Mississippian or older age. "The Leach seems to be older than the Inskip, although their mutual contact is for the most part steeply faulted." (Silberling and Roberts, 1962, p. 13).

In the central part of the Great Basin, upper Upper Devonian (Famennian) rocks, including in places lowermost Mississippian, underlie rock units assigned to interval A (pl. 2). Formations present include the Woodruff Formation in the Pinon Range, Nev., the Pilot Shale in eastern Nevada and western Utah, the Pinyon Peak, Stansbury, and Hanauer Formations in central and northwestern Utah, and the lower part of the Eleana Formation in southern Nevada. Devonian faunas have been recovered from units A and basal B of the Eleana Formation. Upper B, units C to G, and lower H remain undated, but Late Mississippian faunas have been recovered from units H through J (F. G. Poole, oral commun., 1968; Poole, in Ekren and others, 1971, p. 23).

The Woodruff Formation is an allochthonous assemblage of siliceous mudstone and chert exposed in the Pinon Range, Elko County, Nev., which is overlain unconformably by beds of interval A. The age of the Woodruff based on goniatites and conodonts is Middle to late Late Devonian (Smith and Ketner, 1968, 1972). The upper unit of the Pilot Shale where present in eastern Nevada and western Utah is Early Mississippian (Kinderhook) in age (Sandberg and Poole, 1970; Tschanz and Pampeyan, 1970; Kellogg, 1963; Hose, 1966). Sandberg and Poole (1970) reported the Mississippian conodont *Siphonodella* in the basal 55 feet of the upper unit (302 feet thick) of the Pilot Shale in the Pahranagat Range, Nev. The Mississippian part of the Pilot, which locally has been removed by pre-Joana erosion (Langenheim, 1961; Kellogg, 1963; Hose, 1966), was excluded from interval A in the present report, mainly because of the difficulty in determining its regional extent from the literature. For similar reasons it was included in the synthesis of the Devonian System by Poole and others (1967, fig. 6).

The Pinyon Peak Limestone in central and northwestern Utah is considered Late Devonian on the basis of megafaunas (Morris and Lovering, 1961) and conodont microfaunas (Beach, 1961; Clark and Ethington, 1967). Morris and Lovering (1961) considered the uppermost part of the Pinyon Peak as possibly Early Mississippian, whereas Clark and Ethington (1967) reported a latest Devonian conodont fauna from 128 feet above the base of the overlying Fitchville Formation. A Late Devonian age for the lower part of the Fitchville seems established inasmuch as Late Devonian conodonts (J. W. Huddle, written report to Helen Duncan, 1966; C. A. Sandberg, oral commun., 1972) occur in the lower part of the Fitchville near the type section of the Pinyon Peak. These relations are shown on the correlation chart (pl. 15) but, because the Devonian part of the Fitchville is thin, all the Fitchville has been assigned arbitrarily to interval A (pl. 3). The contact between the Pinyon Peak and Fitchville is considered conformable everywhere. An angular unconformity in the northern Gilson Mountains reported by Costain (1960) has been recognized as a thrust contact (Wang, 1970, p. 88). The Stansbury Formation (Stokes and Arnold, in Rigby, 1958) has been correlated with the Pinyon Peak (Rigby, 1959b, p. 31) or considered slightly older (Poole and others, 1967, p. 885). The greater part of the Hanauer Formation is known to be Late Devonian in age but the Devonian-Mississippian boundary may lie in the upper part of the formation (Staatz and Carr, 1964, p. 61).

In areas peripheral to those in which upper Upper Devonian rocks are preserved (pl. 2), Mississippian rocks unconformably overlie lower Upper Devonian rocks. Devonian formations present in these peripheral areas include the Guilmette Formation, the Devils Gate Limestone, and a Devonian limestone and chert transitional unit (Smith and Ketner, 1968, p. 17, 1975, p. A34) in northeastern Nevada; the Guilmette, the Jefferson, Beirdneau, and lower part of the Leatham Formations in northern Utah, the Guilmette in an isolated area of eastern Nye County, Nev., the Lost Burro Formation, the Muddy Peak Limestone, the *Arrow Canyon Limestone* of Langeheim and others (1962), and the Valentine Limestone Member of the Sultan Limestone in southeastern California and southern Nevada. The fact that some younger Devonian rocks are included in interval A in the Crystal Pass Limestone in some areas of Nevada and California is discussed under the section "Interval A." Lower Upper Devonian rocks appear to underlie the Mississippian east of the Sevier belt in southwestern Utah.

Middle and Upper Devonian, Silurian, Ordovician, and Cambrian rocks are unconformably overlain by Mississippian rocks in the vicinity of the Stansbury anticline (fig. 71*A*; Rigby, 1959a; Poole and others, 1967, fig. 1) in northern Utah. Late Devonian uplift of this feature provided coarse detritus for the surrounding Upper Devonian Stansbury Formation which Roberts and Tooker (1969) suggested may extend as far west as the Newfoundland Mountains, and possibly the Silver Island Range. The Middle Devonian Simonson Dolomite, the Silurian Laketown Dolomite, the Ordovician Garden City Limestone, and several Cambrian formations — including the Ajax and Lynch Dolomites, the Maxfield Limestone, and the Ophir Shale — are exposed under Mississippian rocks in the vicinity of the Stansbury anticline.

The Lower Devonian Water Canyon Formation underlies Mississippian rocks in the Promontory Mountains in northern Utah (Olson, 1960).

Cambrian, Ordovician, Silurian, and Lower and Middle Devonian rocks are unconformably overlain by Mississippian rocks in several restricted areas along the eastern edge of the Antler highland in Eureka and Elko Counties, Nev. (pl. 2). In the Pinon Range area, Mississippian rocks of interval A unconformably overlap Devonian and older rocks of both siliceous (western) and carbonate (eastern) assemblages juxtaposed by the Roberts Mountains thrust (Smith and Ketner, 1968, 1972, 1975). The Ordovician Vinini Formation underlies Mississippian rocks in the Carlin Canyon area, east of Carlin, Nev. Ordovician chert and the Silurian *Noh Formation* are unconformably overlain by the Mississippian Diamond Peak Formation in the HD Range (Riva, 1970). Silurian rocks assigned to the *Taylor Canyon* and *Happy Camp Formations* and Ordovician and Silurian rocks of the Hanson Creek Formation are unconformably overlain by the *Waterpipe Canyon Formation* (Kerr, 1962), provisionally assigned to Mississippian interval A. The Ordovician Valmy Formation is unconformably overlain by the Mississippian volcanic rocks of the Schoonover Formation (Fagan, 1962) in the northern Independence Mountains (Churkin and Kay, 1967), and by the limestone and volcanic rocks of the Banner and Nelson Formations in the Mountain City–Owyhee area to the north. Both of these sequences of Ordovician Valmy Formation overlain by Mississippian volcanic rocks are thought to be allochthonous (Churkin and Kay, 1967, p. 664; R. R. Coates, written commun., 1973). A Carboniferous limestone in the northern Independence Mountains, Nev., unconformably overlies undivided Cambrian rocks (Churkin and Kay, 1967). Immediately north and east of the Independence Mountains, Mississippian(?) rocks of various facies overlie the Valmy Formation.

At Swales Mountain in the southern part of the Independence Mountains, Nev., limestones of Mississippian interval A seem to be gradational with the Devonian part of the underlying Roberts Mountains Formation (Ketner, 1970b).

In northwestern Nye County, Nev., along the eastern edge of the Antler belt, conglomerates tentatively assigned to the Mississippian in the central Monitor Range (Nev. loc. 329) are shown as unconformable on the Lower and Middle Devonian Nevada Formation and the Ordovician Vinini Formation by Kleinhampl and Ziony (1967). Faulting may be present here, however. Also, Mississippian limestone at the base of the Eleana Formation assigned to interval A is conformable on the Nevada Formation in the northern Monitor Range (Nev. loc. 328) and in the southern Hot Creek Range (Nev. loc. 330) (Kleinhampl and Ziony, 1967).

In southwestern Nye County and in southern Esmeralda County, parts of the Mississippian Meiklejohn Formation (of former usage) are in fault contact with Cambrian, Ordovician, and Devonian rocks (Cornwall and Kleinhampl, 1961). None of these are shown on plate 2.

In the northwestern Inyo Mountains, Inyo County, Calif., the Mississippian Perdido Formation is unconformable on Silurian-Devonian(?) rocks assigned to the Sunday Canyon Formation and Vaughn Gulch Limestone (Ross, 1966, p. 35; Stewart and others, 1966, p. D25).

The Furnace Limestone in the San Bernardino Mountains of southern California, which is in part assigned to interval A, is underlain by the Saragossa Quartzite of Precambrian and Cambrian age (Stewart and Poole, 1975). According to F. G. Poole, the Furnace Limestone may include carbonate strata ranging in age from Middle Cambrian to Carboniferous (Stewart and Poole, 1975, p. 209).

PALEOTECTONIC IMPLICATIONS

Geosynclinal sedimentation that dominated the Great Basin region during the early Paleozoic was interrupted during Late Devonian time by widespread crustal movements. Lower Paleozoic rocks of eugeosynclinal and transitional facies were uplifted in central Nevada and were thrust eastward over autochthonous miogeosynclinal rocks in latest Devonian and earliest Mississippian time. The uplifted area, the Antler orogenic belt or highland, provided detritus for widespread deposits of shale, siltstone, and minor sandstone of the Pilot Shale, which accumulated in the position of the miogeosyncline in eastern Nevada and western Utah. Local uplifts, such as the Stansbury anticline, formed in Utah by folding and faulting related to the orogeny and became sources for local Upper Devonian clastic deposits (Poole and others, 1967, p. 879). These major structural patterns are shown on the paleogeologic map (pl. 2).

The record of deposition across the Devonian-Mississippian system boundary seems complete in the central and eastern parts of the Great Basin in the Pinyon Peak–Fitchville sequence, and locally, in the Pilot Shale and lower part of the Eleana Formation. Unconformities separate Devonian and older rocks from Mississippian strata along the margins of these continuous sequences, however, reflecting the widespread uplift, faulting, and erosion near the end of Devonian time in much of the Great Basin region.

INTERVAL A

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

CALIFORNIA

Rocks assigned to interval A in California include parts of several formations. Near Taylorsville (pl. 1, Calif. loc. 156), Kinderhook brachiopod faunas from a

blue-gray calcareous slaty mudstone lens in the layered chert and slate of the upper member of the Peale Formation (McMath, 1966), have been identified by G. A. Cooper (written commun., 1965). In the Taylorsville area, the Peale is locally underlain by the Taylor Metaandesite of Mississippian age (Clark and others, 1962; Diller, 1908, p. 84; McMath, 1966), assigned to interval A. South of Taylorsville, Mississippian rocks are absent in the northern Sierra Nevada, where rocks of probable Silurian age are unconformably overlain by rocks of Mesozoic age (pl. 1, Calif. loc. 155; Clark and others, 1962). In the northern Sierra Nevada, a Carboniferous age for the Calaveras Formation and related Blue Canyon Formation was assigned by Turner (1893, 1897) and Lindgren (1896, 1900). Much of what was called Calaveras by these authors has been reassigned to the Shoo Fly Formation (Turner, 1893), of Silurian age (Clark and others, 1962). In Calaveras County, only Permian and probable Triassic faunas have been recovered from both the western and eastern belts of the restricted Calaveras Formation (Clark, 1964). This area has been tentatively assigned a zero Mississippian isopach value even though much of the Calaveras remains undated. In southern Mono County (pl. 1, Calif. loc. 157; Rinehart and Ross, 1964) a point of zero Mississippian has been interpreted between the Mount Baldwin Marble and underlying Silurian and older rocks in the Mount Morrison pendant. Kinderhook rocks are thought to be missing in the northern Inyo Mountains (pl. 1, Calif. loc. 159; Nelson, 1966) and in the western part of the Last Chance Range (pl. 1, Calif. loc. 153a) in northern Inyo County, where the Perdido Formation of Osage, Meramec, and Chester age unconformably overlies Devonian and Silurian rocks (Stewart and others, 1966, p. D25).

In southern Inyo County the lower part of the Tin Mountain Limestone, which "consists of dark-gray limestone in beds 2 to 6 inches thick, separated by much thinner beds of light brownish gray to pale-red shale" (McAllister, 1952, p. 20), is included in interval A with confidence. Coral and brachiopod faunas reported by McAllister (1952, 1956, 1974), Hall and MacKevett (1962), and Langenheim and Tischler (1960) are of Early Mississippian age. Conodont microfaunas from the southern Funeral Mountains (pl. 1., Calif. loc. 152) indicate a probable late Kinderhook age for at least the lower 205 feet of the formation (J. W. Huddle, written commun., 1965; McAllister, 1974). This figure is double the thickness shown on the interval A isopach map (pl. 3*A*) at this locality, and thus interval A isopachs in this area are minimums.

Farther south, in eastern San Bernardino County, the lower part of the Crystal Pass Limestone has been tentatively assigned to interval A. Recent conodont determinations, however, indicate that these rocks are probably of latest Devonian age (table 3), and thus interval A isopachs in this area are maximums. All of the Crystal Pass Limestone in the Nopah Range (pl. 1, Calif. loc. 108) is included in interval A with question.

In southwestern San Bernardino County an undetermined thickness of the lower part of the marblized Furnace Limestone (pl. 1, Calif. loc. 112) is included in interval A inasmuch as faunas recovered from the lower 1,000 feet in Furnace Canyon in the San Bernardino Mountains suggest an Early Mississippian age. (See discussion on p. 282.) Also, a part of the metasedimentary rocks of the Oro Grande Formation arbitrarily has been included in interval A as well as in all other intervals (Dibblee, 1960).

NEVADA

Thick sequences of interval A in Elko County are well documented faunally. Smith and Ketner (1968, 1975) have shown that in the Pinon Range the Webb Formation of interval A overlaps a segment of the Roberts Mountains thurst and rests unconformably on Devonian and Ordovician rocks of both the siliceous- and carbonate-assemblage units. In the southern part of the Carlin–Pinon Range area, the lower part of the Chainman Formation rests unconformably on these Devonian and Ordovician rocks, but to the north, the Chainman and Diamond Peak sequence is conformable on the Webb Formation (pl. 1, Nev. loc. 1).

In the northern part of the Pinon Range the Webb Formation is made up largely of siliceous mudstone and claystone with minor sandstone and limestone lenses; at its most southerly exposure, however, the formation is limestone. Claystone and barite nodules occur at the top of the formation which is at least 800 feet thick in the northern Pinon Range (Smith and Ketner, 1968, p. I9). Conodont faunas from the Webb Formation containing the diagnostic Kinderhook association of *Siphonodella obsoleta* and *Gnathodus* aff. *G. punctatus* were reported by W. H. Hass, J. W. Huddle, and D. L. Clark (in Smith and Ketner, 1968, p. I11, 1975, p. A37).

The lower part of the Chainman Shale–Diamond Peak Formation undifferentiated on the west slope of the Pinon Range consists largely of soft gray shale, some calcareous, interbedded with siltstone and minor sandstone. Megafaunas consisting of brachiopods, corals, and the diagnostic ammonoid genus, *Protocanites*, indicate a late Kinderhook age (Gordon and Duncan, 1961). Gordon (in Brew, 1971) considered the "so-called Chainman in this area * * * [the] temporal equivalent of roughly the upper two thirds of the Joana Limestone of the Ely district." A shaly limestone at Swales Mountain (pl. 1, Nev. loc. 21) with a minimum thickness of 650 feet contains a Kinderhook microfauna of the same age as that found in the Webb Formation. These shaly limestones, the *Camp Creek* sequence, seem to be gradational with Devonian(?) limestones of the Roberts Mountains Formation (Ketner, 1970b).

The thick (5,000 feet minimum) Chainman Formation in the Adobe Range (pl. 1, Nev. loc. 22), which is composed of about 80 percent chert-grain sandstone and 20 percent shale (K. B. Ketner, written commun., 1969; Ketner, 1970a), is probably in part correlative with the Webb and Chainman of the Pinon Range.

The unfossiliferous allochthonous argillite unit of Lee Canyon (Ketner and Smith, 1963a; Smith and Ketner, 1968, 1975), which is estimated to be "about a mile thick," may be in part of Kinderhook age.

A thick incomplete interbedded shale and sandstone and minor limestone sequence in the Windermere Hills (pl. 1, Nev. loc. 24) was reported by Oversby (1969) to contain the Kinderhook conodonts, *Siphonodella* spp., *Polygnathus* sp., and *P. perplanus*.

South of that locality, but also in the Windermere Hills, Oversby (1969) measured a thick (about 1,500 feet) section of impure limestone which he provisionally assigned to the Joana Limestone. The upper beds contain the Kinderhook conodont assemblage, *Siphonodella obsoleta* and *Gnathodus punctatus* which is also common in the Webb. Thorman (1970) reported Kinderhook microfaunas from both the base and top of a series of interbedded argillaceous limestone, siltstone, and limestone, at least 1,745 feet thick, in the northern Pequop Mountains (pl. 1, Nev. loc. 10).

One other clastic sequence tentatively assigned to interval A in Elko County is the *Waterpipe Canyon Formation* of Kerr (1962) which Ketner (1970b, p. D20) considered equivalent to the Chainman Shale. The *Waterpipe Canyon* (pl. 1, Nev. loc. 3) consists of at least 700 feet of unfossiliferous black argillaceous quartz siltstone interbedded with pebbly sandstone which lies unconformably on Silurian and Ordovician rocks.

Areas of zero interval A on the isopach map (pl. 3-A) in northeastern Elko County are attributed to (1) faulting, (2) pre-Tertiary beveling, and (3) complex facies and lack of faunal control. For example, thrusting in the East Humboldt Range (pl. 1, Nev. loc. 8) brings an incomplete section of Mississippian coarse clastic rocks, here assigned to the Upper Mississippian, directly over high-grade metamorphic rocks (Snelson, 1957, p. 115–117). Pre-Tertiary beveling and faulting are present in a drill hole (pl. 1, Nev. loc. 18) in which Ordovician rocks were recovered immediately below Tertiary sediments. Complex facies, lack of faunal control, and thrusting in the Pequop Mountains (pl. 1, Nev. loc. 9) were noted by Sadlick (1965) and Thorman (1962, 1970) who reported Chainman Shale structurally overlying questionable Pilot Shale with no intervening Joana in the southern part of the northern Pequop Mountains (Nev. loc. 9). Immediately northeast of locality 9 they reported 350 feet of Joana (Nev. loc. 9a) and, still farther north, possibly more than 1,700 feet of argillaceous Joana Limestone (Nev. loc. 10). Sadlick (1965, p. 42) pointed out the possibility that some of his *Needle Siltstone Member* at the base of the Chainman at Nevada locality 9 may be "an older pre-Chainman siltstone that may be a clastic facies of the Joana Limestone." This thick siltstone has been assigned entirely to interval C in this report, however, on the basis of a possible Meramec flora reported about 500 feet above the base.

In Eureka County and along the adjacent western border of White Pine County, the lower parts of two Chainman Shale facies, the allochthonous *Black Point facies* and the autochonous *Water Canyon facies* (Brew, 1961b, 1964), are included in interval A (pl. 15, col. 119) without conclusive faunal evidence. Gordon (in Brew, 1971) and Nolan, Merriam, and Williams (1956) have suggested that these rocks may all be much younger than the Chainman in the Pinon Range and may belong entirely to interval C. (See thickness discussions and alternate interpretation of table 5.) The lower part of the *Black Point facies* in the central Diamond Mountains (pl. 1, Nev. loc. 227) consists of about 2,000 feet of more than 70 percent mixed claystone and siltstone, more than 20 percent sandstone, and about 2 percent chert and quartzite pebble and cobble conglomerate (Brew, 1961a, 1964). The Chainman unconformably overlies approximately 200 feet of dark-gray crinoidal limestone of the Joana Limestone which is included in interval A. In the southern Diamond Mountains, the *Black Point facies* is slightly thinner (Nev. loc. 226; Brew, 1964, p. 58–59), and on the western slope of the Diamond Mountains it is underlain unconformably by as much as 400 feet of Joana Limestone (Nolan and others, 1956; Brew, 1961a) which is assigned to interval A. Thin shaly limestone and limy shale beds assigned to the Joana Limestone in the Eureka area and in the western Pancake Range (Nev. loc. 51) contain Kinderhook faunas in the basal beds and have been assigned entirely to interval A (Nolan and others, 1956, p. 55; Gordon, in Brew, 1971).

The lower 500 feet of *Water Canyon facies* of the Chainman Formation on the west side of Newark Mountain (Nev. loc. 55) consists predominantly of claystone and siltstone with an estimated 10 percent sandstone and minor chert pebble conglomerate (Brew, 1964, p. 54–55).

Other outcrops of Chainman Formation in Eureka County tentatively assigned to interval A include (1) about 200 feet of green siliceous mudstone, black shale, and gritty sandstone which lies between two alaskite sills at Devils Gate (Nev. loc. 235; Merriam, 1963), (2) a small outcrop of about 3 feet of undated conglomerate in the southern part of the Tuscarora Mountains (Nev. loc. 230), and (3) a fault-bounded pebbly sandstone and shaly sandstone sequence in westernmost central Eureka County east of Cortez which Harold Masursky

(in Roberts and others, 1967, p. 37) reported may be of Mississippian age. This outcrop lies within the eastern edge of the Antler highland as defined for succeeding intervals. The rocks are tentatively assigned to interval A because of lithologic similarity to the lower part of the Chainman Formation and because the highland is postulated to have first appeared during latest Devonian time and was probably much less extensive and had much less relief in Kinderhook time than in later times.

In central Nye County, an allochthonous sandy limestone, which is thrust over a Devonian "transitional assemblage" in the southernmost Hot Creek Range (Nev. loc. 160; Kleinhampl and Ziony, 1967), contains Kinderhook conodonts *Siphonodella* and *Pseudopolygnathus* (J. W. Huddle, written commun., 1964) and Early Mississippian calcareous foraminifer, *Septaglomospiranella primaeva* (Chernysheva). Huddle (J. I. Ziony, written commun., 1967) also has recovered Kinderhook conodonts 350 feet above the base of the limestone of Dobbin Summit in the Monitor Range which is about 1,000 feet thick in northern Nye County (Nev. loc. 328; Kleinhampl and Ziony, 1967). The limestone of Dobbin Summit, lying with no obvious break on Devonian limestone below, consists of black, fine-grained, thin-bedded limestone laminated with quartz-silt and quartz-sand streaks interbedded with minor sandstone and limestone-and-chert pebble conglomerate (J. I. Ziony, written commun., 1967). The lower contact may be a low-angle fault.

In southern Nye County an estimated 1,500 feet of argillite and subordinate quartzite and conglomerate of units D, C, and the upper part of B of the Eleana Formation (Poole and others, 1961), and an unspecified thickness of silty shale of the *Meiklejohn Formation* (of former usage) are included arbitrarily in interval A. Most argillite and quartzite units in the Eleana Formation contain sparse plant fossils, but preservation is too poor for identification, and invertebrate faunas useful for dating have not been found in strata assigned to interval A. Units B, C, and D of the Eleana Formation are recognized on the Nevada Test Site (Nev. loc. 151) only in the southern Belted Range (Ekren and others, 1971). The Narrow Canyon Limestone, where recognized, is assigned to interval A in southern Nevada. The Early Mississippian conodont *Siphonodella* is present in a limy quartzite at the base of the Narrow Canyon Limestone, which consists of platy brown silty and clayey limestone and minor limy siltstone and quartzite in the Spotted Range (Nev. loc. 137; F. G. Poole, oral commun., 1970). F. G. Poole (oral commun., 1970) correlated the Narrow Canyon Limestone with a widespread Kinderhook unit of similar lithology in the upper part of the Pilot Shale of the eastern Great Basin. Above the Narrow Canyon, the lower part of the Mercury Limestone also may contain some Kinderhook beds (F. G. Poole, oral commun., 1972), but, if present, these are thin and have been included in interval B. A thick section of very cherty Monte Cristo Limestone exposed in the Last Chance Range (Nev. loc. 156) has been assigned entirely to interval B, but the lowermost part of the sequence probably contains some strata of Kinderhook age and the upper part possibly contains some rocks of Meramec age (F. G. Poole, oral commun., 1970). The sequence has both lithologic and faunal similarities to the Tin Mountain Limestone in the southern Funeral Mountains (Calif. loc. 152).

Interval A in eastern Nevada generally is made up of the lower 50–100 feet of the lower massive crinoidal member of the Joana Limestone where recognized (Langenheim, 1960; Chilingar and Bissell, 1957; Stensaas and Langenheim, 1960; Westgate and Knopf, 1932). Conodont faunas from upper beds of the massive crinoidal member at Ward Mountain (Nev. loc. 59) in the southern Egan Range indicate a Kinderhook age for these beds (Huddle, in Brew, 1971); thus, interval A thicknesses indicated on plate 3-A are minimums. Langenheim (1960, p. 77) showed that at Lund (Nev. loc. 58) in the southern Egan Range the lower massive crinoidal member of the Joana may be as much as 330 feet thick.

In the northern Egan Range (Nev. loc. 67), an Osage age for the entire Joana is suggested by megafaunas reported by Fritz (1957, 1960), and at that locality interval A tentatively is considered to be absent.

In Lincoln County, Nev., interval A is represented by the lower part of the Joana Limestone in the Pahranagat Range (Nev. loc. 103; Reso, 1963) and the southern Egan Range (Nev. loc. 107; Kellogg, 1960), by an unnamed limestone of Mississippian age (Tschanz and Pampeyan, 1970) in the Golden Gate Range (Nev. loc. 109), and by an arbitrary thickness of rocks assigned to the thick Monte Cristo Limestone in the Mormon Mountains (Nev. loc. 105) and Meadow Valley Mountains (Nev. loc. 108). The thicknesses of interval A at these localities are probably minimums.

Within Clark County, all but the upper part of the Crystal Pass Limestone has been assigned to interval A with question. Brenckle (1970, 1973) has recorded a foraminiferal zone 7 microfauna (late Kinderhook to earliest Osage) 55 feet below the top of the Crystal Pass Limestone in the Arrow Canyon Range (near Nev. loc. 126). Other foraminiferal collections from the type locality of the Crystal Pass in the Las Vegas Range have indicated only a Famennian (Late Devonian) through Tournaisian (Early Mississippian) age range. The Crystal Pass Limestone, as at present defined in southeastern Nevada, probably contains the Devonian-Mississippian boundary (F. G. Poole, oral commun., 1970; Brenckle, 1970, 1973). In the Arrow Canyon and

Las Vegas Ranges, this boundary may be marked by a quartzite or a sandy zone about 60 feet below the top of the formation (F. G. Poole, oral commun., 1970); thus, thicknesses shown for interval A in Clark County are maximums. The unconformity between the light-gray medium-bedded Crystal Pass Limestone and the dark-gray thin-bedded Dawn Limestone Member of the Monte Cristo in these areas (Langenheim and others, 1962) lies close to the Kinderhook-Osage time boundary (Brenckle, 1970, 1973) as indicated in table 6.

WESTERN UTAH

Units assigned to interval A in northwestern Utah include the Fitchville Formation and the lower part of the Madison Limestone in the Lakeside Mountains (Utah locs. 120, 124; Young, 1955; Doelling, 1964), the Promontory Mountains (Utah loc. 102; Olson, 1960), the Sheeprock Mountains (Utah loc. 121; Cohenour, 1959), and the Dugway Range (pl. 1, Utah loc. 122; Staatz and Carr, 1964). The lower part of the *Gardner Dolomite* (of former usage), as used locally by Rigby (1952, 1958) in the Stansbury Mountains (Utah locs. 111, 112) and in the Selma Hills north of the East Tintic Mountains (Utah loc. 206) and by Bullock (1951) in the Lake Mountains (Utah loc. 205), and the Jefferson(?) Dolomite, as used by Baker (1947) in the Wasatch Range and by Gilluly (1932) in the southern Oquirrh Mountains (Utah loc. 117), also are assigned to interval A.

The Fitchville Formation generally is composed of blue-gray to light-gray and pink commonly medium-bedded limestone and dolomite with minor "bottle-green" chert. In the Tintic district, southern Lakeside Mountains, and the area of the Devonian Stansbury anticline — where Mississippian rocks are underlain by Cambrian, Ordovocian, and Silurian formations (pl. 2) — the basal few feet of the Fitchville consists of a conspicuous thin calcareous quartzite or a sand-streaked limestone that locally contains quartz pebbles. The top of the Fitchville in most areas is marked by "the curley limestone," a stromatolitic limestone a few inches to several feet thick (Morris and Lovering, 1961, p. 84–85; Proctor and Clark, 1956). Corals associated with some brachiopods are present in the Fitchville of the East Tintic Mountains and were studied by Helen Duncan and Mackenzie Gordon, Jr. (in Morris and Lovering, 1961, p. 85–87) who assigned the entire formation an Early Mississippian age. They stated further that "faunal evidence points to the presence of the systemic boundary [Devonian-Mississippian] in the nodular argillaceous limestones that form the upper part of the Pinyon Peak [below the Fitchville]." Conodont faunas from the Fitchville of the East Tintic area, however, indicate a Kinderhook age for all but the basal 130 feet of the formation, from which Clark and Ethington (1967) reported the Late Devonian forms *Polygnathus styriacus* and *Bispathodus aculeatus*. Zone 7 foraminiferal faunas illustrated by Clark (1954, pl. 2) support a late Kinderhook to early Osage age for the upper part of the Fitchville. Late Devonian conodont faunas, identified by J. W. Huddle and C. A. Sandberg, from the lower part of the Fitchville have been discussed under "Paleogeology" in reference to the Pinyon Peak Limestone (p. 281). The entire Fitchville Formation, however, is included in interval A in this report. Morris and Lovering (1961, p. 87–88) noted that "Beds similar lithologically and containing identical fossils have been recognized by the writers in the Stansbury and Oquirrh Mountains and the Wasatch Range, but were assigned to the Jefferson(?) dolomite [by Gilluly (1932) and Calkins and Butler (1943)]."

In central Utah the lower part of the "Gardner Formation," as described by Clark (1954), is lithologically identical to the Fitchville Formation, and extensive faunas listed by Clark (1954, p. 7) support an Early Mississippian age for these beds.

Rocks of interval A are shown to be missing over much of northwestern Utah (pl. 3-A). Faunal evidence proving their absence, however, is not everywhere conclusive. Helen Duncan (in Morris and Lovering, 1961, p. 92) correlated the Madison of the Gold Hill district (Utah loc. 115) with the Gardison of interval B. Faunas from the Joana Limestone of the north Silver Island Range (Utah loc. 113), however, indicate that some of that formation may be Kinderhook in age (Schaeffer, 1960).

Interval A constitutes the lower part of the Joana Limestone where present in the Confusion Range–Needle Range area of west-central Utah. Dark-gray thick to massively bedded variably cherty medium- to coarse-grained crinoidal limestone characterizes this part of the Joana which contains calcareous foraminiferal faunas (Zeller, 1957) in addition to diagnostic coral and brachiopod faunas (Ogden, 1951; Chilingar and Bissell, 1957; Langenheim, 1960).

In southwestern Utah, the lower part of the Whitmore Wash Member of the Redwall Limestone or its correlative is included in interval A where recognized, as in a well in Washington County (Utah loc. 278), in the Beaver Dam Mountains (Utah loc. 276; Reber, 1952; R. L. Langenheim, Jr., written commun., 1967; McKee and Gutschick, 1969, p. 26), and in the Star Range (Utah loc. 301; Baer, 1962). At these localities, thickbedded or massive dolomite or dolomitic limestone makes up the interval. In the Wah Wah Mountains (Utah loc. 303), 80 feet of unnamed thick-bedded limestone has been included arbitrarily in interval A.

THICKNESS TRENDS

An unknown thickness of interval A is present west of the Antler highland in northern California (pl. 3-A). The highland itself as shown on plate 3-A is devoid of

recognizable interval A with the exception of an undated isolated fault sliver in Eureka County, Nev., just east of Cortez (Nev. loc. 236). Possibly, this locality lies at the eastern margin of the highland but isopachs are not drawn to show this because the rocks are undated. A large area of missing or thin (less than 100 feet thick) interval A extends south-southwestward from the area of the Wendover highland in northwestern Utah into Lincoln County, Nev., along the inferred Ely arch.

Thick sequences are present in north-central Nevada in the general vicinity of the Chainman–Diamond Peak trough, in south-central Nevada in the Eleana trough, and in central Utah in the Stansbury Mountains (pl. 3-A). Thicknesses may be as much as 2,000 feet or more in both the southern and northern parts of the Chainman–Diamond Peak trough. In the northern part, the Webb Formation and Chainman Shale of the Pinon Range are known to constitute thick interval A deposits as do related sequences at Swales Mountain, in the Windermere Hills, and in the northern Pequop Mountains. In the southern part of the Chainman–Diamond Peak trough, the lower 2,000 feet of poorly dated Chainman Shale in the vicinity of the Diamond Mountains that is assigned to interval A on plate 3-A may instead belong to interval C (as shown in table 5). If so, the area of the Webb-Chainman trough that excludes the detrital rocks of the Diamond Mountains area is restricted, as illustrated in figure 74, to include only the northern part of the Chainman–Diamond Peak trough in Eureka and Elko Counties.

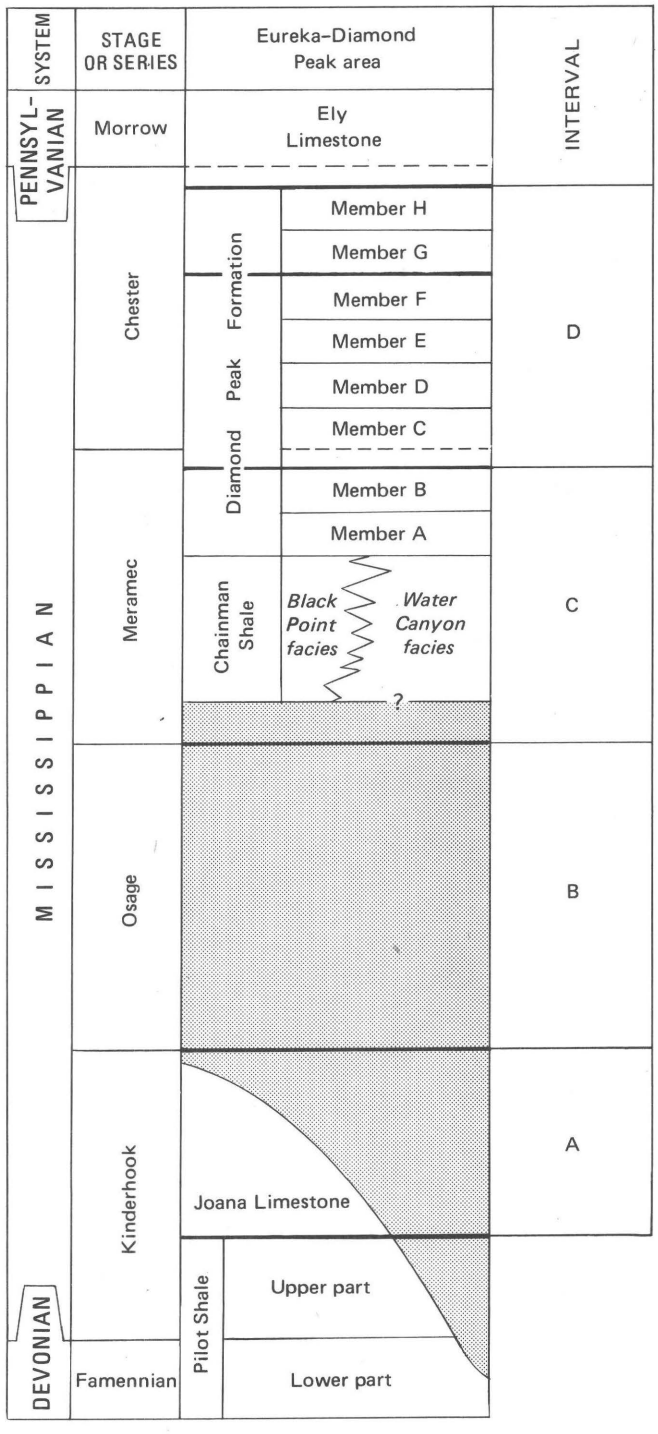
Interval A may have a maximum thickness of more than 1,500 feet in the Eleana trough in south-central Nevada (pl. 3-A), but diagnostic faunal evidence is lacking.

Thin interval A sequences which nowhere exceed 300 feet are present in much of southeastern California. Thicknesses shown on plate 3-A are minimums in northern Inyo County near the Antler highland and maximums farther southeast.

Interval A is notably thin, less than 100 feet, in most of western Utah. Isolated thicker sequences, as much as 650 feet, are postulated in two areas of north-central Utah in Tooele and Box Elder Counties in the general vicinity of the proto-Oquirrh basin. Interval A thins eastward to zero near the Wasatch line in central Utah.

TABLE 5. — *Alternate time-stratigraphic interpretation (compare with plate 15, column 121) of correlation of Mississippian rocks in the Eureka–Diamond Peak area, northeastern Nevada*

[Effect on isopachs for intervals A, B, and C is shown in figures 74, 75, and 77. Interval D is not affected. Modified from Gordon (in Brew, 1971)]



LITHOFACIES TRENDS

Interval A as defined herein for the Great Basin region is dominated by thin sequences of relatively pure shallow-water marine limestone which extend from central Utah westward to the northern Monitor Range in Nye County, Nev., northward into Idaho, and southward through much of southern California (pl. 3-B). The limestones range in texture from coarse-grained crinoidal in the northern and eastern parts of the basin to micritic in the southern part.

Argillaceous limestones are reported from California in scattered localities along the eastern edge of the Antler highland in the Inyo Mountains (Calif. loc. 101)

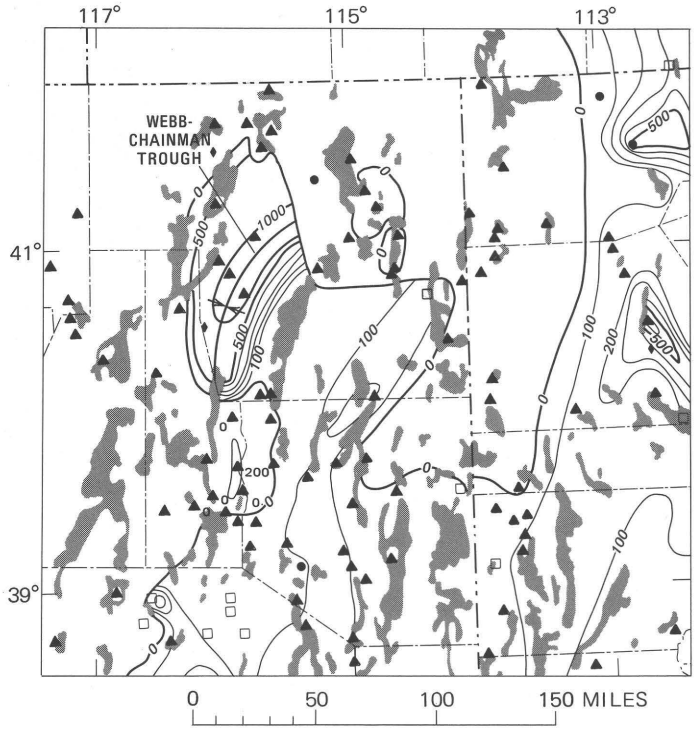


FIGURE 74. — Alternate isopach map of interval A showing thickness in feet in the Eureka –Diamond Peak area, Nevada, and the location of the Webb-Chainman trough when the Chainman Shale is assigned to interval C rather than to intervals A and B. Triangles, dots, diamonds and squares are control points identified on plate 1. Shaded areas are outcrops of pre-Mississippian rocks.

and from Nevada in the central Pancake Range in northeastern Nye County (Nev. loc. 331), at Swales Mountain in the Independence Mountains in western Elko County (Nev. loc. 21), and in the Windermere Hills and the northern Pequop Mountains in eastern Elko County (Nev. locs. 10, 25). Sandy limestone is recorded from a well in northwestern Utah (Utah loc. 104) and from an incomplete section in Nye County (Nev. loc. 160). Fine-grained detrital rocks consisting of claystone and siltstone with minor sandstone and some limestone are present in the Pinon and Adobe Ranges in western Elko County, in the Windermere Hills and Pequop Mountains of eastern Elko County, and in the isolated outcrop (near Cortez, Nev.) which lies either on the Antler highland or along its eastern margin.

Chert forms more than 10 percent of the interval east of the Antler highland (outlined by the zero isopach in central Nevada; pl. 3-A) only in the Promontory Mountains and the Sheeprock Mountains of northern Utah. Elsewhere chert is present generally as scattered nodules and layers in the ubiquitous thick- to massive-bedded interval A limestone and dolomite.

Interval A volcanic activity in northern California is not as firmly documented as that of intervals C and D in central Nevada. The mudstone from which Kinderhook fossils have been identified is, according to Vernon McMath (written commun., 1965), "a lens in the base of a dominantly nonvolcanic unit * * * which is characterized by ribbon chert and slate * * * . All units above and below are pyroclastic and so far undated." The enclosing volcanic rocks and the Mississippian age of the underlying Taylor Metaandesite has led me to speculate that there probably was Kinderhook volcanism in this area.

SOURCES OF SEDIMENT, ENVIRONMENTS OF DEPOSITION, AND CLIMATE

A western deeply weathered terrigenous source area, consisting largely of lower Paleozoic eugeosynclinal and transitional rocks, contributed the detritus which comprises the mudstone, siltstone, and sandstone of the Webb Formation and Chainman Shale in western Elko County, Nev. (Roberts and others, 1958; Ketner, 1970b). These shales and sandstones, many of which are calcareous, may have been deposited in moderately deepwater marine environment (pl. 11, fig. 1); probable radiolarians were reported from the Webb by Smith and Ketner (1968, p. 10, 1975, p. A35). The calcareous shales and siltstones are replaced southward by limy shales and limestones within the Webb Formation and then farther south by the shallow-water marine carbonate rocks of the Joana Limestone. In the southern Independence Mountains, western Elko County, local lower Kinderhook turbidity current deposits which contain no eugeosynclinal detritus were described by Ketner (1970b). The turbidites are interpreted to have been deposited along the western edge of the miogeosyncline in a moderately deep marine environment. Westward-moving turbidity currents derived from the east may have been triggered by early pulses of the Antler orogeny.

Large areas of interval A carbonate rocks in the central part of the Great Basin indicate extensive uninterrupted shallow-water marine shelf deposition. In the central and eastern parts of the Great Basin region most of the limestones are relatively pure, thick bedded, and rich in crinoidal detritus and indigenous corals, brachiopods, and foraminifers, suggesting that interval A deposition was below wave base in warm marine waters of normal salinity, and of shallow to moderate depth. Aphanitic limestone or micrite, much of it with a peloidal texture formed by rounded fragments of micrite and unilocular calcareous foraminifers (McKee and Gutschick, 1969, p. 555) is common in interval A in southern Nevada and southern California. Blue-green algae in these micritic limestones suggest they may have been deposited in a very shallow subtidal, at times perhaps intertidal, marine environment. The biostromal "curley limestone" at the top of the

Fitchville Formation in northern Utah probably was formed in an intertidal or lagoonal environment which extended over 600 square miles of northern Utah in late Kinderhook time (Clark, 1954, p. 29–30). No true basinal carbonate facies have been recognized in interval A. Very little detritus appears to have been shed westward from the craton into the shallow seas that covered large parts of the central Great Basin region during this interval.

South of the Pinon Range in Eureka and southern Elko Counties in central Nevada, none of the detrital sequences that have been assigned to interval A are dated. Shallow epicontinental seas in which the Joana Limestone was deposited may have covered the entire area during most of the interval. The thick shales and siltstones shown in this position on plates 3-A, 4-A, and 5-A, however, would require a western land source — a segment of the Antler highland. To the south, in western Nye County, dated interval A limestones are sandy in part, which suggests a nearby landmass, as does the fine-grained detritus of the lower part of the Eleana and *Meiklejohn* Formations in southern Nevada and eastern California. Most of the Eleana Formation of southern Nye County appears to have been deposited under basinal conditions in moderately deep waters. Many quartzites in units B, C, and D assigned to interval A have features characteristic of turbidites (F. G. Poole, in Ekren and others, 1971, p. 22).

Probable Early Mississippian volcanic activity in northern California at Taylorsville is recorded in the Taylor Metaandesite and the overlying Peale Formation. A Kinderhook fauna from the upper part of the Peale Formation indicates that some of the slate and chert of that formation were deposited in a eugeosynclinal or island arc environment during interval A time.

PALEOTECTONIC IMPLICATIONS

Interval A volcanism similar to volcanism in earlier periods of the Paleozoic probably took place west of the Antler orogenic belt and, therefore, west of the contemporaneous continental margin in what is now northern California.

No record of interval A rocks exists in the broad area between Taylorsville in northern California and Cortez in north-central Nevada, but it seems reasonable to assume that the eastern part of this area was occupied by the Antler highland (pl. 10, fig. 1), and the western part by the Ancestral Pacific Ocean. It is also possible that some of the volcanic activity recorded in undated greenstones and metavolcanics of the "Leach Formation" took place near the western edge of the Antler highland at this time.

The Antler orogenic belt came into existence shortly before interval A time and was characterized by uplift and erosion of lower Paleozoic transitional and eugeosynclinal facies rocks and telescoping of these facies along extensive low-angle faults with major horizontal displacements. The major thrusting has been dated as latest Devonian in northern Nevada in the Pinon Range (Smith and Ketner, 1968) and as possibly earliest Mississippian in the southern Independence Mountains (Ketner, 1970b).

Uplift of the Antler belt was accompanied by subsidence of the Webb-Chainman trough (fig. 74) during interval A time along the eastern edge of the highland in western Elko County, Nev., where a minimum of 1,000 feet of well-dated calcareous shales and siltstones was deposited (pl. 9-G, sec. *R*–*R'*). Siltstones of the Chainman Shale in Eureka County were deposited during either interval A or C.

Subsidence of the Eleana trough also probably began in interval A time, although the rocks that might contain this record remain undated.

Relatively thin calcareous shales and siltstones of the upper part of the Pilot Shale, which are excluded from interval A, were deposited over a large area of the central part of the Great Basin (pl. 2) at the same time that shale and limestone of the Webb Formation and *Camp Canyon Member* of the Chainman of north-central Nevada and the Narrow Canyon Limestone of southern Nevada and southeastern California were laid down. Deposition of the Pilot Shale ended during Kindherook time and was followed by a period of uplift and erosion on most of the shelf area of the Great Basin.

As the epicontinental seas again flooded the shelf area, pure bioclastic marine limestone of the Joana Limestone of interval A was deposited on an erosional surface developed within the lowermost part of the Joana or on the Pilot Shale and older rocks exhumed during emergence. Westerly derived detritus which produced the Pilot Shale was no longer available in the eastern part of the region, but, to the west, deposition probably continued without major interruption in the troughs adjacent to the orogenic belt.

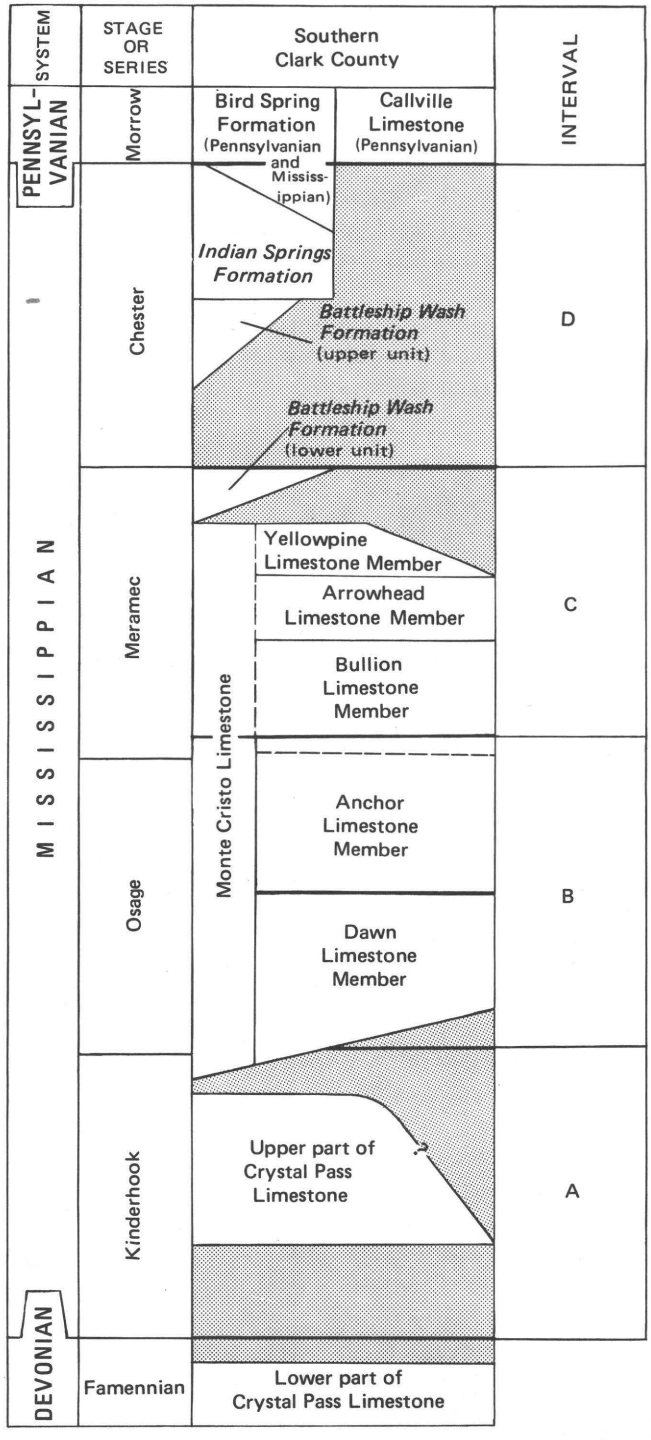
Well-dated shallow marine, sandy in part, limestone in north-central Nye County is present (Nev. loc. 328) between and west of the Webb-Chainman and the Eleana troughs. This limestone may have been deposited in an embayment along the eastern side of the highland or possibly in a shallow seaway which may have divided the Antler highland into northern and southern segments. The sandy nature of the limestone indicates the presence of a nearby terrigenous source, but one which contributed much less detritus than the highlands adjacent to the Webb-Chainman and Eleana troughs.

Local interruption in sedimentation near the end of Kinderhook time is indicated in southern Nevada and

southeastern California by an erosional hiatus at the top of the *Crystal Pass Limestone* (table 6; Langenheim and others, 1962). The top of interval A also marks the end of deposition of the pure massive crinoidal member of the Joana Limestone in the central Great Basin. The Joana of the succeeding interval is impure and thin bedded indicating the introduction of either cratonic or orgeonic detritus which was not available during interval A time. Local crustal instability in the vicinity of the Wendover highland at the end of interval A is suggested by the fact that Kinderhook limestones which probably covered parts of the highland were removed by erosion before interval B deposition (pl. 9-G, sec. *R*–*R'*).

TABLE 6. — *Alternate time-stratigraphic interpretation (compare with plate 15, column 126) of relations between the Crystal Pass Limestone and the Monte Cristo Limestone*

[Based on Brenckle (1973) and F. G. Poole (oral connum., 1970). Isopachs on plate 3-A have not been adjusted to fit this interpretation]



Interval A thins to zero along the western edge of the craton from central Utah southwestward into southern California. The zero line shown on plate 3-A probably represents the easternmost transgression of the late Kinderhook sea onto a very low positive or neutral area (pl. 10, fig. 1).

PALEOGEOGRAPHIC INTERPRETATIONS

Much of the Great Basin region in eastern Nevada and western Utah was covered by shallow epicontinental seas during interval A time (pl. 12, fig. 1). The maximum transgression of the sea upon the broad cratonic lowland to the east probably occurred in late Kinderhook time and roughly coincides with the position of the Wasatch line.

On the western edge of the epicontinental sea, the Antler highland emerged as low hills in central Nevada and southern California. The low hills, probably deeply weathered (Ketner and Smith, 1963b), slowly were eroded by gentle gradient streams which carried mud, silt, and minor sand into the narrow restricted marine troughs along the eastern edge of the highland.

No record is preserved of detritus shed northwestward from the highland. The record of interval A in northern California is that of interbedded chert, slate, and volcanics — a eugeosynclinal assemblage — formed, perhaps, in an oceanic environment on the continental side of a volcanic island arc, within an inner arc basin.

INTERVAL B

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

SOUTHERN CALIFORNIA

Units assigned to interval B (approximately Osage time) include the upper part of the Tin Mountain Limestone and the lower part of the overlying Perdido Formation, the Dawn and Anchor Limestone Members of the Monte Cristo Limestone and the uppermost part of the underlying Crystal Pass Limestone, where recognized (pl. 15), and several unnamed metasedimentary sequences.

The upper part of the Tin Mountain Limestone is recognized in California in the northern Panamint Range (Calif. locs. 101, 158), Inyo Mountains (Calif. loc. 104), Darwin area (Calif. loc. 107), Argus Range (Calif. loc. 118), and southern Funeral Mountains (Calif. loc. 152). In all these areas (fig. 70), this part of the Tin Mountain consists of medium-gray thin- to medium-bedded cliff-forming, variably cherty limestone with silty or shaly partings. The limestone is micritic or calcarenitic, and it contains crinoidal debris and numerous corals and brachiopods, which in the northern Argus Range are "clearly related to [those] found in the upper part of the Lodgepole" (J. T. Dutro, Jr., in Hall and Stephens, 1963, p. 15).

The lower part of the Perdido Formation gradationally overlies the Tin Mountain Limestone, and the contact is commonly placed at or near the base of the lowermost bedded chert unit (McAllister, 1952, 1974; Hall and MacKevett, Jr., 1962). Gordon (1964, p. A2) reported an Early Mississippian coral fauna from the lower part of the Perdido which, in the southern Funeral Mountains, is bracketed by Kinderhook conodont and foraminiferal faunas below and Meramec conodont and foraminiferal faunas above (McAllister, 1974). In the northern Panamint Range (Calif. loc. 101), in the type area, 225 feet of the lower part of the Perdido was described (McAllister, 1952) as interbedded limestone, siltstone, silty limestone, and chert. In the southern Panamint Range (Calif. loc. 109), the Perdido is dominantly interbedded cherty limestone and chert, and it contains as much as 30 percent nodular and bedded chert (Hall and Stephens, 1962, 1963; Pelton, 1966, p. A21). In the Inyo Mountains (Calif. loc. 105), the lower part of the Perdido unconformably overlies Silurian and Devonian transitional facies rocks and consists of shale, siltstone, boulder to pebble conglomerate with an argillaceous-siliceous matrix, and minor chert (Ross, 1966, p. 59–60; and Ridley, 1970). Ross (1966) tentatively considered the Perdido here to be entirely of Late Mississippian age, but the lower 250 feet is included arbitrarily in interval B in this report. Pelton (1966, p. A17) described several hundred feet of varicolored siltstone and shale above the Devonian Lost Burro Formation on the east flank of the Last Chance Range (Calif. loc. 153b) that is included in interval B. Scott (1960) described a minimum of 300 feet of cherty and sandy limestone and dolomite, siltstone, sandstone, and shale with 60 feet of conglomerate consisting of dolomite clasts as much as 2 feet in diameter in a limy sandstone matrix in the Waucoba Springs area of the Inyo Mountains (Calif. loc. 154) which is assigned to interval B. Cherty fine- to coarse-grained limestone and marble in the northern Slate Range (Calif. loc. 148) are considered lithologically similar to the Tin Mountain Limestone and Perdido Formation (G. I. Smith, written commun., 1965) and therefore are assigned to interval B. The lower cherty part of an unnamed limestone, "younger than Tin Mountain," also has been included in interval B in the northern Panamint Range (Hunt and Mabey, 1966, p. 43–45).

The Anchor and Dawn Limestone Members of the Monte Cristo Limestone and the uppermost part of the Crystal Pass Limestone, where recognized, are included in interval B from the northern Nopah Range (Calif. loc. 108; Hazzard, 1954a) south to the northern Providence Mountains (Calif. loc. 110; Hazzard, 1954b) and at Old Dad Mountain east of the Soda Mountains in the southwest corner of the Ivanpah quadrangle (Calif. loc. 131; Barca, 1960). The Anchor Limestone Member is characterized by abundant chert and crinoidal debris and therefore is correlated with the Thunder Springs and the Mooney Falls Members of the Redwall Limestone in northern Arizona. The Anchor and Dawn Limestone Members and similar cherty limestone in southeastern California are assigned to interval B on the basis of lithologic correlation with the Arizona Redwall Limestone and the Monte Cristo Limestone in the Arrow Canyon Range, Nev.

In southwestern San Bernardino County, Calif., tentatively included in interval B, as well as in the other three Mississippian intervals, are metasedimentary rocks, particularly the marble and minor quartzite and phyllite, of the lower part of the Furnace Limestone (Calif. locs. 112, 113, 150; Dibblee, 1964a, b, c; Richmond, 1960; Woodford and Harriss, 1928; Merriam, 1954) and a part of the quartzite, schist, and marble of the Oro Grande Formation (Calif. locs. 120, 122, 141, 142; Bowen, 1954; Dibblee, 1960). The Early Mississsippian brachiopods "*Dictyoclostus*" and "*Spirifer*" aff. "*S.*" *centronatus* have been recovered from the lower 1,000 feet of the Furnace Limestone (Richmond, 1960).

South of the northern Providence Mountains, limestone and dolomite, in places altered to marble, as in the southern Marble Mountains and Ship Mountains (Calif. loc. 139; Hazzard, 1933; Clark, 1921), are thought to include Mississippian rocks. A somewhat similar, thin metamorphosed variably cherty calcite marble section is present in the northern Riverside Mountains (Calif. loc. 119; Hamilton, 1964). This marble section is assigned to intervals B and C on the basis of general lithologic similarity to a nearby unmetamorphosed section of 415 feet of Escabrosa Limestone (Miller, 1970; Miller and McKee, 1971) near Quartzsite (Ariz. loc. 157), from which a late Kinderhook(?) to early Meramec coral fauna, collected by F. G. Poole 10–15 feet above the base, was identified by W. J. Sando and Helen Duncan (written commun., 1967).

In northern California, part of the metavolcanic and metasedimentary rocks of the upper member of the Peale Formation near Taylorsville (Calif. loc. 156) is in-

cluded in interval B because Mississippian brachiopods found in the sequence have been recognized as Kinderhook and possibly earliest Osage in age (G. A. Cooper, written commun., 1965). The fossils occur in a lens about 50 feet thick, of blue-gray calcareous slaty mudstone interbedded with ribbon chert and slate (Vernon McMath, written commun., 1965). Sixty miles south of the Taylorsville area on the north fork of the American River (Calif. loc. 155), Triassic(?) chert breccias unconformably overlie the Shoo Fly Formation of Silurian age (Clark and others, 1962); Mississippian rocks apparently are absent.

NEVADA

Formations assigned to interval B in Nevada with some faunal support include the upper part of the Joana Limestone, where present, the Dawn and Anchor Limestone Members of the Monte Cristo Limestone in Clark County, the Mercury Limestone and limestone of Timpi Canyon (of F. G. Poole, oral commun., 1965) of northwestern Clark County and adjacent Nye County (Nev. locs. 137, 138), and the upper part of the Chainman Shale in the Pinon Range–Carlin area (Nev. loc. 1; Smith and Ketner, 1968, p. I14; Gordon, in Brew, 1971).

The upper part of the Chainman in the Pinon Range–Carlin area in north-central Nevada is made up of gray shale and minor sandstone, quartzite, and mudflow conglomerate. Osage faunas from these beds, reported by Gordon and Duncan (1961) consist of corals, brachiopods, and mollusks. Near Carlin late Kinderhook goniatites have been found about 200 feet above the base of the Chainman where it overlies the Webb Formation in beds here assigned to interval B. Also, the contact of the Chainman with the overlying Diamond Peak Formation is gradational, and, in places, the lower part of the Diamond Peak contains Osage faunas (J. F. Smith, Jr., oral commun., 1972). Where the Webb Formation is present, the overlying Chainman is assigned entirely to interval B; where the Webb is absent, an arbitrary thickness of the upper part of the Chainman is assigned to B (pl. 15).

The Dawn and Anchor Limestone Members of the Monte Cristo Limestone in southern Nevada have been accepted as Early Mississippian in age since Girty (in Hewett, 1931, 1956) first described faunas from the Goodsprings and Ivanpah areas in southern Clark County, Nev., and northeastern San Bernardino County, Calif. Recent work on calcareous foraminfers from the Arrow Canyon Range (Brenckle, 1970, 1973; Mamet and Skipp, 1970b) has shown that the Dawn Limestone Member contains microfaunas indicative of the lower and upper(?) Osage (foraminiferal zones 7, 8, and 9?).

Diagnostic foraminiferal faunas have not been recovered from the thick upper crinoidal part of the Anchor Limestone Member in the Arrow Canyon Range, but questionable middle Meramec faunas have been recovered from the overlying Bullion Limestone Member. Brenckle (1973), therefore, has assigned an Osage age to only the lower cherty part of the Anchor, an assignment which would reduce interval B thicknesses at Nevada localities 126 and 131 by about 400 feet, with a corresponding increase in interval C thicknesses. In this report, the entire Anchor Limestone Member is included in interval B because faunas are sparse and because the member has the same stratigraphic position and bears lithologic resemblance to the Thunder Springs Member (Osage) of the Redwall and the lower part of the Mooney Falls Member (Osage) of the Redwall Limestone in the western Grand Canyon area (McKee and Gutschick, 1969; fig. 7). The Osage-Meramec boundary based on conodonts has been placed within the Yellowpine Limestone Member overlying the Bullion Limestone Member in the Arrow Canyon Range (Pierce and Langeheim, 1974). This boundary seems too high as a middle Meramec foraminiferal fauna was recovered from near the base of the Yellowpine Limestone Member in the same area (Brenckle, 1973). Recent work (Brenckle and others, 1974), however, has shown that the Osage Series may contain somewhat younger strata than reported in the Mamet foraminiferal zonation (Mamet and Skipp, 1970a), thus tending to close the gap a little between the boundaries based on foraminiferal and conodont determinations.

A late Kinderhook to early Osage fauna (foraminiferal zone 7) has been found in the Arrow Canyon Range in a limestone bed 55 feet below the top of the Crystal Pass Limestone just above a thin quartz sandy interval (Brenckle, 1970). This thin sequence has been assigned, with reservation, to interval B in much of Clark County; Brenckle (1970) considered it to be late Kinderhook. (See "Interval A" discussion and table 6.) A very cherty unit in the Monte Cristo Limestone crops out in the Last Chance Range (Nev. loc. 156) according to F. G. Poole (written commun., 1965). This cherty unit is here correlated with the Anchor Limestone Member and is tentatively assigned to interval B. (See "Interval A" discussion.)

Part of the Rogers Spring Limestone (Longwell, 1921), which was mapped in eastern Clark County (Longwell and others, 1965), is included in interval B. The Rogers Spring Limestone has long been considered approximately equivalent to the Monte Cristo Limestone (Hewett, 1931).

The early Osage age of the upper part of the Joana Limestone in eastern Nevada is well established. Spencer (1917) first reported a "Madison" fauna when he defined the Joana in the Ely area. Early Mississippian megafossils were later reported by Westgate and

Knopf (1932, p. 20) in the Pioche district from the Bristol Pass Limestone, here considered equivalent to the Joana; elsewhere in eastern Nevada, similar fossils from the Joana were reported by Nolan (1935) and Stensaas and Langenheim (1960). Kinderhook to early Osage calcareous foraminiferal faunas were reported by Chilingar and Bissell (1957, p. 2262), Zeller (1957), and Woodland (1958). Recent work by Brenckle (1973) on the Joana at Ward Mountain southwest of Ely in the southern Egan Range (Nev. loc. 59) shows that in a 408-foot-thick section of Joana the lower 70–80 feet of crinoidal limestone contains no foraminifera, but that above this, calcareous foraminiferal zones 7 and 8, and possibly 9, are present representing late Kinderhook or early Osage to late(?) Osage time time. J. W. Huddle reported a *Siphonodella* (late Kinderhook) conodont zone at the top of the lower massive crinoidal member at Ward Mountain (Gordon, in Brew, 1971, p. 36). In the present report, all the thin-bedded locally impure limestones and all but the basal 50–100 feet of massive crinoidal locally cherty limestone of the Joana are included in interval B. In some places, notably at Ward Mountain, interval B thus includes about 50 feet of dated Kinderhook beds. In very thick sections of Joana, such as that at Sunnyside in the southern Egan Range (Nev. loc. 104), lowermost Meramec rocks may be present in the uppermost beds, but faunal evidence is inconclusive.

A faulted section of Joana-like limestone in the San Antonio Mountains (Nev. loc. 158; Kleinhampl and Ziony, 1967) in northwestern Nye County contains the Foraminifera, *Spinoendothyra* and *Tuberendothyra*, indicative of the upper Osage; these rocks are assigned to interval B.

In addition, limestone pebbles and cobbles from the middle of the upper part of the Permian Diablo Formation in the southern Toiyabe Range (near Nev. loc. 155) contain corals of Osage aspect (N. J. Silberling and Helen Duncan, written commun., 1959), probably indicating that limestones of interval B once were present farther west (Ketner, 1967, p. 237). The thin shaly limestone and limy shale beds assigned to the Joana in the Pancake Range (Nev. loc. 51) and Diamond Mountains area contain Kinderhook faunas in the basal beds and have been assigned entirely to interval A (Nolan and others, 1956, p. 55; Gordon, in Brew, 1971).

Other rocks assigned to interval B in Nevada include part of the volcanic rocks of the Inskip and Goughs Canyon Formations in Pershing and Humboldt Counties west of the Antler highland. The Goughs Canyon is reported to contain corals and brachiopods suggestive of both Osage and Meramec age (Hotz and Willden, 1964, p. 26–27). The upper parts of the two facies of the Chainman Shale in the Eurkea–Diamond Peak area are assigned to interval B (Brew 1961b, 1964). Possibly, the entire Chainman in this area belongs to interval C (table 5) as suggested by Nolan (1943) and Gordon (in Brew, 1971).

In the Nevada Test Site area (Nev. loc. 151) approximately 2,500 feet of argillite, quartzite, and conglomerate belonging to units E and F of the Eleana Formation, which contain only plant fossils and invertebrate trails, has been assigned to interval B arbitrarily (Poole and others, 1961, 1965), as has an unspecified thickness of the *Meiklejohn Formation* in southwestern Nye County. Reworked Early Mississippian fossils have been recovered from conglomeratic limestone beds in unit I of the Eleana Formation (Poole and others, 1965, p. A52–A53) in association with both Devonian and Late Mississippian forms, indicating that Devonian and Early Mississippian beds were uplifted and eroded in intervals C (Meramec) and D (Chester) time. The conglomerate, siltstone, and phyllite of the Grossman Formation in northern Elko County (Nev. loc. 12) is assigned to interval B because of its stratigraphic position below the Banner Formation of interval C, but there is no faunal evidence in the Grossman itself. As pointed out by Coats (1969, p. A24), however, "the clastic character and obvious derivation from the Valmy suggest that the formation postdates the beginning of the Late Devonian Antler orogency * * * it is therefore tentatively assigned to the Devonian or Mississippian."

An unspecified thickness of the argillite-quartzite unit in Lee Canyon of Ketner and Smith (1963a, p. B10) is assigned to interval B on the basis of lithologic similarity to the nearby Chainman Shale which contains Osage faunas.

WESTERN UTAH

In northwestern Utah, units assigned to interval B (pl. 15) include the Gardison Limestone of the East Tintic Mountains (Utah loc. 201) and southern Wasatch Range (Utah locs. 202, 203; Morris and Lovering, 1961; Crittenden, 1965a, b; Baker and Crittenden, 1961; Baker, 1964), the Madison Limestone of the southern Oquirrh Mountains (Utah loc. 117; Gilluly, 1932), Gold Hill district (Utah loc. 115; Nolan, 1935), and Crater Island (Utah loc. 108; Anderson, 1957). The upper part of the Madison Limestone has been assigned to interval B in the Lakeside Mountains (Utah locs. 120, 124; Young, 1955; Doelling, 1964), the Promontory Mountains (Utah loc. 102; Olson, 1960), the Sheeprock Mountains (Utah loc. 121; Cohenour, 1959), and the Dugway Range (Utah loc. 122; Staatz and Carr, 1964). The combined upper part of the Madison and the lower part of the overlying *Pine Canyon Limestone* (of former usage) make up interval B in the southern Stansbury Mountains (Utah loc. 112; Teichert, 1959). The combined lower part of the *Pine Canyon Limestone* and upper part

of the underlying Gardner Formation (considered Gardison equivalent by Morris and Lovering, 1961) constitute interval B in the Lake Mountains (Utah loc. 205; Bullock, 1951; Clark, 1954) and a part of the southern Wasatch Range (Utah loc. 111; Rigby, 1958). South of the Wasatch Range, in the Gilson Mountains (Utah loc. 227), Costain (1960) mapped approximately 350 feet of Gardison Limestone which is here included in interval B. The Gardison Limestone in the East Tintic Mountains consists of a lower thin- to medium-bedded nearly chert free fine- to coarse-grained fossiliferous blue-gray limestone containing many euomphalid gastropods, and an upper massive- to medium-bedded, very cherty (up to 30 percent), richly fossiliferous (crinoid columnals, gastropods, brachiopods, and corals) blue-gray limestone (Morris and Lovering, 1961, p. 90). This description applies, in general, to other interval B limestone sequences in northwestern Utah except for two subsurface Madison(?) sequences which contain appreciable amounts of terrigenous material and Osage microfaunas (Utah locs. 103, 104; H. J. Bissell, in Peace, 1956). The Gardison Limestone of the Gilson Mountains (Utah loc. 227) contain pockets and lenses as much as 10 feet thick of dolomite and chert pebble-to-cobble size conglomerate and breccia.

The Early Mississippian age of the Gardison Limestone and associated rocks is established on the basis of abundant megafaunas (Morris and Lovering, 1961). Calcareous foraminferal faunas of Osage age from Rock Canyon (Utah loc. 202) in the southern Wasatch Range were illustrated by Woodland (1958). Also, Clark (1954) illustrated foraminifers, interpreted by me to be of Osage age, from Rattlesnake Spur in the East Tintic Mountains, in the upper part of the *Gardner Formation* and overlying *Pine Canyon Limestone* beds here assigned to interval B. Davis (1956, p. 34, 38) described Osage coral faunas from all but the lowermost beds of the upper part of the *Gardner Formation* in central Utah which he stated contain Kinderhook faunas; all these beds are included here in interval B. Clark (1954) also described Osage coral faunas from several sections in central Utah and emphasized the presence of the gastropods *Euomphalus* and *Loxonema* in the upper part of the *Gardner Formation.*

Recently, it has been recognized that several rock units in central Utah assigned to interval C in this report constitute a starved basin facies of Osage to Early Meramec age (Sandberg and Gutschick, 1976; Poole and Sandberg, 1977). These units include the lower phosphatic unit of the Deseret Limestone, the lower part of the Woodman Formation, and the *Needle Siltstone Member* of the Chainman Formation and are discussed under interval C.

In the Confusion Range and vicinity and in the Silver Island Range (Utah loc. 113), the thin-bedded crinoidal locally sandy upper part of the Joana Limestone, where recognized, is assigned to interval B. The late Early Mississippian age of the upper part of the Joana in the Confusion Range was verified by Zeller (1957, p. 687). My recent work in the Conger Range (Utah loc. 252), southeast of Zeller's section, and at Skunk Springs (Utah loc. 253) supports his conclusions. Foraminiferal zone 7 (upper part) and foraminiferal zone 8(?) (lower Osage) are present in at least the upper 100 feet of these two sections.

South and east of the Needle Range (Utah loc. 258) in southwestern Utah, carbonate rocks assigned to interval B include parts of the Topache Limestone (Butler, 1913; Brill, 1963), Joana Limestone (Barosh, 1960), and Redwall Limestone (Stokes, 1963), or they have remained unnamed (Miller, 1966) as in the Wah Wah Mountains (Utah loc. 303). The sequences are lithologically and, locally, faunally similar to the Redwall of northern Arizona. Lithologic correlatives of the upper part of the Whitmore Wash Member and the Thunder Springs and Mooney Falls Members of the Redwall, where tentatively recognized, were assigned to interval B. The Whitmore Wash is generally fine grained, oolitic, and thick bedded; the Thunder Springs is very cherty and thin bedded; the Mooney Falls is largely a bioclastic crinoidal thick- to very thick bedded cliff-forming pure limestone (McKee and Gutschick, 1969).

THICKNESS TRENDS

An unknown thickness of volcanic rocks may comprise interval B in eastern Humboldt and Pershing Counties in northwestern Nevada, and near Taylorsville in northern California where a part of the metavolcanic and sedimentary sequences of the Peale Formation may belong to interval B. In both areas, however, interval B assignment is more tentative than are other interval assignments involving the same sequence of rocks.

Interval B rocks are absent, except as detritus, in younger conglomerates in the area of the Antler highland. They attain a maximum thickness of more than 1,575 feet in the Pinon Range along the axis of the Chainman–Diamond Peak trough, which is shown (pl. 4-A) to extend into the Eureka area along the eastern margin of the highland. If, as suggested by Nolan (1935) and Gordon (in Brew, 1971), the entire Chainman Shale in the Diamond Peak–Eureka area is of Late Mississippian age, then the area of thick interval B is more restricted as shown in figure 75, and the area of zero interval B is extended to include the Diamond Mountains as well as the southern Ruby Mountains and the Pancake Range.

Interval B is estimated to have a thickness of more than 2,000 feet in the southern Belted Range in the

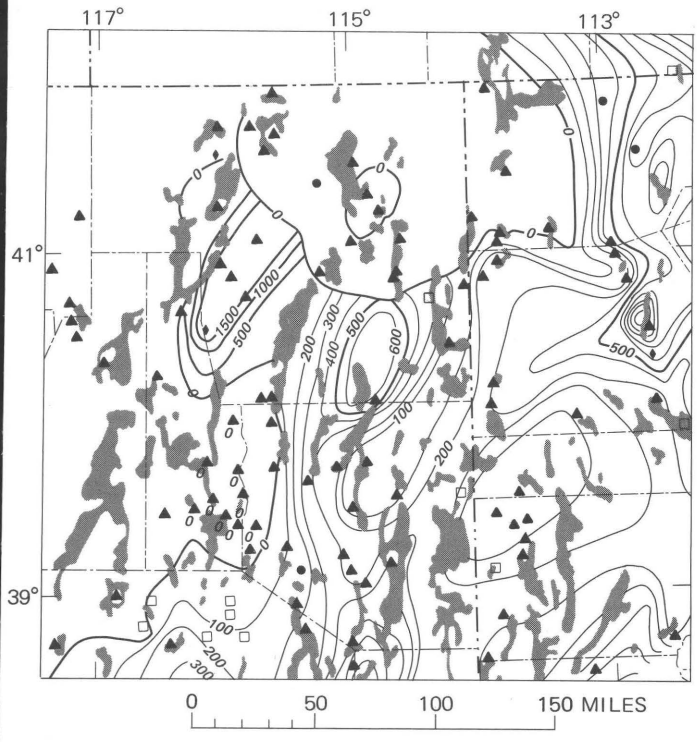


FIGURE 75. — Alternate isopach map of interval B showing thickness in feet in the Eureka–Diamond Peak area, Nevada, when the Chainman Shale is assigned to interval C rather than to intervals A and B. Triangles, dots, diamonds and squares are control points identified on plate 1. Shaded areas are outcrops of pre-Mississippian rocks.

area of the Nevada Test Site. Elsewhere in eastern Nevada and southern California interval B generally is less than 1,000 feet thick (pl. 9-G, sec. *U–U'*). Thicknesses near 1,000 feet are reported in southern Lincoln County and adjacent northern Clark County, Nev., in the northern Nopah Range (Calif. loc. 108), and in northern Inyo County (Calif. loc. 159). These are maximum thicknesses for interval B in these areas. In Lincoln County, some Kinderhook rocks may be included; in northern Clark County and in the Nopah Range less than 200 or 300 feet of Meramec rocks probably are included. Most interval B thicknesses in southeastern California and adjacent Nevada are also maximum thicknesses which may be reduced as more faunal evidence is obtained (see "Interval A" discussion.)

In western Utah interval B is absent in a large area of western Box Elder County in the vicinity of the Devonian(?) and Mississippian Wendover highland (pl. 9-G, sec. *R–R'*). The interval is also absent in places in the northern Confusion Range. Interval B strata are 1,100 feet thick in the northern Stansbury Mountains (Utah loc. 111) and approximately 1,000 feet thick in the Star Range (Utah loc. 301). Interval B rocks in northern Utah are thickest in the vicinity of the Oquirrh basin, but generally are less than 500 feet thick. The northeast-trending Ely positive area (Langenheim, 1956) or Ely arch (Langenheim, 1960, p. 73) is first apparent in the isopachs of interval B as a linear, relatively thin area extending southwestward from the position of the Wendover highland. The arch, however, is a postdepositional tectonic feature and interval B has been thinned by erosion.

Interval B thicknesses in southwestern Utah along the southern part of the Wasatch line are probably maximums (Utah locs. 301, 303) and may include as much as 300 feet of younger beds.

LITHOFACIES TRENDS

West of the Antler highland, calcareous mudstone interbedded with chert and slate and overlain by volcanic tuffs are tentatively assigned to interval B in the vicinity of Taylorsville in northern California (pl. 4-B). On and adjacent to the highland in eastern Humboldt County and northern Pershing County, Nev., volcanic rocks interbedded with limestone and minor shale and chert of the Goughs Canyon and Inskip Formations may belong in part to interval B.

Immediately adjacent to and east of the Antler highland in north-central Nevada, chert- and quartzite-pebble conglomerate, siltstone, and phyllite of the Grossman Formation tentatively are assigned to interval B. If this assignment is correct, the conglomerate of the Grossman, in addition to units C and F of the Eleana Formation in southern Nevada, represents the coarsest detritus recorded from Lower Mississippian rocks of the Great Basin. The Grossman and associated formations in northeastern Nevada may all be allochthonous (R. R. Coats, written commun., 1973).

In the Pinon Range, light- and medium-gray soft shale interbedded with fine- to coarse-grained sandstone and minor pebbly mudstone and ferruginous siltstone make up interval B. The sandstone is composed of angular to subrounded chert grains and minor rounded quartz grains. The pebbles and cobbles in the mudflow conglomerates are composed of chert, sandstone, and gray limestone probably derived from the Devonian Devils Gate Limestone (J. F. Smith, Jr., and K. B. Ketner, written commun., 1965). Locally, sandstone and conglomerate make up as much as 25 percent of the section.

Two facies in the upper part of the Chainman Shale in the Diamond Mountains area have been described (Brew, 1961b, 1964, 1971). The coarser of the two, the *Black Point facies*, consists of interbedded gray shale, silty claystone, and siltstone, and minor gray and brown sandstone and conglomerate (Nev. locs. 226, 227). Silicified chert and quartzite pebble-and-cobble conglomerate is present as lenses in the sandstone and

siltstone. Some of the conglomerates have siltstone matrices. Sandstone and conglomerate compose about 25 percent of the upper 2,000 feet of the *Black Point facies*. The coarser grained *Black Point facies* was thrust eastward over the finer grained *Water Canyon facies* (Nev. loc. 55) which consists generally of different proportions of the same lithologies as those just described. Sandstone and minor chert-pebble conglomerate form an estimated 10 percent of the *Water Canyon facies*, and siltstone and claystone the remainder (Brew, 1964, p. 55). The upper contacts of each of the two Chainman facies with the overlying Diamond Peak Formation are gradational; the possibility that these facies of the Chainman are Meramec in age has been discussed earlier.

Interval B is missing in the northern Pancake Range immediately south of the Diamond Peak area where Chainman Shale of interval C rests disconformably on Joana Limestone of interval A.

In the Windermere Hills (Nev. locs. 24, 25), interval B may be present in sandstone beds assigned to the Diamond Peak by Oversby (1969). Kinderhook conodont faunas are reported from limestone beds that seem gradational with the overlying sandstones. This occurrence extends the apron of interval B clastic sediments eastward into eastern Elko County.

Mudstone with minor sandstone and conglomerate has been assigned to interval B in southern Nye County, Nev.

Elsewhere in eastern and southern Nevada, interval B includes the slightly silty and sandy fossiliferous variably encrinitic calcarenitic limestone of the upper part of the Joana Limestone, the cherty and encrinitic lower part of the Monte Cristo Limestone (Dawn and Anchor Limestone Members), the micritic limestones of the upper part of the Crystal Pass Limestone (Carss and Carozzi, 1965), the cherty limestone of the Mercury Limestone, and the overlying very fine grained limestone of Timpi Canyon. Bedded and nodular chert makes up more than 10 percent of the interval in much of Clark County, Nev.

The cherty limestone facies persists south and west into southern California, where it is reported in the lower part of the Monte Cristo Limestone, upper part of the Tin Mountain Limestone, and lower part of the Perdido Formation. A nonpersistent facies in the lower part of the Perdido Formation in northern Inyo County (Calif. locs. 105, 154) locally includes conglomerate with subangular boulders of quartzite, limestone, and dolomite, as much as 42 feet in diameter (Ridley, 1970; Ross, 1966; Scott, 1960). Farther northwest in the northern Inyo Mountains (Calif. loc. 159) the Perdido that is assigned to interval B has a local basal limestone gravel conglomerate, but this thick section is mostly fine-grained dolomite, chert, and quartzite (Nelson, 1966).

Where recognized throughout western Utah (pl. 9-G, secs. $R-R'$, $S-S'$), interval B consists of limestone and minor argillaceous or sandy limestone. Chert constitutes 10 percent or more of interval B in part of southwesternmost Utah and in scattered localities in the vicinity of the Oquirrh basin.

Fine-grained sandstone with sandy and silty limestone from a well in central Box Elder County, Utah, contains Osage microfossils according to H. J. Bissell (in Peace, 1956). This well is the only known interval B locality in western Utah that contains significant amounts of clastic material.

SOURCES OF SEDIMENT, ENVIRONMENTS OF DEPOSITION, AND CLIMATE

Interval B detrital rocks in the Chainman–Diamond Peak trough are made up largely of clay and lesser amounts of chert- and quartzite-rich sandstone and conglomerate derived from a relatively low lying Antler highland to the west. Pebbly mudstones in the Chainman of the Pinon Range were probably emplaced catastrophically during earthquake activity. Available evidence suggests that more than 2,000 feet of fine detrital sediments accumulated locally in the axial portions of the major troughs adjacent to the highland on the east (pl. 9-G, sec. $R-R'$). Downwarping of the troughs seems to have been less pronounced in interval B than in succeeding Mississippian intervals, and sediment accumulation probably was slower. Brachiopods, corals, and mollusks from limestone lenses within the detrital rocks indicate that the limestones were deposited in warm marine waters of normal salinity and shallow depth (pl. 11, fig. 2). The area of zero interval B south of the Diamond Mountains is probably an erosional feature inasmuch as the Chainman here is unconformable on rocks of interval A. If the lower part of the Chainman of the Diamond Peak area is of Meramec age, a large eastward projection of the Antler highland (pl. 4-A; pl. 10, fig. 2) may have contributed detritus to the basin in the vicinity of the Pinon Range during interval B time.

A southern, but not necessarily connected, extension of the Antler highland presumably contributed detritus to the areas of the Eleana-*Meiklejohn* and lower part of the Perdido Formations in southern Nevada and southeastern California. Deposition within the trough apparently was slow and took place under basinal conditions (F. G. Poole, oral commun., 1972). Convoluted laminae, graded bedding, and small-scale crossbedding in the fine quartzites of unit E of the Eleana are features characteristic of turbidites (Poole, in Ekren and others, 1971, p. 22; Poole, 1973, 1974). The abundant chert in the limestones bordering the areas of shale and sand accumulation probably was introduced into the areas of carbonate deposition by upwelling sea

waters from the Eleana trough near the margins of the carbonate shelf. Local conglomerate, containing subangular boulders of quartzite, limestone, and dolomite as much as 42 feet in diameter, which are reported from beds assigned to interval B (Calif. locs. 105, 154) in northern Inyo County (Ridley, 1970; Ross, 1966; Scott, 1960), have been interpreted as Mississippian submarine channel rockslide deposits with a southeastern source (Ridley, 1970). The conglomerate beds are interbedded with turbidites and may have been derived from the western edge of the carbonate shelf and deposited along the eastern edge of the Eleana trough.

Elsewhere in southern California, southern Nevada, and southwestern Utah, chert is concentrated along the Wasatch line. In this belt the chert is coextensive with chert in the Redwall Limestone to the southeast. In the Arrow Canyon Range of southeastern Nevada the chert is younger than the main chert body in the Redwall. E. D. McKee (in McKee and Gutschick, 1969, chap. 11 and p. 560–561) has suggested that the chert in southeastern Nevada bordered a westward (or northwestward) regressing shoreline, which has been documented in northern Arizona, during interval B time. This concept, supported by foraminiferal determinations, is diagramed and extended westward into the Arrow Canyon Range of Nevada in figure 76.

The thin slightly silty and noncherty limestone that represents interval B across much of eastern Nevada and northwestern Utah was laid down in warm shallow marine waters as indicated by indigenous faunas and bioclastic textures. In addition, Ca/Mg ratios of Joana Limestone samples increase from the bottom to the top of the formation suggesting a possible gradual deepening of the Cordilleran seaway in this area during Early Mississippian time (Chilingar and Bissell, 1957, p. 2257). The increase in Ca/Mg ratios is accompanied by an increase in percentage of coarse-grained bioclastic limestone in the Joana Limestone of west-central Utah which suggests slightly deeper waters than those in which the micritic limestone of interval A accumulated.

A sedimentological and petrographic study of the upper part of the *Gardner Formation* of central Utah by Clark (1954, p. 36) shows that the overall marine environment was one of clear, temperate, and well-circulated shallow water with abundant bryozoans, corals, and crinoids. An increase in total percent of insoluble material compared with the lower part of the *Gardner*, along with more matrix and an increase in grain size in the upper part, suggests slightly more unstable shelf conditions than existed during interval A. Detrital quartz forms less than 1 percent of the total volume of the *Gardner* carbonate rocks and implies transport from a distant source area (Clark, 1954).

Significant amounts of terrigeneous material in subsurface strata reported from a well north of the Promontory Mountains are hypothesized to have been derived from the Wendover highland.

Volcanic activity west of the Antler highland in western Nevada and northern California is less well documented for interval B than for other Mississippian intervals. This apparent decrease in volcanism during interval B time is particularly interesting in view of the fact that east of the highland layered chert in the limestones occurs in greater volume than in any of the other intervals.

PALEOTECTONIC IMPLICATIONS

The generally thin limestone sequences of interval B, which cover the eastern part of the Great Basin (pl.9-G, sec. *U–U'*), were deposited under shelf conditions (Roberts and others, 1958; Schleh, 1968) slightly less stable than those of interval A (Chilingar and Bissell, 1957). The carbonates extend far to the east beyond the Wasatch line and as far west as north-central Nye County, Nev. (pl. 10, fig. 2). Fossiliferous gravels of interval B (Osage) and interval C (Meramec) limestone in a Permian conglomerate in the southern Toiyabe Range (near Nev. loc. 155; N. J. Silberling and Helen Duncan, written commun., 1959; F. G. Poole, oral commun., 1968) suggest that in Osage and Meramec time, intervals B and C limestones may have extended farther west and shallow marine waters may have covered part of the slowly rising, locally inundated, Antler highland. Because the Permian conglomerate also contains greenstone gravels that represent submarine mafic lavas, the limestones may have been deposited west of the highland and subsequently thrust eastward and eroded in late Paleozoic time (F. G. Poole, written commun., 1973). Faulted sequences of Mississippian limestone of intervals B and C have been identified in west-central Nye County (Nev. locs. 157, 158). In this report, an embayment on the eastern margin of the highland is hypothesized to have been the site of deposition of this limestone.

Gray shale and lesser sandstone and pebbly mudstone, whose aggregate thickness is about 1,600 feet in the center of the northern part of the Chainman–Diamond Peak trough, are the only well-dated interval B detrital rocks in the Great Basin. They were deposited in a slowly subsiding trough on the eastern edge of the Antler highland. The trough probably underwent continuous slow subsidence without major interruption into interval C time. There is no evidence for local post-interval B uplift and erosion in the Carlin – Pinon Range–Windermere Hills area as there is in almost all other Great Basin sections to the east and southeast.

The area shown as a southern lobe of the Chainman–Diamond Peak trough on plates 4-B and 4-C may have been a neutral or low positive area during all or part of interval B time (fig. 75). If it was a low positive

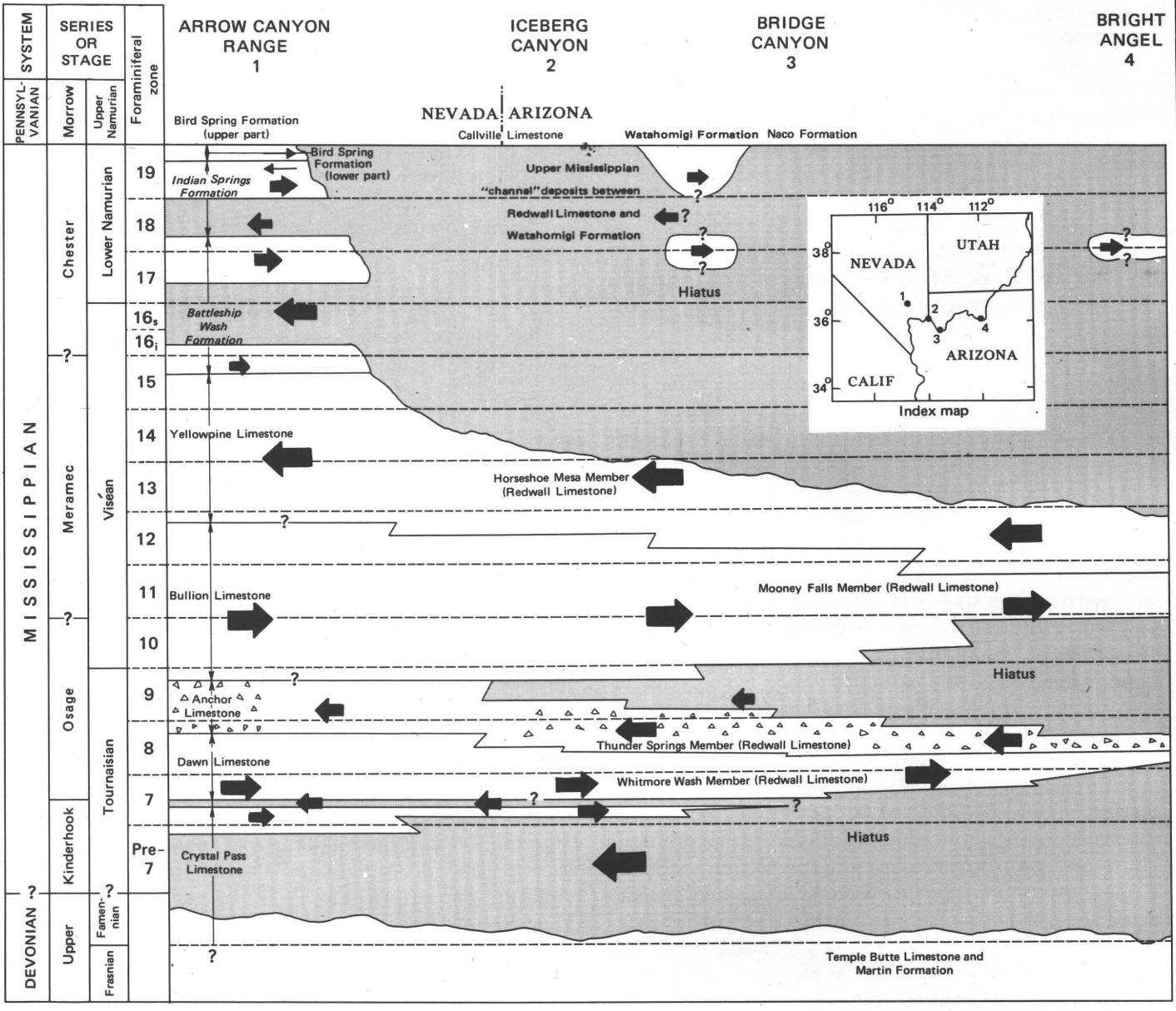


FIGURE 76. — Diagram showing the history of transgression and regression and changes in depositional environment of the Mississippian seas across southern Nevada and northern Arizona based on time lines represented by the international foraminiferal zones of Mamet (in Mamet and Skipp, 1970b) modified as suggested in Brenckle, Lane, and Collinson (1974). Microfaunal studies of Brenckle (1973) in the Arrow Canyon Range, Nevada, and Skipp (in McKee and Gutschick, 1969) in northern Arizona and unpublished data by Skipp, form the basis for zonal ranges of the formations. The pattern of transgressions and regressions of the Redwall Sea in Arizona is from McKee and Gutschick (1969, fig. 26). Large arrows indicate major transgressions and regressions. Triangles indicate chert. Unnamed Upper Mississippian channel deposits recently were discovered in the western Grand Canyon area by G. H. Billingsley (written commun., 1977).

area, it did not contribute much detritus to surrounding areas where carbonate deposition was dominant.

Evidence for a southern segment of the Antler highland shown on plate 10, figure 2 is indirect. None of the detrital rocks shown in southern Nye County and southeastern California are well dated. The shale, sandstone, and conglomerate of the Eleana Formation assigned to interval B were deposited in the subsiding Eleana trough next to a low or moderately positive area. The cherty carbonate rocks of the Tin Mountain Limestone and the Monte Cristo Limestone are well dated, but the sporadic basal conglomerate, shale, siltstone, and sandy carbonate of the lower part of the Perdido are assigned arbitrarily to interval B although the lower part of the Perdido may be of Late Mississippian age. Silica for the chert-rich shelf carbonate

deposits may have been transported eastward by upwelling currents from the deeper parts of the Eleana trough.

The Wendover highland in northwestern Utah may have been emergent during part of interval B time. Detrital sedimentary rocks are present both east and west of the proposed highland (pl. 4-B); possibly, however, all the detritus was derived from the Antler highland and was removed from the Wendover area by Pennsylvanian and Permian erosion. Limestones of the Confusion Range and vicinity in west-central Utah contain virtually no terrigenous material, suggesting that the thin or missing interval B limestones in that area are probably the result of early Meramec (interval C) epeiric uplift and erosion on the Ely arch and are not related to an interval B positive area.

PALEOGEOGRAPHIC INTERPRETATIONS

During interval B time, warm, shallow, and well-circulated marine waters covered most of the Great Basin, where thin sequences of relatively pure bioclastic limestone accumulated (pl. 12, fig. 2). The emergent Antler highland of generally low relief probably stood largely as low hills, with local areas of high (more than 1,000 feet) relief which supplied minor conglomeratic detritus to small areas in north-central Nevada. Conglomeratic mudstones of the Chainman Shale in the Pinon Range were emplaced, probably catastrophically, during earthquakes which accompanied the rise of the Antler highland.

Abundant chert in the limestones of southeastern California and southern Nevada may provide indirect evidence for a moderately deep Eleana trough to the west from which silica-rich upwelling marine currents moved eastward onto the shelf. The chert-rich limestones of southwestern Utah, however, are interpreted as being formed in a nearshore marine environment into which silica-rich fresh waters were introduced by northwest-flowing rivers that originated on the low-relief craton to the east. It is inferred that chert formed as silica selectively replaced carbonate layers shortly after deposition. The shoreline must have fluctuated repeatedly, but the overall sum movement was a withdrawal of the Redwall–Monte Cristo Sea toward the northwest from a broad emerging craton to the southeast at the end of interval B time (fig. 76). This concept is one of several proposed by E. D. McKee to explain chert distribution in the Redwall Limestone (McKee and Gutschick, 1969, p. 560–561). There is some evidence in southern Nevada (Spring Mountains and Arrow Canyon Range) and southwestern Utah (Wah Wah Mountains) for significant chert formation in early Meramec time, which is slightly younger than chert formation in the Redwall of northern Arizona. This regional trend supports McKee's concept of a general southeast-to-northwest regression.

Although volcanic islands are shown in Humboldt and Pershing Counties, Nev., and in the Taylorsville area of northern California, faunal evidence for volcanism in these areas in interval B time is less compelling than for any of the other Mississippian intervals.

INTERVAL C

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

SOUTHERN CALIFORNIA

In southeastern California the upper part of the Perdido Formation and the Bullion, Arrowhead, and Yellowpine Members of the Monte Cristo Limestone are included in interval C (pl. 15). The upper part of the Perdido Formation in its type section consists of calcareous siltstone, sandstone, limestone boulder conglomerate, chert, locally silty limestone, and some shale (McAllister, 1952, p. 22–25). It is reported to contain Upper Chester cephalopods in its uppermost bed and an Early Mississippian coral fauna in its lower part (Gordon, 1964, p. A2). All the upper part of the Perdido is here included in C for convenience. Conodont and foraminiferal faunas of Meramec age have been recovered from 205 feet above the base of the Perdido Formation in the southern Funeral Mountains. The faunas include the conodont *Taphrognathus varians* (J. W. Huddle, written commun., 1965), and the Foraminifera *Endothyra bowmani* and *Globoendothyra tomiliensis*, all indicative of a Meramec age (McAllister, 1974).

All but the basal 40 feet of a "limestone, younger than the Tin Mountain, in part of Mississippian age" (Hunt and Mabey, 1966, p. A44) in the northern Panamint Range (Calif. loc. 158), is included in interval C. The late Meramec index coral *Faberophyllum*, is present in the upper beds of this limestone unit.

The lower part of the Lee Flat Limestone 10 miles north of Darwin is thought to be a "time-stratigraphic equivalent of the upper clastic part of the Perdido Formation * * * of the Ubehebe Peak quadrangle" (Hall and MacKevett, 1962, p. 19). Although this relationship is shown on the correlation chart (pl. 15) and has been corroborated by Pelton (1966, pl. 1), the entire Lee Flat Limestone is included in interval D, and the Perdido in interval C in the present report.

The Bullion Limestone Member of the Monte Cristo Limestone is included in interval C because questionable Meramec calcareous foraminiferal faunas were recovered from that member in the Arrow Canyon Range, Nev. (Brenckle, 1973). Conodant faunas reported by Pierce and Langenheim (1974) suggest an

Osage age for the Bullion Limestone Member. (See discusion on p. 292.) The Meramec age of the Arrowhead and Yellowpine Limestone Members of the Monte Cristo was largely determined from megafaunas in the Goodsprings area in southern Clark County, Nev., by Girty (in Hewett, 1931, p. 18–19) and has been supported by foraminiferal faunal evidence (Brenckle, 1970, 1973) and, in part, by conodont determinations (Pierce and Langenheim, 1974).

A part of the Furnance Limestone of southwestern San Bernardino County is included in interval C as in all other intervals, owing to Mississippian fossils found by J. F. Richmond (1960, p. 15–18) in the lower part of the formation in the San Bernardino Mountains. Other metasedimentary rocks that may contain interval C equivalents are the Oro Grande and Bean Canyon Formations and related rocks in southern California (Bowen, 1954; Dibblee, 1960, 1967).

A part of the upper part of the Peale Formation at Taylorsville in northern California is included with question in interval C. No Late Mississippian fossils have been reported from the sequence, but tuff, siliceous mudstone, and bedded chert units similar to those found in the Peale are also found in the lower part of the Baird Formation of Late Mississippian age in the Klamath Mountains (Yates, chap. Q) of northern California.

NEVADA

A part of the Goughs Canyon and Inskip Formations in northwestern Nevada are included in interval C (pl. 15). Coral and brachiopod faunas identified by Gordon and Duncan (in Hotz and Willden, 1964, p. 26–28; Willden, 1964, p. 19–20) from these formations suggest both Osage and Meramec age (Tournaisian and Visean). A calcareous foraminiferal fauna from the Goughs Canyon contains *Endothyranopsis* cf. *E. compressa* which indicates a middle and early late Visean (Meramec) age. The Goughs Canyon and Inskip Formations, considered as possible time correlatives (Duncan, in Roberts and others, 1958, p. 2846–2847; Hotz and Willden, 1964, p. 28), consist largely of altered andesitic and basaltic volcanic rocks, breccias, pillow lavas, and interbedded impure limestone and minor shale and chert (Willden, 1964, p. 19). In the Mountain City–Owyhee area (Nev. loc. 12), similar interbedded limestones and volcanic rocks occur at the base of the Nelson Formation which is gradational with impure limestone of the underlying Banner Formation (Coats, 1969). Meramec faunas have been reported from the Banner Formation (Duncan, in Coats, 1969, p. A25), and therefore both the Banner and Nelson Formations are included in interval C. The volcanic rocks of the Nelson Formation (Nev. loc. 13) have been metamorphosed to amphibolite and greenschist (Coats, 1969, p. A25–A26).

In the Carlin area (Nev. loc. 1) more than 3,000 feet of interbedded sandstone, siltstone, and conglomerate of the undivided Chainman Shale and Diamond Peak Formations is assigned to interval C on the basis of megafaunas identified by Gordon and Duncan (in Brew, 1971). A similar thickness has been estimated for interval C strata in the Adobe Range (Nev. loc. 22) to the north (Ketner, 1970a). Both the interval C and D sequences in the Carlin area contain 40–60 percent conglomerate (Smith and Ketner, 1968, p. 113). The percentage of conglomerate generally decreases southward through the Pinon Range area. In the Diamond Mountains only members A and B of the type Diamond Peak (Nev. loc. 226) are assigned here to interval C. These members contain less than 7 percent conglomerate, one-third of which consists of limestone pebbles and cobbles in a siltstone matrix (Brew, 1964, 1971). If as, suggested by Gordon (in Brew, 1971), the entire several thousand feet of underlying Chainman Shale in the Diamond Peak area is also part of interval C, the percentage of conglomerate is reduced to much less than 1 percent. (See discussion under "Thickness Trends" for interval C.)

Less than 1 percent conglomerate occurs in the 2,300 feet of siltstone and sandstone that comprises the lower part of the Chainman Shale and Diamond Peak Formation undivided in the northern Pancake Range (Nev. loc. 52; Stewart, 1962). These undivided rocks are assigned to interval C on the basis of Meramec faunas from near the base of the sequence reported by Gordon (in Brew, 1971), who also reported Meramec ammonoids from the lower part of the Chainman Shale in the White Pine Range to the east.

Farther south in north-central Nye County, one-half of the estimated thickness of the incomplete Eleana Formation or undivided Chainman Shale–Diamond Peak Formation is arbitrarily assigned to interval C. Meramec faunas from impure limestones on the east side of the northern Hot Creek Range (near Nev. loc. 166) have been reported by Mackenzie Gordon, Jr., and Helen Duncan (written commun., 1967). And, farther west in the southern Toiyabe Range (near Nev. loc. 155), cobbles of limestone containing Meramec conodonts were found in a Permian conglomerate by F. G. Poole (oral commun., 1968), and the conodonts were identified by J. W. Huddle (written commun., 1968). In the northern part of the Nevada Test Site, 1,700 feet of strata comprising units G and lower H of the Eleana Formation is assigned to interval C (Poole and others, 1961; Orkild, 1963; Poole and others, 1965; F. G. Poole, oral commun., 1972). Interbedded conglomerate and quartzite and minor argillite are the dominant lithologies composing interval C in the test site. Although reworked Meramec and older megafossils occur in Unit I of the Eleana, early Chester ammonoids are present in the upper part of Unit H and are the

basis for assigning lower H and G to interval C (F. G. Poole, oral commun., 1972). An unspecified thickness of silty claystone of the *Meiklejohn Formation* in southwestern Nye County is assigned provisionally to interval C on the basis of stratigraphic position below Chester microfaunas and lithologic similarity to the Eleana and Chainman Shale–Diamond Peak Formation (Cornwall and Kleinhampl, 1964).

In the northern Pequop Mountains, Sadlick (1965, p. 142–145) measured approximately 2,700 feet of siltstone, sandstone, and claystone belonging to his *Needle Siltstone Member* and lower part of the overlying *Camp Canyon Member* of the Chainman Shale, which are included here in interval C (Nev. loc. 9). Early Late Mississippian faunas have been recovered from several stratigraphic levels in this section (Sadlick, 1965, p. 63; Thorman, 1962).

Elsewhere in eastern Nevada interval C is thin or absent. In the Wendover area, Utah, Sadlick reported 81 feet of *Needle Siltstone Member* of the Chainman Shale that he assigned to the late Meramec. In the northern Egan and Schell Creek Ranges, in the Ely area, and at Dutch John Mountain, interval D rests unconformably on the Joana Limestone of interval B (Drewes, 1967; Gordon, in Brew, 1971; Langenheim and Peck, 1957). The area wherein interval C is missing extends southward (pl. 5-A) into northwestern Clark County and adjacent Nye County in the southern Spotted Range (Nev. loc. 137), where rocks of interval D unconformably overlie the limestone of Timpi Canyon of interval B (F. G. Poole, written commun., 1965). The absence of interval C is due to an erosional hiatus between Lower and Upper Mississippian strata first described by Nolan (1943, p. 154; Nolan and others, 1956, p. 55) in the eastern and central Great Basin.

A thin siltstone sequence at the base of the Chainman Shale, locally called Peers Spring Formation by R. L. Langenheim, Jr., (written commun., 1967) is included in interval D in the southern Egan and Schell Creek Ranges, Pahranagat Range, and associated areas in southern Nevada even though no faunal control is available for these rocks at present, and they possibly represent a thin blanket of interval C rocks.

In Clark County, the basal part of the Battleship Wash Formation, where present, and the Yellowpine, Arrowhead, and Bullion Limestone Members of the Monte Cristo Limestone constitute interval C. A calcareous foraminiferal fauna in the middle part of the Bullion Member is considered to be Meramec in age (Brenckle, 1970, 1973; Mamet and Skipp, 1970b).

WESTERN UTAH

Interval C in most of north-central Utah is made up of the Deseret Limestone and the overlying Humbug Formation (pl. 15). These formations are recognized as far north as the Promontory Mountains (Utah loc. 102) and as far south as the Gilson Mountains (Utah loc. 227) southwest of the East Tintic Mountains. In this area, the overlying Great Blue Limestone reportedly contains Meramec beds (Weller and others, 1948), but these are not included in interval C because of insufficient paleontologic and lithologic information on which to make a regional subdivision. Morris and Lovering (1961, p. 112) noted that Crittenden and others (1952, p. 10) had included 200–275 feet of blue limestone, which may be part of the Great Blue, at the top of the redefined Humbug Formation in the Cottonwood Canyon area near Salt Lake City (Utah loc. 127). These beds are included in interval C.

The Deseret Limestone was named and defined by Gilluly (1932, p. 25) in the southern Oquirrh Mountains (Utah loc. 117) where he described 650 feet of thin- to massive-bedded blue limestone, some of it sandy and cherty with 9 feet of "black, red-weathering shale with a phosphatic oolite bed at the top at the base of the formation." No fossils were collected from this basal shale member, but it was tentatively assigned to the Upper Mississippian. Morris (1957) and Morris and Lovering (1961) divided the Deseret Limestone of the East Tintic Mountains (Utah loc. 201) into three members: (1) the phosphatic shale member at the base, 91 feet thick; (2) the Tetro Member, cherty limestone, 475 feet thick; and (3) the Uncle Joe Member at the top, massive bioclastic limestone, 544 feet thick. Faunas near the top of the phosphatic shale member northeast of the Tintic mining district contain keeled crinoid columnals and the brachiopods *Leiorhynchus carboniferum polypleurum* Girty and "*Productus*" *hirsutiformis* Walcott (Gordon, in Morris and Lovering, 1961, p. 97). Recently, Petersen (1969, p. 63) described the ammonoids *Dahaprakoceras* and *Beyrichoceras* from the basal phosphatic shale unit of the Deseret in the northern Stansbury Mountains (Utah loc. 111) and considered them indicative of a Meramec age. *Beyrichoceras* is common in the middle Visean of the British Isles, though Gordon (1964, p. 183) stated that: "All the American occurrences are probably in beds of Osage age." *Quadratia hirsutiformis* and *Leiorhynchus carboniferum polypleurum* were considered indicative of the middle Meramec (megafaunal Zone pre-E) in southeastern Idaho and northeastern Utah by Sando, Mamet, and Dutro (1969, p. E10), but a paper by Sando, Dutro, Sandberg, and Mamet (1976) revises the age of these beds, and an Osage age for the lower part of the section is indicated. I have included all of the Deseret in interval C (pl. 15, col. 110), though it is now recognized that a few hundred feet of limestone and the basal phosphatic shale member of the formation are of Osage age.

The lower beds of the Deseret as used herein have been called the upper member of the Pine Canyon Formation by Bissell (1959, p. 47–49) in the southern

Oquirrh Mountains. In many localities the basal phosphatic shale member of the Deseret is either absent or concealed, as in the Promontory Mountains (Utah loc. 102), northern Lakeside Mountains (Utah loc. 110), northern Sheeprock Mountains (Utah loc. 121), southern Wasatch Range (Utah loc. 202), Lake Mountains (Utah loc. 205), and Gilson Mountains (Utah loc. 227).

The Humbug Formation gradationally overlies the Deseret and, in places, is mapped with it. The Humbug consists of interbedded lenticular crossbedded sandstone and limestone and sandy limestone with minor dolomite and shale. The upper contact with the overlying Great Blue is also gradational and is placed at the top of the uppermost prominent quartzite or sandstone bed (Morris and Lovering, 1961, p. 104). Microfaunas in the Humbug (Woodland, 1958) and its position between strata of established Meramec age assures assignment to interval C.

The Woodman Formation makes up interval C in the Deep Creek Mountains (Nolan, 1930, 1935) and Dugway Range (Staatz and Carr, 1964), where it consists of silty limestone and calcareous siltstone that is correlated with the Deseret-Humbug sequence in the East Tintic Mountains. The lower part of the Woodman Formation is now known to be of Osage age (Sandberg and Gutschick, 1976).

In northwestern Utah interval C is absent in the Newfoundland Mountains and northern Silver Island Range, where Pennsylvanian and Permian conglomerate has been reported to rest with angular unconformity on Devonian limestone and shale (Paddock, 1956; Anderson, 1957, 1960; Schaeffer and Anderson, 1959). Roberts and Tooker (1969), however, suggested that these same conglomerates may be a northwestward extension of the Stansbury Formation (Devonian). The conglomerates in the Newfoundland Mountains and Silver Island Range are overlain by dated Permian carbonate rocks. These conglomerates are considered to be Pennsylvanian and Permian in age in this report as suggested by Paddock and Anderson. The area of zero interval C is extended northward to include the Muddy Range, Grouse Creek Mountains (Utah loc. 105) and Vipont Mountains (informal name; Utah loc. 109) and westward to include the northern and southern Pilot Range (Blue, 1960, p. 35; O'Neill, 1968).

Interval C probably is absent or may be very thin in the northern Confusion Range at Granite Mountain (Utah loc. 226) where the Chainman Formation rests on siltstone of the Pilot Shale (R. K. Hose, written commun., 1965) or is separated from the Pilot Shale by a thin wedge of Joana Limestone (Mackenzie Gordon, Jr., written commun., 1976). Foraminiferal faunas representative of foraminiferal zone 16i containing *Neoarchaediscus* of Early Chester age occur 150 feet above the base of the Chainman Shale at this locality. A few hundred feet of uppermost Meramec rocks in the same locality were reported by Gordon (1971) who studied the goniatite fauna. Farther south, in the Needle Range (Utah loc. 258), 550 feet of calcareous siltstone and sandstone, which makes up the type section of the *Needle Siltstone Member* of the Chainman Formation and contains a Meramec brachiopod and goniatite fauna, unconformably overlies the Joana Limestone (Sadlick, 1965, p. 40–45). The *Needle Siltstone Member* is overlain by the upper part of the *Camp Canyon Member* of interval D. The *Needle Siltstone Member* of the Chainman Formation recently has been correlated with the lower (Osage) parts of the Deseret and Woodman Formations (Sandberg and Gutschick, 1976).

In southwestern Utah a sequence of interbedded limestone and sandstone which has been called Redwall, Joana, or Topache Limestone (Hintze, 1963; Barosh, 1960; Earll, 1957) is assigned to intervals B and C in this report. This sequence is lithologically and, in places, faunally similar to the well-studied Redwall Limestone in adjacent northern Arizona (McKee and Gutschick, 1969).

THICKNESS TRENDS

In northern California an unknown thickness of interval C volcanic rocks is present in the Klamath Mountains (Yates, chap. Q) and also may be present in the Taylorsville area. On the western edge of the Antler highland in Humboldt and Pershing Counties, Nev., 5,000 feet or more of allochthonous interbedded volcanic rock and limestone of the Goughs Canyon and Inskip Formations may belong to interval C (pl. 5-A). These rocks, with thickness indicated by "plus marks" on plate 5-A, were deposited in the eugeosynclinal area west of the highland and were moved into the present position by later thrusting.

More than 4,000 feet of interval C volcanic rocks of the Nelson Formation, the basal part of which is interbedded with the upper part of the Banner Formation, may have been erupted on or near the eastern edge of the Antler highland in northern Elko County, Nev. Possibly, however, the entire sequence is allochthonous (R. R. Coats, written commun., 1973) and its site of deposition unknown.

The Antler highland probably received no sediments, but the Chainman–Diamond Peak and Eleana troughs along the eastern edge of the highland received about 1,500 to more than 3,000 feet of detrital sediments during interval C time (pl. 5-A). Thicknesses in the Diamond Mountains are based on the assumption that intervals A and B are represented by the fine-grained clastic rocks of the Chainman Shale as was suggested

by Brew (1964, 1968). A recent synthesis of faunal data by Gordon (in Brew, 1971) has suggested, however, that all the Chainman Shale in the Diamond Peak area may be Meramec in age and has agreed with the Late Mississippian age assigned to the Chainman of the Eureka district by Nolan and others (1956, p. 60). At this writing, the faunal evidence seems inconclusive, but an alternate isopach map for the area is presented in which all the Chainman Shale in the Diamond Mountains area is assigned to interval C (fig. 77). The position of the Chainman–Diamond Peak trough is not changed in this construction but interval C thicknesses exceed 5,600 feet.

In the area of the northern part of the Diamond Peak–Chainman trough in northeastern Nevada, post-Ely Limestone erosion locally has reduced the thickness of strata to a zero edge, though complete Mississippian sections may be preserved in the unpenetrated autochthon in this part of Nevada. (See discussion of post-Mississippian structures, p. 318.)

The Wendover highland, an area of zero Mississippian in northwestern Utah, is located east of the northern part of the Diamond Peak–Chainman trough. South and east of the Wendover highland, interval C is more than 1,500 feet thick in two depocenters which occupy the general position of the Oquirrh basin.

Over much of eastern Nevada and westernmost Utah, rocks of interval C are thin, being less than 500 feet thick, or they are absent (pl. 9-G, secs. *S*–*S'*, *U*–*U'*). This is the area occupied by the Ely arch which extends from the northern Egan Range, Nev., east to include the House, Confusion, and Needle Ranges of Utah and south to include most of Lincoln County and part of Clark County in southeastern Nevada.

East of the Ely arch is another elongate northeast-trending trough containing more than 1,000 feet of interval C strata. From this trough eastward, interval C thins to the south and east along the west edge of the craton, and an area of zero interval C is thought to be present in the Mineral Mountains (Utah loc. 302) and along the southwest flank of the Pavant Range (Utah loc. 251).

Separation of the Chainman–Diamond Peak and Eleana troughs is not as pronounced in this interval as in intervals A and B. Thicknesses between the troughs are about 1,500 feet. More than 5,000 feet of strata may be present in the Chainman–Diamond Peak trough, but only about 1,700 feet is assigned to interval C in the Eleana trough. The thickest interval C detrital sequence in southeastern California is 1,000 feet in the upper part of the Perdido Formation on the east side of Owens Valley (Calif. loc. 159). This thick section may be in an offset segment of the Eleana trough.

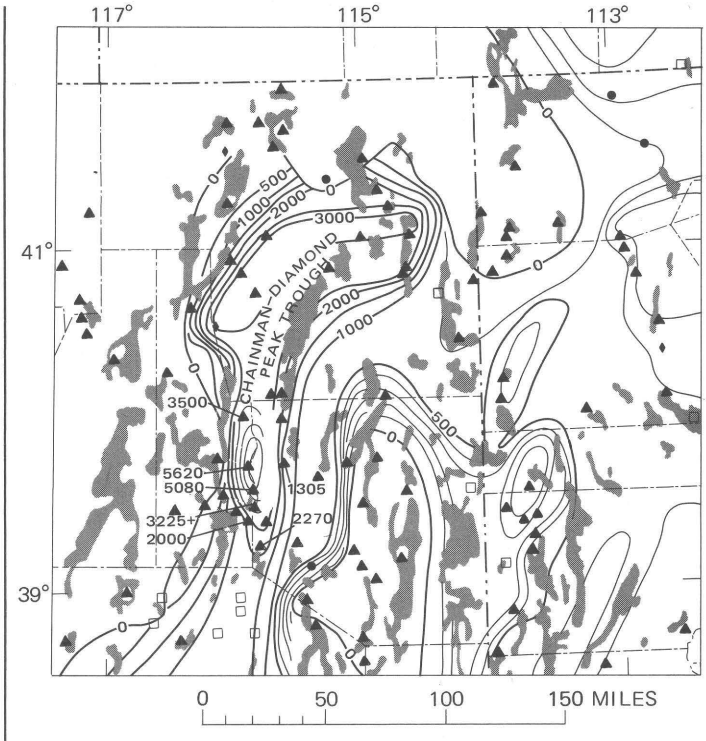


FIGURE 77. — Alternate isopach map of interval C showing thickness in feet in the Eureka–Diamond Peak area, Nevada, when the Chainman Shale is assigned to interval C rather than to intervals A and B. Triangles, dots, diamonds, and squares are control points identified on plate 1. Shaded areas are outcrops of pre-Mississippian rocks.

LITHOFACIES TRENDS

Evidence for volcanism during interval C at Taylorsville west of the Antler highland is inconclusive, but well-dated Upper Mississippian volcanic rocks are present in the Klamath Mountains to the north (Yates, chap. Q).

The Goughs Canyon Formation, located on the west edge of the Antler highland in north-central Nevada in the Hot Springs Range and Osgood Mountains, consists of approximately 70 percent altered andesitic and basaltic volcanic rocks, 25 percent limestone, and 5 percent or less calcareous sandstone, calcareous shale, siliceous shale, and chert (Willden, 1964, p. 19). The Goughs Canyon is well dated by megafaunas and microfaunas, and remnants of the formation preserved in thrust slices probably represent but a small part of the original thickness and extent of the volcanic material assigned to interval C. The possibly related Inskip Formation contains less limestone and more altered mudstone, sandstone, and minor conglomerate.

In northeastern Nevada, coarse to fine detritus shed from the Antler highland fills the Chainman–Diamond Peak trough to the east. Conglomerate and coarse sandstone composed of quartzite and chert detritus are

restricted to a narrow belt extending southward from the Adobe Range (Nev. loc. 22) through the Pinon Range (Nev. loc. 1) and into the Diamond Mountains (Nev. loc. 226). If all the Chainman Shale of the Diamond Mountains is assigned to interval C (fig. 77), it would be dominated by fine detritus rather than medium- to coarse-grained material shown on plate 5-B. (See discussion of "Interval C, Thickness Trends".)

The Banner Formation (Nev. loc. 12), which consists of calcareous sandstone and sandy limestone, is an anomalous lithologic assemblage on the northwest flank of the Chainman–Diamond Peak trough in northern Elko County. The Banner grades vertically and, perhaps, laterally, into the andesitic and basaltic volcanic rocks of the overlying Nelson Formation (Coats, 1969, p. A24–A25), the basal part of which is assigned to interval C.

To the south, in the northern part of the Nevada Test Site (Nev. loc. 151) in the Eleana trough, units G and lower H of the Eleana Formation, which are assigned to interval C, are composed mainly of conglomerate and quartzite and minor argillite (Poole and others, 1961). These beds probably correlate with part of the *Meiklejohn Formation* (of local usage) to the southwest (Nev. loc. 154). The *Meiklejohn*, in general, however, is slightly finer grained than the Eleana. According to Cornwall and Kleinhampl (1961), it consists of "siltstone, claystone and shale with subordinate but locally dominant intercalated beds of dark-gray chert, brown sandstone and conglomerate, and long thin lenses of medium-gray limestone. Minor lithologies are commonly cyclical, with graded bedding in clastic rocks."

Farther south and west in southeastern California, interval C is represented by the upper part of the Perdido Formation (pl. 9-G, sec. T–T') which, on the east side of Owens Valley in northern Inyo County (Calif. loc. 159), consists of thin-bedded gray shale with interbeds of medium-grained brown quartzite and chert pebble-and-granule conglomerate (Nelson, 1966). In the vicinity of the type section in the Quartz Spring area (Calif. loc. 101) in the northern Panamint Range, the Perdido consists of interbedded cherty limestone, siltstone, bedded chert, and minor limestone-boulder conglomerate (McAllister, 1952). Farther south and east of the type section, the cliff-forming limestones of the upper part of the Perdido are relatively pure, but they are still interbedded with as much as 30 percent chert in the southern Panamint (Calif. locs. 109, 158) and Argus (Calif. loc. 118) Ranges (Hall and Stephens, 1962, 1963; Hunt and Mabey, 1966; Pelton, 1966) and in the southern Funeral Mountains (Calif. loc. 152; Denny and Drewes, 1965). Abundant chert in limestones of interval C is common in southern Nevada east of the belt of major clastic deposition in the Last Chance Range (Nev. loc. 156) and near Indian Springs in northwestern Clark County (Nev. loc. 138; pl. 5-B; F. G. Poole, written commun., 1965).

Except for the limestones of the *Battleship Wash Formation*, and the Yellowpine, Arrowhead, and Bullion Limestone Members of the Monte Cristo Limestone which represent interval C in Clark and southern Lincoln Counties of southeasternmost Nevada, siltstone, shale, and mudstone, locally calcareous, of the lower part of the Chainman Shale are assigned to interval C along most of the eastern edge of Nevada and in the Confusion Range and Needle Range areas of west-central Utah (pl. 9-G, sec. S–S'). North of the Confusion Range, the Chainman siltstone and shale grade into sandy limestone and calcareous sandstone of the Deseret Limestone and Humbug and Woodman Formations of northern Utah. Both bedded and nodular chert, which locally comprises 30 percent of the Deseret Limestone, makes up more than 10 percent of the total thickness of interval C only in the northern Stansbury Mountains (Utah loc. 111; Rigby, 1958). Sandstone and siltstone make up more than 50 percent of interval C in the northern Lakeside Mountains (Utah loc. 110; Doelling, 1964) and Promontory Mountains (Utah loc. 102; Olson, 1960). The prominent basal phosphatic shale member of the Deseret Limestone, where recognized, makes up less than 5 percent of the total volume of interval C. A unit of sandy limestone, sandstone, and cherty dolomite is assigned to interval C in the Wah Wah Mountains (Utah loc. 303). To the south in southwestern Utah and southern Nevada, interval C is characterized by pure limestone of the Redwall and Monte Cristo Limestones.

SOURCES OF SEDIMENT, ENVIRONMENTS OF DEPOSITION, AND CLIMATE

Widespread volcanism in the northwestern Great Basin region occurred during interval C time. From the Klamath Mountains of northern California (Yates, chap. Q) eastward to eastern Humboldt County in north-central Nevada and northern Elko County in northeastern Nevada, the lower Upper Mississippian rocks record intermittent volcanic activity that locally added large volumes of andesite and basalt to the geologic record. The allochthonous Goughs Canyon and Inskip Formations in Pershing and Humboldt Counties on and near the western edge of the Antler highland contain altered pillow lavas which indicate emplacement on an ocean floor. Limestones associated with the volcanics are contaminated with volcanic debris and are fossiliferous. The corals, brachiopods, algae, and rare Foraminifera which make up the biota indicate a warm shallow subtidal marine environment (pl. 11, fig. 3) analogous to modern shallow marine environments surrounding volcanic islands in the South Pacific.

Detrital material forms less than 5 percent of the Goughs Canyon Formation. The Inskip Formation contains less limestone and more altered shale, sandstone, and minor conglomerate than does the Goughs Canyon. This detritus may have been derived from the west side of the Antler highland. Present evidence suggests that both formations were deposited west of the highland and were thrust to their locations in post-Mississippian time.

Interval C volcanism was not restricted to the area west of the Antler highland. The Banner and Nelson Formations in north-central Elko County (Nev. loc. 12) locally rest depositionally on allochthonous Ordovician eugeosynclinal rocks of the Valmy Formation and presumably were deposited near the eastern edge of the highland. The sandy limestones of the Banner Formation, which contain detritus very like local quartzites of the Valmy, are gradational with submarine peperites at the base of the overlying Nelson Formation (Coats, 1969, p. A25).

The Antler highland controlled sediment distribution east of the uplift during interval C time. Abundant chert derived from lower Paleozoic eugeosynclinal rocks exposed on the highland have been recognized in conglomerates of interval C along the entire length of the eastern edge of the highland in the Great Basin — from northeastern Nevada to southeastern California (pl. 9-G, sec. *U–U'*). This is the only interval of the Mississippian for which this evidence is conclusive. Coarse detritus is limited to a rather narrow belt along the east margin of the highland where the detritus accumulated as a series of coalescing deltas (pl. 11, fig. 3) deposited by east-flowing steep-gradient streams. East of the deltas, but within the troughs, basinal deposition in moderately deep water continued from the Early Mississippain into interval C and, locally, was associated with deltaic sedimentation.

Local limestone boulder conglomerate beds in the upper part of the Perdido in the northern Inyo Mountains of southeastern California (Calif. locs. 101, 105) are interpreted as trough slope deposits derived from the southeast along the northeastern edge of the carbonate shelf as in interval B (p. 297). Chert pebble-and-granule conglomerates found farther west on the east side of Owens Valley in northern Inyo County (Calif. loc. 159) may be basin deposits derived from the Antler highland to the west and laid down in the southern segment of the Eleana trough. C. H. Stevens and A. P. Ridley (written commun., 1972), however, cited evidence for these chert pebbles and granules being derived from the southeast and, hence, believed that they cannot be considered as evidence for a western highland in southeastern California. Sediment contributed to these areas by the Antler highland, if any, was mostly fine grained and basinal.

Thick accumulations of fine clastic sediments, mainly siltstone, sandstone, calcareous mudstone, and minor conglomerate in the Pequop Mountains (Nev. locs. 9, 9a, 10), were probably winnowed from the coarser detritus in the Chainman–Diamond Peak trough to the west. Sadlick (1965, p. 45–46) suggested that the well-sorted and evenly bedded lower siltstones and sandstones assigned to his *Needle Siltstone Member* of the Chainman Shale were deposited in a low-energy neritic marine environment relatively far from a shoreline. Abundance of benthonic, nonburrowing brachiopods indicates deposition within the photic zone in water less than 200 feet deep. The overlying very thick bedded medium- to dark-gray siltstone and claystone of the lower part of the *Camp Canyon Member* of the Chainman were deposited in relatively unagitated bottom waters, which may have been inhabited by nematode and nemertean worms (Sadlick, 1965, p. 54). The succeeding siltstones and sandstones of the lower part of the *Camp Canyon Member*, also assigned to interval C, contain abundant carbonaceous films of plants and exhibit common laminations and ripple marks indicative of a deltaic environment (pl. 11, fig. 3; Sadlick, 1965, p. 56).

Fine-grained siltstones, shales, and sandstones of the lower part of the Chainman Shale in the area surrounding the Ely arch probably were derived both from winnowing of fines from the detritus shed from the Antler highland and from the arch itself. Even though an extensive erosional hiatus is recognized between intervals B and C in the area of the Ely arch, it seems that limestones of interval B exposed on the emergent areas of interval C contributed only fine clastic detritus to the surrounding areas of marine sedimentation. The basal Chainman Shale of interval C is composed largely of quartzose siltstone; coarser sediments are rare. The arch must have been an area of low relief which was only partly submerged in latest interval C time (pl. 12, fig. 3). Available faunal evidence indicates that interval C rocks located near the flanks of the arch are probably of latest Meramec age (Gordon, in Brew, 1971; Sadlick, 1965; Gordon, in Hose and Repenning, 1963).

The limestone and sandstone and phosphatic shale of the Deseret limestone and of the Humbug and Woodman Formations of central and northern Utah contain a more complete record of interval C time than does the basal Chainman Shale of the Confusion Range area. The phosphatic shale unit at the base of the Deseret Limestone, formerly considered to be as young as middle Meramec (Sando, Mamet, and Dutro, 1969, p. E10, E21), or early Meramec (Morris and Lovering, 1961, p. 99), is now correlated with rocks of Osage age (Sando and others, 1976; Sandberg and Gutschick, 1976; Roberts, chap. N, this paper). The environment of deposition of the phosphatic shale is discussed here

because the unit is included in the construction of plate 11. In the recent past, the depositional environment of the phosphatic shale member was believed to have been similar to that postulated for the Phosphoria Formation (McKelvey and others, 1953; Sheldon and others, 1967) but of much shorter duration (Morris and Lovering, 1961, p. 103). According to Morris and Lovering (1961, p. 103), the interbedded carbonate and phosphatic shale

> probably accumulated on a shelving platform or in an embayment of rather large size. The absence of coarse detrital sediments * * * indicates that the bordering land masses were relatively low and contributed only small amounts of fine-grained quartz and clay. The increase of clastic sediments in the Deseret Limestone [to the] northwest * * * however, points to a land mass of at least small size in this direction. The abundance and preservation of carbonaceous material suggests * * * an oxygen-deficient depositional environment.

More recently, a deepwater starved basin origin for the phosphatic shale member of the Deseret Limestone has been proposed (Sandberg and Gutschick, 1976; Poole and Sandberg, 1977). The starved basin is interpreted to have been located just west of a Mississippian carbonate shelf margin in the East Tintic and Oquirrh Mountains (Rose, 1976), and the phosphatic sediments of the basin intertongue with flysch-derived siltstone to the south in the Confusion and Needle Ranges, and to the west in the Dugway Range (Sandberg and Gutschick, 1976; Poole and Sandberg, 1977).

Above the thin phosphatic shale member, more than 1,500 feet of Meramec fossiliferous limestone including some cherty dolomite, calcareous sandstone and siltstone, and minor shale make up the main bulk of interval C in northwestern Utah. Lithofacies patterns (pl. 5-B) seem to indicate that the relatively low-relief Wendover highland with lower Paleozoic rocks exposed in its core was the main source of sand and silt in the Deseret Limestone and Humbug and Woodman Formations. Evidence in support of this inferred facies pattern was presented by Livingston (1955) who cited ripple marks, regional thinning to the east, crossbedding attitudes, and increasing grain size to the west as evidence of a western source for detritus in the Humbug Formation. Crossbedding in the sandstone beds indicates that much of the Humbug was deposited in a high-energy, perhaps nearshore, environment (pl. 11, fig. 3). Some disagreement remains, however, about whether the highland actually was emergent during interval C time. It is possible that interval C rocks were deposited across the area of the highland and subsequently removed during Pennsylvanian and Permian time or that the detritus of the Humbug and Woodman was derived from the west but from the Antler highland (M. D. Crittenden, Jr., written commun., 1973). An emergent Wendover highland is preferred in this report.

The carbonates assigned to interval C, which extend from northwestern Utah southwestward through southern Nevada into southern California, are abundantly fossiliferous. They contain common bryozoans, corals, brachiopods, calcareous foraminifera, conodonts, and algae which indicate a warm shallow subtidal marine shelf environment (pl. 11, fig. 3).

Bedded and nodular chert forms more than 10 percent of interval C carbonate rocks in several areas just east of detrital rocks in southeastern California (Calif. locs. 101, 107), southern Nevada (Nev. locs. 156, 138), and western Utah (Utah locs. 303, 111). Elsewhere, areas of carbonate rocks contain smaller amounts of chert. The spatial distribution of abundant chert (pl. 5-B) suggests that upwelling currents from moderately deep waters of the Chainman–Diamond Peak and Eleana troughs were the source of silica for the shallow carbonate shelf areas to the east. The high concentrations of chert seem to occur near the western edge of the carbonate shelf. Disrupted trends of chert concentrations in interval C carbonate rocks near the Nevada-California border are probably related to the broad zone of Cenozoic right-lateral strike-slip faulting that offsets the Eleana trough in southwestern Nevada and southeastern California (Stewart and others, 1968). Abundant chert in the Wah Wah Mountains (Utah loc. 303) in southwestern Utah and the Deseret Peak area in northern Utah (Utah loc. 111) probably was derived from similar upwelling currents which originated farther west in deeper parts of the shelf sea.

PALEOTECTONIC IMPLICATIONS

Interval C was a time of crustal instability and change in tectonic patterns of the Great Basin. Relatively large volumes of andesitic and basaltic volcanic rocks were extruded near the western and, perhaps, on the northeastern flanks of the Antler highland. These volcanics are considered by Burchfiel and Davis (1975, p. 369) to be evidence for post-Antler eastward subduction of oceanic basement. The highland itself was intermittently active along its entire length. The relatively thin Lower Mississippian carbonate deposits along the eastern edge of the highland initially were uplifted and eroded and finally buried under a flood of interval C detritus that was shed eastward from the rising highland from north-central Nevada to southeastern California (pl. 10, fig. 3).

The Chainman–Diamond Peak and Eleana troughs subsided more during interval C time than in previous intervals, and a clastic wedge 3,000 feet to perhaps more than 5,000 feet thick accumulated in northeastern Nevada (pl. 10, fig. 3; pl. 9-G, sec. R–R').

A large part of eastern Nevada and adjacent areas of west-central Utah, which were sites of shelf sedimentation during intervals A and B, became part of an exten-

sive north-northeast-trending elongate low positive area, the Ely arch, during interval C time (pl. 9-G, secs. *S–S'*, *U–U'*). The shelf carbonates of earlier intervals were uplifted and locally folded and eroded to form a subaerial surface of low relief. Late in interval C time, parts of the arch in west-central Utah and eastern Nevada were submerged, and thin blankets of residual fine-grained detrital sediments covered them. Possibly, these locally derived sediments blanketed more of the southern part of the arch than is shown on plate 3-A.

The Wendover highland, as interpreted here, remained emergent during interval C time and contributed large volumes of detritus to the northwestern Utah marine shelf, particularly during the latter part of the interval (Humbug Formation).

An orogenic phase — the "Wendover phase" of the Antler orogeny — is recorded in the Silver Island Range where northeast-trending folds involving interval B rocks (pl. 10, fig. 3) are overlain unconformably by the Chainman Shale (Schaeffer and Anderson, 1959; Sadlick and Schaeffer, 1959). Comparable large broad gentle folds in interval B rocks are also present on the Ely arch in the Confusion Range area where the Joana Limestone has been removed from crests of anticlinal folds by pre-Chainman erosion (R. K. Hose, written commun., 1965).

An area of moderate subsidence is hypothesized to have developed between the Ely arch on the south and the Wendover highland on the north. This depressed area, or strait, would have provided a connection for the upwelling cold phosphate- and silica-rich waters of the Chainman–Diamond Peak trough and the warm shallow marine waters of the northern Utah carbonate shelf during at least early and middle interval C time.

Subsidence, which allowed accumulation of 1,000 feet of carbonate sediment, took place in a small elongate area east of the Ely arch in southwestern Utah. This trough and a small positive area just east of it, which is probably a post-Mississippian erosional feature, have been telescoped by folding and thrusting along the Sevier orogenic belt.

PALEOGEOGRAPHIC INTERPRETATIONS

During interval C time, volcanic islands (pl. 12, fig. 3) were present in northern California in the Klamath Mountains (Yates, chap. Q) and possibly in the Taylorsville area. The Antler highland from north-central Nevada to southeastern California stood as a series of high mountain peaks with local relief of several thousand feet (pl. 12, fig. 3). Volcanic islands, perhaps part of an island arc system, bordered the highland on the west. Some volcanic islands also may have been present near the northeastern edge of the highland in northern Elko County, Nev. Along the east front of the highland from northeastern Nevada to southeastern California coalescing deltas were formed by east-flowing steep-gradient streams. Deltaic sedimentation took place simultaneously and locally alternated with brief marine transgressions and emplacement of turbidites in the Chainman–Diamond Peak and Eleana troughs. The Ely arch stood as an elongate emergent area of low relief with local elongate fold-belt hills east of the Chainman–Diamond Peak and Eleana troughs. Early in interval C, a relatively deepwater basin connected with the Chainman–Diamond Peak trough, may have existed north of the Ely arch and south and west of the Wendover highland.

The Wendover highland during interval C time probably consisted of low hills with a maximum relief of less than 1,000 feet. Sand and mud derived from the Wendover highland were shed into the shallow sea which occupied the general area of the Oquirrh basin. Some deltas formed in the vicinity of the Promontory and northern Lakeside Mountains. In the southeastern Great Basin, bedded and nodular chert that formed along the western edge of the carbonate shelf of the Eleana trough was offset along the California-Nevada border by post-Mississippian strike-slip faults. Southeast of this outer shelf area of chert accumulation, a more unstable shelf area covered by warm shallow marine waters extended into eastern Utah and northwestern Arizona (pl. 12, fig. 3).

INTERVAL D

FORMATIONS INCLUDED AND STRATIGRAPHIC RELATIONS

SOUTHEASTERN CALIFORNIA

Interval D in southeastern California consists of the Rest Spring Shale, Lee Flat Limestone, and Chainman Shale (pl. 15). These formations are exposed in the Panamint Range (McAllister, 1952; Hall and Stephens, 1962, 1963; Pelton, 1966; Hunt and Mabey, 1966), the Inyo Mountains (Ross, 1962, 1966; Merriam, 1963; Nelson, 1966), near Darwin southeast of the Inyo Mountains (Hall and MacKevett, 1962), the Argus Range (Hall and Stephens, 1963), the Last Change Range (Stewart and others, 1966; Pelton, 1966), and in nearby areas. Gordon (1964, p. A5) described Chester cephalopod faunas from several of these localities, and stated that "The Chainman Shale of the Inyo Range [Mountains] appears * * * to be approximately equivalent to the Rest Spring Shale of the Panamint Range, and represents only a small part of the Chainman Shale of eastern Nevada." The Rest Spring Shale and the Lee Flat Limestone may contain some Lower Pennsylvanian beds according to Hall and MacKevett (1962, p. 20).

A part of the recrystallized Furnace Limestone in the San Bernardino Mountains (Dibblee, 1964a, b) is ar-

bitrarily assigned to interval D, as well as to all other Mississippian intervals, because Early(?) Mississippian fossils were found by Richmond (1960, p. 15–18) in the lower part of the formation (Dibblee, 1964a). (See p. 282.) Interval D is absent in the Ivanpah area (Hewett, 1956) where beds of Pennsylvanian and Permian age of the Bird Spring Formation rest unconformably on the Monte Cristo Limestone of interval C. Metasedimentary rocks in the western Mojave Desert believed to contain strata of Carboniferous(?) age include the Oro Grande and Bean Canyon Formations and related rocks in the Tehachapi Mountains and southern Sierra Nevada (Dibblee, 1967, p. 28–31). Other metasedimentary rocks tentatively assigned to the Carboniferous are the *Riggs Formation* of Kupfer (1960) in the Silurian Hills (Calif. loc. 115) and the formation at Carbide (Calif. loc. 117) near Lane Mountain (McCulloh, 1960).

NEVADA

In northwestern Elko County, north-central Nevada, two different lithologic suites, a volcanic and a detrital, are well-dated interval D sequences. The first suite consists of the interbedded chert, mudstone, conglomerate, and basaltic volcanic rocks of the *Schoonover Formation* (Fagan, 1962). The second suite consists of chert-pebble and quartzite-pebble conglomerate and sandstone assigned to the Diamond Peak Formation (Nev. loc. 22) in the northern Adobe Range (Ketner, 1970a).

The *Schoonover Formation* (Nev. loc. 9) consists of a sequence of turbidites, some radiolarian cherts, and basaltic volcanic rocks derived from "a tectonic land not far to the west" (Fagan, 1962, p. 595), and subsequently thrust southeastward over Paleozoic miogeosynclinal rocks. Fagan (1962, p. 595) suggested that the thrusting may be as old as Late Mississippian, but later work by Churkin and Kay (1967, p. 664) demonstrated that the thrust on which the *Schoonover* was transported may truncate Permian miogeosynclinal rocks. Fossils collected 250 feet above the base of the Schoonover have been identified as Late Mississippian (Chester) by Gordon (in Churkin and Kay, 1967, p. 663). The *Schoonover* has a maximum thickness of 9,000 feet and was thought by Churkin and Kay (1967, p. 664) to include rocks of Pennsylvanian and Permian age. Recent faunal information (R. R. Coats, oral commun., 1972) indicates that the entire formation is of Chester age. The *Schoonover* rests unconformably on eugeosynclinal rocks of the Ordovician Valmy Formation. Both the Valmy and Schoonover are part of an allochthonous block.

A thick sequence of autochthonous interbedded calcareous sandstone, siltstone, and limestone above quartzite-pebble conglomerate which has been described in Elko County, is faulted against the allochthonous *Schoonover Formation* (Nev. loc. 4). This unnamed sequence was assigned a Carboniferous(?) and Permian age by Churkin and Kay (1967, p. 663). Probable Permian brachiopods were identified by Gordon (in Churkin and Kay, 1967, p. 664) from 1,000 feet above the base of the conglomerate which rests unconformably on Cambrian limestone. Some of the sequence, which includes calcarenitic limestone, shale, and about 200 feet of dark chert, has been correlated with part of the Bird Spring Limestone of late Chester age by J. W. Huddle on the basis of conodont faunas (R. R. Coats, written commun., 1973) and is included in interval D.

The upper part of the Diamond Peak Formation in the Adobe Range (Nev. loc. 22) contains both Late Mississippian and Early Pennsylvanian fossils (Ketner, 1970a), but the position of the systemic boundary has not been established in the Adobe Range section. Thus, the estimated thickness of the entire upper part of the Diamond Peak is included in interval D. The Diamond Peak in the Adobe Range consists mostly of chert and quartzite pebble-and-boulder conglomerate with minor sandstone and siltstone.

Upper parts of the Chainman Shale and Diamond Peak and *Tonka* Formations are assigned to interval D in central and eastern Nevada (pl. 15, cols. 118, 119). The usage of the name *White Pine Shale* of Hague (1883, 1892) has been largely abandoned following the suggestion of Nolan, Merriam, and Williams (1956, p. 56–57), though relevant arguments for its retention locally have been presented by Langenheim (1963) and Humphrey (1960). The Chainman Shale and Diamond Peak Formation are treated as an undifferentiated sequence in much of the Eureka–Pinon Range area (Nolan and others, 1956, p. 60; Smith and Ketner, 1968, p. I13); however, in the northern Pinon Range (Nev. loc. 1), the formations can be separated, and in that area the upper parts of the Diamond Peak and *Tonka* Formations (1,560 feet thick) are assigned to interval D (pl. 15, col. 121). Conglomerate, composed of subrounded to subangular pebbles and cobbles of quartzite and chert as much as several inches across, comprises 40–60 percent of interval D at this locality (J. F. Smith, Jr., and K. B. Ketner, written commun., 1965; Smith and Ketner, 1968, p. I13). In the type section of the Diamond Peak Formation in the Diamond Mountains (Nev. loc. 226) near Eureka, Brew (1961b, 1964, 1971) has divided the formation into eight local mappable members, designated A through H (pl. 15, col. 119). Members C through H, which have an aggregate thickness of 1,975 feet, are assigned to interval D on the basis of calcareous foraminfer, brachiopod, coral, and cephalopod faunas of Chester age (Skipp and Mamet, 1970; Gordon, in Brew, 1971). Conglomerates are com-

mon in members C, E, G, and H and make up about 15 percent of the Chester part of the sequence. A part of the overlying Ely Limestone, as much as 80 feet thick, contains a latest Mississippian fauna (Gordon, in Brew, 1971), but this thin sequence has not been included in interval D.

In the northern Pancake Range (Nev. loc. 52), undifferentiated Chainman Shale and Diamond Peak Formation have been divided into four informal units (Stewart, 1962, p. C58–C59). The upper two units, which have a total minimum thickness of about 1,250 feet (top of the section is eroded), consist of more than 75 percent siltstone and claystone, minor sandstone, and less than 1 percent conglomerate. These upper two units have been arbitrarily assigned to interval D herein, although Gordon (in Brew, 1971) reported that at least the lower 200 feet of the upper siltstone unit is Meramec in age. The part of the Chainman–Diamond Peak sequence in the White Pine Range (Humphrey, 1960, p. 37; W. H. Easton, written commun., 1965) assigned to D is predominantly shale, sandstone, and siltstone with minor conglomerate near the base of the Diamond Peak. In the southern Diamond Mountains, Pancake Range, and White Pine Range, the Mississippian-Pennsylvanian systemic boundary lies within the lower member of the Ely Limestone (Gordon, in Brew, 1971), but these thin Ely Limestone beds are excluded from interval D in this report because of insufficient thickness data.

In northern Nye County, interval D strata are poorly known. No definite Chester faunas have been reported from coarse clastic rocks assigned to either the Diamond Peak (H. W. Dodge, Jr., oral commun., 1969) or Eleana (Kleinhampl and Ziony, 1967) Formations in the Monitor and Hot Creek Ranges. Quartzite, sandstone, and conglomerate make up about 50 percent of the upper part of the Mississippian clastic sequence in this area, and about half of the estimated thicknesses for these faulted and incomplete sections have been arbitrarily assigned to interval D. A Chester microfauna containing eostaffellid and cornuspirid foraminiferal forms has been recovered from a sandy limestone collected by F. J. Kleinhampl (Nev. loc. 166) in a sequence of shales north of Morey Peak in the northern Hot Creek Range.

In the areas east of the belt of coarse clastic deposits of the *Tonka*, Diamond Peak, and Eleana Formations, the siltstone, shale, sandstone, limestone, and minor conglomerate of the Chainman Shale compose interval D. Sadlick (1965) has divided the Chainman Shale of the southern Pequop Mountains (Nev. loc. 9) into five formal lithologic members and two informal lithologic units. The upper three members, which total about 1,200 feet in thickness, are considered Chester in age. Following are brief descriptions of Sadlick's divisions of the Chainman Shale in the southern Pequop Mountains.

The uppermost member (*Jensen*) consists of about 450 feet of interbedded calcareous siltstone, limestone, and quartzitic sandstone with minor conglomerate. This member contains the late Chester zonal fossils *Cravenoceras hesperium* and *Rhipidomella nevadensis*. The brachiopod, *R. nevadensis* also occurs in the lower 50 feet of the overlying Ely Limestone (Sadlick, 1965, p. 96), and thus the Mississippian-Pennsylvanian boundary is thought to lie within the lower beds of the Ely Limestone. The *Jensen Member* is believed to be a time-stratigraphic equivalent of the Scotty Wash Quartzite in southeastern Nevada.

The *Willow Gap Limestone Member*, which underlies the *Jensen* gradationally, consists of about 440 feet of ledge-forming bioclastic limestone interbedded with slope-forming brownish siltstone. The *Willow Gap* is characterized by the Chester fossils *Spirifer brazerianus, Caninia* cf. *C. nevadensis,* and *Eumorphoceras.*

The *Donner Member* (about 300 feet thick) is gradational with the overlying *Willow Gap* where present and is gradational both laterally and vertically with the upper part of the *Camp Canyon Member.* The *Donner* consists mostly of chert-pebble conglomerate and sandstone with minor limestone. The chert pebbles probably were derived from the Ordovician Valmy Formation (Roberts, in Sadlick, 1965, p. 75) to the west. The late Chester (Namurian) brachiopods, *Brachythyris chesterensis* and *Spirifer brazerianus* occur in this member. The *Donner Member* is a clastic wedge that thins to a zero edge about 70 miles east of the Antler highland (Sadlick, 1965, p. 76) and probably thickens westward into the Diamond Peak and *Tonka* Formations.

The Meramec-Chester Series boundary in the northern Pequop Mountains (Nev. loc. 9) is believed to lie within the top 171 feet of the underlying lower part of the *Camp Canyon Member* (Sadlick, 1965, p. 142). Thorman (1970, p. 2422) allocated about 300 feet of siltstone and sandstone in the lower part of the Chainman to the Meramec. Because the exact location of the series boundary in this area is in doubt, the 171 feet in Sadlick's section is included in interval C (including the basal *Needle Siltstone Member*). About 15 miles north of Nevada locality 9, 130 feet of beds assigned to the Diamond Peak Formation by Thorman (1962, 1970) has been correlated with the *Donner Member* of the Chainman Shale by Sadlick. These Upper Mississippian rocks are overlain unconformably by the Permian *Pequop Formation* (Thorman, 1962, p. 72). Lithologic correlatives of the *Willow Gap Limestone* and *Jensen Members* have been removed entirely by post-Ely erosion in this area. North of the Pequop Mountains, in the Windermere Hills (Nev. loc. 25), Oversby (1969, 1972)

has estimated the thickness of the Diamond Peak Formation to be about 4,000 feet. Half of this thickness is assigned to interval D. South of the Pequop Mountains, in the Cherry Creek Mountains (Nev. loc. 5), more than 2,000 feet of undivided Chainman Shale and Diamond Peak Formation has been arbitrarily assigned to interval D. Cross section *U–U'* (pl. 9-G) shows the intermediate position of this section between thick interval C strata in the Pequop Mountains and absence of interval C strata in the northern Egan Range to the south. It seems possible that more of the Chainman in this area may prove to be of Meramec age and, if so, interval D thicknesses will be reduced.

In the Egan and Schell Creek Ranges, the entire Chainman Shale is assigned to interval D. The Chainman in these ranges consists of dark-gray clay-shale and minor amounts of sandstone, conglomerate, limestone, and traces of gypsum. The Chainman lies unconformably on carbonate rocks of the Joana Limestone assigned to interval B (Gordon, in Brew, 1971; Drewes, 1967, p. 40). In the Egan, Schell Creek, and Snake Ranges, according to Drewes (1967, p. 43), "the lowest few tens of feet of [Ely Limestone] limestone contain Late Mississippian fossils." As no further quantitative estimates of the thicknesses of these Upper Mississippian transitional beds are available in the central and northern parts of these ranges, the thin limestone-bearing units are included in the overlying Ely Limestone and thus are excluded from interval D. In the southern Egan and southern Schell Creek Ranges, however, thicknesses have been published for the Upper Mississippian "red limestone facies" of the Ely Limestone; therefore, these Mississippian rocks are included in interval D (Tschanz and Pampeyan, 1970).

At the eastern edge of Nevada in the southern Deep Creek Mountains, Kern Mountains, and northern Snake Range, two different interval D lithologies are juxtaposed by thrust faults (Nelson, 1966, p. 924–936). The autochthonous facies of the southern Deep Creek Mountains consists of calcareous shale and limestone similar to the Manning Canyon Shale and Ochre Mountain Limestone of the Gold Hill district in Utah. The allochthonous facies or assemblage in the northern Snake Range and Kern Mountains (Nev. loc. 68) consists of shale and minor limestone and siltstone referred to as Chainman Shale, which also is recognized in the southern part of the Snake Range (Whitebread, 1969). Both assemblages are finer grained than any lithologies reported to the north and west in east-central Nevada.

In southern Nevada in the Nevada Test Site, interval D consists of more than 1,600 feet of argillite interbedded with sandstone and limestone which makes up upper unit H, and units I and J of the Eleana Formation (Poole and others, 1961, p. D104; Orkild, 1963; Poole and others, 1965, p. 53). The contact of the Eleana with the overlying Tippipah Limestone coincides with the Mississippian-Pennsylvanian systemic boundary in the vicinity of the test site and is marked by an erosional hiatus representing early and middle Morrow time (Gordon and Poole, 1968, p. 157–159; Gordon, 1969, p. C1). The upper part of unit H of the Eleana Formation contains earliest Chester cephalopods (F. G. Poole, oral commun., 1972) and unit J of the Eleana contains late Chester (early Namurian) cephalopods including *Eumorphoceras* sp., *Cravenoceras hesperium*, and *Cravenoceras merriami*.

Siltstones and silty claystones with interbeds of conglomerate, sandstone, limestone, and chert — similar to strata in units H, I, and J of the Eleana Formation — have been called the *Meiklejohn Formation* in southern Nye County (Cornwall and Kleinhampl, 1961, 1964) and southernmost Esmeralda County (Albers and Stewart, 1965). Chester microfaunas have been recognized from the upper 1,400 feet of the *Meiklejohn* in Nye County, and a late Viséan (early Chester) fauna occurs in similar strata in Esmeralda County (Mackenzie Gordon, Jr., written commun., 1962). These upper beds of the *Meiklejohn* and correlative beds in Esmeralda County are included in interval D.

Interval D, where recognized in Lincoln and Clark Counties in southeastern Nevada, consists of a variety of lithologies and formations. These include the Scotty Wash Quartzite and "red limestone facies" of the overlying Ely Limestone in the southern Egan and southern Schell Creek Ranges and Pioche region (Kellogg, 1960, 1963; Westgate and Knopf, 1932; Langenheim and Peck, 1957; Tschanz and Pampeyan, 1970) in northern Lincoln County; Peers Spring Formation, upper part of the Chainman Shale, Scotty Wash Quartzite, and member A of the Bird Spring Formation in the Pahranagat Range (Reso, 1963) in southwestern Lincoln County; the upper part of the *Battleship Wash Formation* (Langenheim and Langenheim, 1965; Mamet and Skipp, 1970b; Brenckle, 1973), the overlying *Indian Springs Formation* (Webster and Lane, 1967, p. 503–516), and the basal part of the overlying Bird Spring Formation (Webster, 1969; Rich, 1961) in northern Clark County (pl. 15, col. 126); beds assignable to either the upper part of the Eleana Formation or the upper part of the Chainman Shale (Nev. loc. 137) in the southern Spotted Range (F. G. Poole, written commun., 1965; Longwell and others, 1965, p. 28) in northwestern Clark County; units BSa and BSb of the Bird Spring Formation (Langenheim and others, 1962, p. 603) in the Arrow Canyon Range (Nev. loc. 126) in north-central Clark County; and the *Bluepoint Limestone* of Longwell (1928; Longwell and others, 1965, p. 29–34)

in the Muddy Mountains (Nev. loc. 134) in southeastern Clark County.

In paleotectonic investigations of the Pennsylvanian System (McKee, Crosby, and others, 1975), the *Indian Springs* was treated as a member of the Bird Spring Formation as defined by Longwell and Dunbar (1936), and thus the upper part was assigned to Pennsylvanian interval A (Morrow). in the present paper the *Indian Springs* has been restricted to the basal clastic unit of the original *Indian Springs Member* of the Bird Spring Formation and redefined as a separate formation, thus raising the base of the original Bird Springs Formation (Webster and Lane, 1967). The newly defined *Indian Springs Formation* contains Chester faunas (Webster and Lane, 1967; Gordon and Poole, 1968) and is assigned entirely to Mississippian interval D.

Interval D strata are absent in southernmost Nevada (southern Clark County) where Pennsylvanian rocks of the Bird Spring Formation and Callville Limestone unconformably overlie the Monte Cristo Limestone (Hewett, 1931, p. 17–21; 1956, p. 42–43; Longwell and others, 1965, p. 26–29).

Interval D rocks have been recognized with certainly in just one area of Mississippian eugeosynclinal deposition in western Nevada (Humboldt and Pershing Counties). The thick Pumpernickel Formation, which consists largely of interbedded chert and argillite with minor limestone, sandstone, conglomerate, and greenstone, has been assigned a Pennsylvanian(?) age by Roberts (1964, p. A44). It may contain some Mississippian rocks in the lower part, but it has not been included in the Mississsippian of the present report. Recently, Late Mississippian (Chester) conodonts have been found in a limestone assigned to the Pumpernickel Formation in the Gold Run Creek quadrangle southwest of Nevada locality 281 in southeastern Humboldt County (R. L. Erickson, oral commun., 1973; C. A. Sandberg, oral commun., 1971; J. W. Huddle, written commun., 1972). The structural and stratigraphic relations of the limestone to the main body of the Pumpernickel are obscure, however, and only the limestone is assigned to interval D in this report.

Both the Goughs Canyon and Inskip Formations in western Nevada are thick (about 5,000–9,000 feet) sequences of dominantly volcanic rocks that contain Meramec faunas. Chester volcanic rocks are recognized farther west in the Baird Formation in northern California (Hotz and Willden, 1964, p. 28) and to the east in the *Schoonover Formation*; hence, it is possible that part of the thick volcanic sequence assigned to the Goughs Canyon and Inskip in western Nevada may be Chester in age.

WESTERN UTAH

In much of north-central Utah, the Great Blue Limestone and lower part of the overlying Manning Canyon Shale make up interval D (pl. 15, col. 111). These formations are recognized in the Promontory Mountains (Olson, 1960), Lakeside Mountains (Doelling, 1964; Young, 1955), Stansbury Mountains (Teichert, 1959; Rigby, 1958), southern Oquirrh Mountains (Gilluly, 1932; Bissell, 1959a), East Tintic Mountains (Morris and Lovering, 1961; Morris, 1964), southern Wasatch Range, and nearby areas to the west (Baker and Crittenden, 1961; Baker, 1947, 1964; Bullock, 1951; Rigby, 1952; Foutz, 1960), Gilson Mountains (Costain, 1960; Wang, 1970), and Sheeprock Mountains (Cohenour, 1959). Crittenden (1965a, b) mapped the Doughnut Formation in the general stratigraphic position of the lower part of the Manning Canyon Shale and the upper part of the Great Blue Limestone in the southern Wasatch Range. (See Roberts, chap. N, and Mallory, chap. M, this report, for further discussion of the Doughnut Formation.)

The Ochre Mountain Limestone (pl. 15, col. 112), defined in the Gold Hill district in the northern Deep Creek Mountains by Nolan (1930, 1935) and extended eastward into the Dugway Range by Staatz and Carr (1964), correlates, in general, with the Great Blue Limestone.

Morris and Lovering (1961, p. 107–113) divided the Great Blue Formation of the East Tintic Mountains into four members. These are, in ascending order: (1) the Topliff Limestone Member, (2) the Paymaster Member, which is largely limestone but which also contains abundant brown-weathering olive-green shales and quartzites, (3) the Chiulos Member, which is made up of black shale and olive-green quartzite sandstone (this member was first described in the Sheeprock Mountains by Cohenour (1959)), and (4) the Poker Knoll Limestone Member, which is composed almost entirely of thin-bedded argillaceous or silty limestone. The contact between the Poker Knoll Member and the overlying Manning Canyon Shale is gradational. The Long Trail Shale Member of the Great Blue Limestone, was defined in the southern Oquirrh Mountains by Gilluly (1932, p. 29). According to Morris and Lovering (1961, p. 112) the member appears to be "represented by one or more of the black shale beds found in the lower part of the Paymaster Member [in the East Tintic Mountains]."

In much of northern Utah, the Meramec-Chester series boundary lies within the Great Blue and Ochre Mountain Limestones (Weller, 1948; Arnold and Sadlick, 1962; Bissell, 1959a). Although Roberts (chap. N) has assigned the lower member of the Great Blue Limestone to interval C in the vicinity of Pocatello,

Idaho, and units 1 and 2 of the *Brazer Formation* to interval C in north-central Utah, time-stratigraphic correlatives in western and central Utah are herein assigned to interval D because no faunal and lithologic data were available to locate the series boundary.

Only the lower part of the Manning Canyon Shale (400–740 feet) is assigned to interval D in the Promontory Mountains, Oquirrh Mountains, East Tintic Mountains, Lake Mountains, southern Wasatch Range, and nearby areas. In the Lakeside Mountains, 1,000 feet of Manning Canyon is assigned to interval D, whereas none of the formation is included in interval D in the Gold Hill district (pl. 15, col. 112) where it totals approximately 500 feet in thickness and was considered to be of Pennsylvanian(?) age by Girty (Nolan, 1935, p. 31–33). The Manning Canyon Shale may be entirely of Mississippian age in the central and southern Oquirrh Mountains (Welsh and James, 1961; Tooker and Roberts, 1963), where it is more than 1,000 feet thick. In the northern Oquirrh Mountains, Tooker and Roberts (1963, p. E33) suggested that even the lower part of the overlying Oquirrh Formation locally may be Chester in age. In the Lakeside Mountains (Utah loc. 120), Sadlick (in Young, 1955, p. 33) recognized Chester (early Namurian) goniatites in the upper one-third of the Manning Canyon Shale and concluded that the entire formation — 1,500 to 2,500 feet thick in that area — may be of Mississippian age. Farther southeast in the southern Wasatch Range (Utah loc. 202), however, goniatite and brachiopod faunas studied by Gordon (in Baker, 1964) indicate that only the basal one-third of the formation is Mississippian (Chester). If more of the Manning Canyon Shale were to be included in interval D in the Oquirrh, East and West Tintic, Promontory, Lakeside, and Deep Creek Mountains, then fine-grained clastic rocks would make up nearly one-half of the total interval in most of the Oquirrh basin.

In northwestern Utah, thin stratigraphic units are assigned to interval D, which have been called undifferentiated Chainman Shale–Diamond Peak Formation in the northern Pilot Range by Blue (1960), in the southern Pilot Range by O'Neill (1967), in the southern Grouse Creek Mountains (Muddy Range) by Baker (1959), and in central Crater Island (Utah loc. 108) in the northern Silver Island Range by Anderson (1957).

Interval D strata are missing entirely in the Newfoundland Mountains (Utah loc. 106) and northern Silver Island Range north of central Crater Island (Utah loc. 107), where Pennsylvanian and Permian conglomerate has been reported to rest with angular unconformity on Devonian limestone and shale (Paddock, 1956; Anderson, 1957, 1960; Schaeffer and Anderson, 1959). Roberts and Tooker (1969, p. 69), however, suggested that this conglomerate is not of Pennsylvanian and Permian age but is instead a northwestward extension of the Devonian Stansbury Formation which is overlain by Permian carbonate rocks. The conglomerate is considered to be of Pennsylvanian and Permian age in the present report.

Elsewhere in most of western Utah, all or part of the Chainman Shale is included in interval D. From north to south, areas in which this terminology is used are: the Vipont Mountains northwest of the Grouse Creek Mountains (Stokes, 1963); the Wendover area southwest of the Silver Island Range (Sadlick, 1965); and the House, Confusion, Conger, and Needle Ranges (Campbell, 1951; Hose, 1963, 1965a, b; Hose and Repenning, 1963, 1964; Hose and Ziony, 1964; Sadlick, 1965).

Rocks of interval D are largely absent in areas south of the Needle Range in southwestern Utah. In the Wah Wah Mountains (Utah loc. 303), the uppermost 200 feet of an unnamed Mississippian limestone and shale is assigned to interval D here because *Caninia*, a Chester (Zone K) coral (Sando and others, 1969) is present above *Faberophyllum*, a Meramec (Zone F) coral (Miller, 1966).

THICKNESS TRENDS

Interval D is thickest in the north-northeast-trending Chainman–Diamond Peak and Eleana troughs of the Antler highland from east-central Nevada to southeastern California and in the Oquirrh basin area of northwestern Utah. A thickness of more than 2,500 feet (pl. 6-A) is preserved in the Chainman–Diamond Peak trough in the southern Diamond Mountains, Nev. (Nev. loc. 51). Elsewhere in the troughs, thicknesses of 1,500–2,000 feet are common.

Thicknesses of interval D strata in northwestern Utah in the Oquirrh basin are as much as 4,000–4,500 feet in the Deep Creek, Lakeside, and East Tintic Mountains. The thickest sequences are shown as two connected northeast-trending troughs (pl. 6-A) similar to those shown on an isopach map of the Mississippian System compiled by Foutz (1966, p. 69). However, the same thickness data could be isopached as a triangular-shaped basin with an east-trending southern margin, such as that interpreted by Bissell (1962, p. 194) and Armstrong (1968b, p. 30) for their total Mississippian isopach maps. As mentioned above, thicknesses of Great Blue and Ochre Mountain Limestones selected for interval D may be too great, whereas the thicknesses of Manning Canyon Shale may not be great enough. Even if 1,000 feet of strata is subtracted from the total thickness of interval D in this area, however, the general location and shape of the basin remain the same.

An elongate north-northeast- to south-southwest-trending area with thicknesses of 1,000 feet or less separates the Chainman–Diamond Peak trough on the west from the Oquirrh basin on the east. This inter-

mediate platform area extends southwestward from southern Idaho through the northwestern corner of Utah into east-central Nevada near the northern edge of Lincoln County. The small area of missing interval D strata in southwestern Box Elder County, Utah, is a remnant of the eroded Wendover highland (pl. 6-A). This area lies on the northern one-third of the linear feature that is a subdued reflection of the Ely arch and Wendover highland which were emergent areas during interval C time.

In southeastern Nevada and southwestern Utah, interval D thicknesses are generally 1,500 feet or less, diminishing eastward to zero in the vicinity of the Wasatch line (Roberts and others, 1965, p. 1928). The zero isopach extends southwestward across southeastern Inyo County into eastern San Bernardino County in southeastern California (pl. 6-A).

Interval D strata are not present in a large part of northern Elko County, Nev., where post-Ely erosion has removed all Mississippian rocks. Mississippian rocks are not present on the Antler orogenic belt, owing to nondeposition on this persistent positive area (Nolan, 1943; Roberts and others, 1958).

Thick accumulations of basaltic volcanic rocks interbedded with chert, mudstone, siltstone, sandstone, conglomerate, and limestone, which make up the allochthonous *Schoonover Formation*, are assigned to interval D in northwestern Elko County, Nev. This sequence has a maximum thickness of about 9,000 feet (Fagan, 1962).

In Humboldt County, Nev., west of the Antler highland, a limestone assigned to the Pumpernickel Formation has yielded a Chester fauna (p. 311). This is the only well-documented occurrence of interval D strata in this area, though it seems likely that a part of the thick (5,000–9,000 feet) volcaniclastic rocks of the Goughs Canyon and Inskip Formations may be correlative with the *Schoonover Formation* (interval D) east of the highland. Also, the Taylorsville area in northern California may contain some beds equivalent to part of the lower part of the Baird Formation of the Klamath Mountains in northern California. (See Yates, chap. Q, fig. 78.)

LITHOFACIES TRENDS

Conglomerate and coarse-grained sandstone along the northeastern edge of the Antler highland, which were derived from quartzites and cherts of lower Paleozoic transitional and eugeosynclinal rocks, grade eastward into finer sandstone, siltstone, and shale across most of eastern Nevada (pl. 6-B; pl. 9-G, sec. *R*–*R'*). These fine-grained detrital rocks are calcareous in easternmost Nevada and west-central Utah and grade northeastward into thick accumulations of impure limestone with as much as 40 percent interbedded fine-grained clastics in the Oquirrh basin of northern Utah. Thinner accumulations of limestone are present along the northeast-trending Wasatch line and contiguous Sevier orogenic belt in Utah (p. 278), southern Nevada, and southeastern California. The coarsest detrital rocks in interval D, which consist of cobble-and-boulder conglomerates, occur in the Pinon and Adobe Ranges in west-central Elko County, Nev. (Nev. locs. 1, 22). Sandstone with some pebble-and-granule conglomerate is common in the Diamond Mountains and Pancake Range in easternmost Eureka County and westernmost White Pine County, and in some areas of northern Nye County. In southern Nye County, Nev., and adjacent southeastern California, shale or argillite with interbedded chert, sandstone, and conglomerate make up the clastic sequence. In southeastern California, the shales grade southeastward into relatively pure carbonates over a distance of about 100 miles. In southeastern Nevada, shale and sandstone and minor limestone make up interval D strata except in the Muddy Mountains (Nev. loc. 134) where limestone is dominant (Longwell, 1928).

In the northern two-thirds of the Great Basin region, the lithologic progression from west to east of coarse clastics grading laterally into impure carbonates is complete and proceeds with only minor structural interruptions (Sadlick, 1965). Across the northernmost part of the area, clastic deposits flank the Wendover highland in northwestern Utah and extend east into north-central Utah (pl. 6-B). Interval D carbonates, contemporaneous with dominantly clastic sequences in the Pequop Mountains in northeastern Nevada and Grouse Creek Mountains in northwestern Utah, are recorded only in the subsurface (Utah loc. 103; Peace, 1956) east of the Grouse Creek Mountains.

The eastward lithofacies progression is much less obvious in the southern one-third of the Great Basin region in southern Nevada and southeastern California where no coarse clastics are recognized and where the fine-grained detrital and carbonate rocks are preserved in faulted incomplete sections.

Chert is not an important constitutent in interval D limestone, except in the Promontory Mountains, Utah (Utah loc. 102), where it forms a significant part of interval D strata (Olson, 1960).

Chert, mudstone, conglomerate, and basaltic volcanic rocks of the *Schoonover Formation* in northern Elko County, Nev. (Nev. loc. 4) rest unconformably on allochthonous Ordovician eugeosynclinal rocks and appear to have had a western source (p. 316; Fagan, 1962). In the present report, they are interpreted to have been deposited near the northeastern edge of the Antler highland and thus have not moved far from their depositional site. However, the possibility that part or all of the *Schoonover* sequence may have been

deposited north or west of the highland and thrust into its present position has not been ruled out. The chemistry of the basalts suggests deposition on oceanic crust (Rogers and others, 1974, p. 1918).

Volcanic rocks of interval D have not been recognized west of the Antler highland in Pershing and Humboldt Counties in northwestern Nevada, or at Taylorsville in northern California (Calif. loc. 156). It is possible that they are or were present, however, for eugeosynclinal rocks of interval D are found to the northwest in the Klamath Mountains (Yates, chap. Q), and an interval D limestone associated with the Pumpernickel Formation is present in southeastern Humboldt County (p. 311).

SOURCES OF SEDIMENT, ENVIRONMENTS OF DEPOSITION, AND CLIMATE

During interval D time the eastern Great Basin was dominated by a moderately shallow marine environment (pl. 11, fig. 4; Sadlick, 1965; Morris and Lovering, 1961; Brew, 1971) in which a large volume of detritus, derived from lower Paleozoic rocks exposed on the Antler highland, was deposited. The detrital rocks became progressively finer grained eastward, from central Nevada into western Utah, and southeastward in southern Nevada and southeastern California (pl. 9). In western Utah, interval D strata contain much limestone, even in areas of dominantly fine-grained detrital rocks, as in the Confusion Range (Utah locs. 226, 252, 253, 254, 256, 257).

The poorly sorted conglomerates of the Diamond Peak and *Tonka* Formations, which are composed of subrounded to subangular quartzite and chert pebbles, cobbles, and a few boulders, were derived from alluvial gravels (pl. 11, fig. 4) rapidly deposited in front of the active orogenic belt and subsequently redistributed in a marine environment in the Chainman–Diamond Peak trough by wave action, currents, and submarine slides (Brew, 1971, p. 25).

Interval D strata in the Diamond Peak Formation in the Diamond Mountains were interpreted by Brew (1964, p. 99) to have been

> deposited as a series of coalescing steep-fronted deltas in a basin that subsided rapidly enough to prevent extensive regression of the sea; [the deltas are composed of] material derived from a nearby tectonically active provenance terrane * * * by closely spaced steep-gradient streams.

He further stated (Brew, 1964, p. XVI) that

> The conglomerate, limestone, siltrock, and sandstone of members C and D are poorly winnowed sediments; conditions were unstable, but subsidence was less continuous than [in interval C]. The * * * siltrock, sandstone, conglomerate, and limestone phenoplast conglomerate of Members E, F, and G [consist of] alternations of winnowed and poorly winnowed sediments [indicating] conditions which changed repeatedly. The highest member H of the Diamond Peak Formation is transitional [with] the conformably overlying Ely Limestone of Pennsylvanian age. This member was deposited in an environment similar to that of the underlying three members, but winnowed sediments are more common, indicating longer periods of relative stability.

Much of the siltstone of members E, F, and G contains locally abundant diagenetic pyrite and pyrite casts interpreted by Brew (1964, p. 94) as

> indicative of a reducing bottom environment within the neritic zone. [Also], the local abundance of poorly preserved plant fragments suggest that land was not far away * * * The preservation of worm trails indicates deposition below wave base and rapid burial * * * A few of the poorly sorted thin sandstone beds exhibit poorly developed graded bedding, suggesting that turbidity currents deposited some of the sandstone layers.

He further stated (Brew, 1964, p. 94–96) that

> The poorly sorted chert and quartzite pebble conglomerates occur in extensive lenticular layers which are thought to represent coalescing bodies of coarse detritus deposited below the limit of effective wave action for the most part; * * * submarine slides [may] have played a part in transport.

The interbedded coarse and fine sediments of both the Chainman and Diamond Peak Formations have been recognized as flysch deposited in an elongate foreland basin (exogeosynclinal trough) by Poole (1974, p. 58), who interprets the flysch sediments as having been deposited in "a complex system of submarine slope-fan-basin floor environments."

A general succession of sedimentary environments was suggested by Sadlick (1965) for the Chainman Shale of eastern Nevada and western Utah. Rocks of the upper part of the *Camp Canyon Member*, the base of interval D in this paper, in northeastern Nevada were considered a fondothem facies — dark shale formed as bottom sediments in a neritic environment (Sadlick, 1965, p. 71). This shale is gradational laterally and vertically with coarse clastic rocks of the *Donner Member* in northeastern Nevada (p. 301; pl. 9-G, sec. *R–R'*). The *Donner* was deposited in a mixed fluvial and deltaic environment described by Sadlick (1965, p. 81) as a "series of coalescing deltaic plains." In northwestern Utah, the *Camp Canyon Member* grades both laterally and vertically into the *Willow Gap Limestone Member*, which is a limestone sequence considered analogous to hydroclastic limestone of the modern Bahama Banks that formed in marine waters no deeper than 10 fathoms (Sadlick, 1965, p. 88). The overlying *Jensen Member* was considered an alluvial coastal plain and a "normal marine" neritic facies, in part similar to the neritic limestone of the overlying Ely Limestone of Mississippian and Pennsylvanian age (Sadlick, 1965, table 4, pl. 1).

The Great Blue Limestone of the Oquirrh basin is mostly a shallow-water marine limestone, locally cherty, with some interbedded shale, sandstone, and siltstone. The thick shale units of the Chiulos and Long Trail Members may represent eastward extensions of detrital sediments derived from western tectonic lands

(pl. 9-G, sec. $R-R'$). Local reworking and redeposition of parts of the thick clastic sequence provided silt and sand for the impure limestone. Marine shale of the Manning Canyon Shale represents an influx of large amounts of terrigenous material very late in interval D time. Moyle (1959, p. 88) suggested that some of the carbonaceous shales were derived from mature soils developed on the craton to the east. Coarse clastic rocks in the Lake Mountains area probably had a local source and suggest that erosion of the Wendover highland and nearby interval D conglomerates in northwestern Utah and northeastern Nevada may have been sources of some detritus during much of interval D time. Moyle (1959, p. 84) also suggested that the Manning Canyon was deposited on an unstable shelf directly adjacent to deeper zones of the geosyncline. Nearshore tidal flat, and shallow neritic environments may have prevailed. One major marine recession, possibly the temporal equivalent of the Chainman–Ely sedimentation break, and one major marine transgression, equivalent to the lower Ely transgression, have been inferred from the strata in the lower part of the Manning Canyon.

In southeastern California, the correlative Rest Spring and Chainman Shales consist of gray and brown siltstone and shale with a few thin interbeds of light-gray quartzite and poorly sorted intraformational conglomerate. Pelton (1966) referred to these rocks as "the fine clastic lithosome" and reported that they do not change laterally as much as the underlying rocks of interval C.

Most of the fine detrital materials of the Rest Spring and Chainman Shales probably were not subjected to current winnowing and are the result of rapid deposition and burial. Pelton (1966, p. 73–75) noted that "The lack of extraformational conglomerate [in these rocks] suggests that the source had lost much of its relief during [earlier] denudation." The western source area for the fine clastic rocks of the Rest Spring and Chainman Shales, therefore, is postulated to be an area of low hills and plains (pl. 12, figs. 3, 4). Conglomerates in western Nye County in both intervals C and D suggest that the northwest highland was elevated in approximately the same areas during both Meramec and Chester times.

The Rest Spring and Chainman Shales grade southeastward (pl. 9-G, sec. $T-T'$) into pure to slightly silty and argillaceous limestone of the moderately stable shelf carbonate rocks of the Lee Flat Limestone (Pelton, 1966, p. 75) and, farther to the southeast, possibly into the relatively pure recrystallized shelf limestones of the Furnace Limestone.

Interval D strata are missing in much of southeastern California and southeasternmost Nevada. The absence of coarse detritus in interval D rocks in the southern Great Basin and nearby areas, plus the indication that sandstones and shales in these areas have graded southeastward into limestone, such as the *Bluepoint Limestone* of Longwell (1928), suggests that as much as 500 feet of D may have been removed by erosion during earliest Pennsylvanian uplift in southern California and Nevada (pl. 9-G, sec. $U-U'$).

Detailed studies of calcareous Foraminifera from the *Battleship Wash Formation* (Langenheim and Langenheim, 1965) and *Indian Springs Formation* Webster and Lane, 1967) in the Arrow Canyon Range in southern Nevada (Nev. loc. 126) have shown that a major faunal hiatus (fig. 76) representing all of late Visean time (early Chester) is present within the *Battleship Wash Formation* (Mamet and Skipp, 1970b; Brenckle, 1973). This hiatus probably represents a shoaling of the seas in this area and an interruption of sedimentation accompanied by significant erosion (fig. 76). Two minor hiatuses were recognized by Webster (1969, p. 24) in the Arrow Canyon Range area — one between the limestone of the *Battleship Wash Formation* and the interbedded shale, limestone, and orthoquartzite of the overlying *Indian Springs Formation*, and the other at the base of the "marker" conglomerate in the *Indian Springs*. Webster (1969, p. 24) also reported that the evenly bedded fine-grained character of the sediments of the *Indian Springs Formation*, the lack of channeling in the sediments, the diversity of faunas, and the pervasive occurrence of sparse fragmented plant material all suggest a low-energy warm shallow-water marine environment close to land. The lithofacies patterns of plate 6-B, however, suggest that interval D detrital materials in southern Nevada were derived either from reworking of contemporaneous deposits to the north and west, or, possibly, from a low positive area to the northwest. Shallow-water marine limestones are dominant to the east and southwest of this area. Sadlick (1965, p. 30) reported that ripple marks in the Scotty Wash Quartzite suggest an open sea east of Lincoln County and indicate that the detritus in the Scotty Wash came from a western source.

In northwestern Elko County, Nev., the *Schoonover Formation* contains thick sequences of basaltic flows and tuff-breccias (Coats, 1969, p. A25; Fagan, 1962, p. 595–610). The volcanic rocks of the *Schoonover* are interbedded with fine-grained argillaceous limestones, bedded cherts, quartzose sandstones, and conglomerates. Fagan (1962, p. 595) interpreted the sandstones as "turbidites intermittently deposited by currents from the west." In thin section, the chert appears to be a pelagic accumulation composed mostly of radiolarian fragments. According to Fagan (1962, p. 595), "layers of pebbly mudstone record chaotic submarine slides into the relatively deep-water site of sedimentation. The source of the turbidites and slide deposits was a tec-

tonic land not far to the west, composed of rocks formed in the eugeosyncline."

This fragmentary record of interval D volcanism suggests that volcanic activity may have been common along the northeastern edge of the Antler highland near the Nevada-Idaho border. Glass spherules of possible volcanic origin have been reported in a Lower Mississippian limestone north of the Snake River Plain in Idaho (Thomasson, 1959, p. 225–227). No volcanic detritus has been reported in the contemporaneous Diamond Peak Formation to the south; therefore, the possibility remains that the *Schoonover* was deposited north or west of the Antler highland and thrust into its present position in post-Mississippian time. The paleogeographic reconstruction (pl. 12, fig. 4) proposing a relatively deepwater site of deposition for the *Schoonover* is based on a hypothetical paleogeography developed for the *Mikes Creek Member* by Fagan (1962, fig. 10, p. 609). Although the shale, chert, and greenstone in the lower part of the Pumpernickel Formation in the Sonoma Range in north-central Nevada (Roberts, 1964, p. 37–44) may record interval D volcanism and deepwater sedimentation west of the Antler highland in northwestern Nevada, faunal evidence is inconclusive, and the main body of the Pumpernickel Formation is assigned to the Pennsylvanian in this report. A bioclastic limestone containing Chester conodonts, which has been assigned to the Pumpernickel, is included in interval D in this report (p. 311).

A warm climate in the Great Basin region during interval D time is indicated by the presence of limestone with diverse megafaunal and microfaunal assemblages — corals, brachiopods, cephalopods, crinoids, mollusks, foraminifers, ostracodes, conodonts, and algae.

PALEOTECTONIC IMPLICATIONS

During interval D time, warm shallow seas covered the miogeosynclinal area of southern California, eastern Nevada, and western Utah. Emergent, actively rising tectonic highlands (pl. 10, fig. 4) in the Antler orogenic belt contributed large volumes of coarse detritus to an elongate, irregularly subsiding trough along the northeastern margin of the highland. The detritus is composed largely of fragments of lower Paleozoic eugeosynclinal quartzite, chert, and argillite that were exposed on the highland. Detritus shed from the Antler highland and the satellitic Wendover highland were deposited by turbidity and other submarine currents as far east as the Tintic area in Utah. The trough area adjacent to the highland locally was occupied by large coalescing deltas formed by the steep-gradient streams which eroded the highlands.

Positions of highlands and troughs were well established in the Great Basin before interval D time (pl. 10). A late Visean–early Namurian (late Chester) period of renewed uplift and erosion of the northern part of the Antler highland is recorded in a coarse eastward-thinning clastic wedge made up of conglomerates of the *Tonka* and Diamond Peak Formations and the *Donner Member* of the Chainman Shale (pl. 9-G, secs. *R–R', S–S'*). Evidence for this late Chester period of intense diastrophism is found as far north as Mackay, Idaho, in the White Knob Limestone (Skipp, 1961; Skipp and Mamet, 1970). Orogenic activity subsided in the Antler highland area in latest Chester time, and neritic limestones of the *Jensen Member* of the Chainman Shale and the lower part of the Ely Limestone were deposited in much of eastern Nevada.

In northern Utah, shallow-water impure limestone and shale accumulated throughout interval D time except on the Wendover highland which was probably emergent during much of interval D time. The highland, however, may have been mostly submerged and may have received a thin sediment cover during deposition of the *Willow Gap Limestone* and part of the *Jensen Members* of the Chainman Shale. South of the highland (Utah loc. 114) these members contain more than 200 feet of fossiliferous shallow-water limestone below an uppermost thin quartzite conglomerate (pl. 9-G, sec. *R–R'*; Sadlick, 1965). In the Tintic area, the thick Chiulos Member of the Great Blue Limestone may be a stratigraphic equivalent of orogenic conglomerate in the *Donner Member* of the Chainman Shale in Utah and Nevada and the White Knob Limestone in Idaho. A very late Chester cycle of regression and transgression, which succeeded the carbonates that dominated interval D in the Oquirrh basin and areas east of the Wasatch line, has been reported in the lower part of the Manning Canyon Shale (Moyle, 1959).

Shale and siltstone dominate the orogenic sediments in southeastern California, indicating that the Antler highland in this area, if present, was probably a positive area of low relief. It is possible that post-Mississippian erosion has removed the record of coarser conglomerates to the west in California, but the former interpretation is preferred in this report.

Interval D strata in southern Nevada consist of a thin but complex succession of neritic marine sediments — shale, limestone, sandstone, and minor conglomerate — indicating proximity to a landmass of low relief to the east. Interval D deposition is extended to the east, however, to include the top few feet of the Redwall Limestone which contains a Chester brachiopod fauna in the central Grand Canyon area, Arizona (fig. 76; Dutro, in McKee and Gutschick, 1969). The occurrence of this fauna plus the absence of recognized continental facies in interval D permits the speculation that Chester rocks were deposited in southern California (pl. 10, fig. 4) also but were removed by earliest Pennsylva-

nian uplift and erosion (Gordon, 1969; Wilson, in McKee, Crosby, and others, 1976).

Sadlick (1965, p. 71–72) interpreted the sedimentary framework of intervals C and D in eastern Nevada as being analogous to "complementing miogeosynclines" and exogeosynclines of Kay (1951, p. 14–15, p. 87) because the clastic sediments consist of low-rank graywackes and argillites derived from eugeosynclinal source rocks and transported to a site on the preexisting miogeosyncline.

Basaltic volcanism, mostly marine, is interpreted as having taken place near the northeastern edge of the Antler highland, probably on oceanic, rather than continental, crust (Rogers and others, 1974), in northern Elko County, Nev., near the Idaho-Nevada border. Interval D volcanic rocks have not been identified with certainty west of the highland in Nevada nor in the Taylorsville area in northern California. They are present in the Klamath Mountains of northern California (Yates, chap. Q), however, and may have been present in parts of western Nevada.

PALEOGEOGRAPHIC INTERPRETATIONS

Volcanoes were probably present near the northeastern edge of the Antler highlands in the vicnity of the Idaho-Nevada border, and volcanic islands are inferred to have been present in the seas that presumably covered the Great Basin area west of the highland (pl. 12, figs. 4*A*, *C*). Sediments associated with the volcanics are basinal marine deposits, and at least some lava flows are submarine. It seems possible that all or most of this volcanism was submarine or that explosive volcanism was very minor, and hence pyroclastic material was not available for deposition with the contemporaneous detrital rocks in the Diamond Peak Formation farther south in Elko County.

During a large part of interval D time the northern part of the Antler highland in central Nevada was a mountain range with relief of several thousand feet. The sedimentary record indicates that renewed uplift of the range occurred midway through interval D (Chester) time.

After medial Chester time, the northern part of the highland was steadily eroded until latest Chester time, when the average relief was probably less than 1,000 feet, while fine-grained detrital rocks and limestones of the upper part of the Diamond Peak Formation and the lower part of the Ely Limestone were deposited along the eastern edge.

Large coalescing deltas fed by steep-gradient streams draining in the highland probably occupied the eastern margin of the highland in Nevada during most of interval D time. In southeastern California the southern part of the Antler highland probably was made up of low hills with a total relief of less than 1,000 feet throughout interval D time, as indicated by the consistently fine-grained Chester sedimentary rocks in the Eleana trough.

Warm shallow-marine waters, which probably were muddy most of the time, covered eastern Nevada east of the Antler highland during Chester time. The area of the Ely arch, which was emergent during interval C time, now was under water and receiving significant volumes of fine detrital sediment.

In northern Utah, the Wendover highland was an area of hills of moderate relief from which some of the conglomerate in the *Donner Member* of the Chainman Shale may have been derived during medial Chester time. In latest interval D time, the highland possibly was submerged and may have received a thin cover of fine sediments.

Warm shallow-marine waters covered most of the remaining area in western Utah, and the Oquirrh basin southeast of the Wendover highland began to subside at a rapid rate, which resulted in the accumulation of more than 4,000 feet of shallow-water carbonate rocks.

Warm shallow-marine waters in which silt and mud accumulated also probably covered most of southern Nevada and southeastern California. The abundant mud and silt probably represent winnowed or eroded sediments derived from the Antler highland to the west. Warm marine waters, largely free of detrital material, covered the shallow shelf area along the Wasatch line as well as the low, flat terrain of the craton to the southeast.

TOTAL THICKNESS OF MISSISSIPPIAN ROCKS

THICKNESS TRENDS

The isopach map showing total thickness of the Mississippian System (pl. 7) in the Great Basin region is partly restored, as are each of the four interval isopach maps; isopachs are dashed through areas of unpenetrated younger rocks in order to give them continuity.

Patterns of the total thickness isopachs resemble those of intervals C and D and reveal the minor influence from underlying intervals. In north-central Nevada, thick sequences (5,000 feet or more) of allochthonous volcanic rocks and volcanic-rich sedimentary rocks are present in large klippen on both the western and northeastern edges of the Antler highland (Antler orogenic belt of Roberts and others, 1958). Thicknesses of these displaced rocks are shown only as "plus marks" on the isopach map (pl. 7). In the Chainman–Diamond Peak trough of central Nevada, thicknesses of detrital sedimentary rocks derived from the

highland approach 10,000 feet, and in the related Eleana trough of southern Nevada, thicknesses exceed 7,500 feet. In southeastern California thicknesses closest to the highland only locally exceed 3,000 feet. An isolated faulted sequence of detrital rocks of unknown thickness (Nev. loc. 236) may be present on the projected Antler highland near Cortez, Nev. (Roberts and others, 1967, p. 37). A linear area of thinner Mississippian rocks (less than 2,500 feet thick), east of the troughs, extends from northwestern Utah southwestward into southern Nevada and includes the Wendover highland and the Ely arch, both contemporaneous Mississippian positive structural features.

In northern Utah, more than 7,000 feet of Mississippian rocks is present in two connected northeast-trending elongate depressions which terminate abruptly along an east-west axis south of Gold Hill and the East and West Tintic Mountains. These areas of thick Mississippian strata occupy the general position of the succeeding Pennsylvanian-Permian Oquirrh basin. Foutz (1966, p. 69) has compiled a map which shows a similar configuration for the Mississippian isopachs of this area. Bissell (1962, p. 194) and Armstrong (1968b, p. 30) have isopached similar thickness data as a large triangular-shaped basin with an east-trending southern margin parallel to, but south of, the Devonian Cortez-Uinta arch of Roberts and others (1965). Either isopach configuration seems permissible from the control points used in this report. The configuration of the northeast-trending troughs does not crosscut the trends of other structural features as much as the triangular basin configuration. Thicknesses of Mississippian rocks within the Oquirrh basin range from 3,800 feet to more than 7,000 feet.

An area of missing Mississippian strata is indicated in the vicinity of the northernmost Silver Island Range (Anderson, 1957) and Newfoundland Mountains (Paddock, 1956) in northern Utah, owing to either nondeposition or erosion of Mississippian rocks.

Mississippian isopachs in northern Utah are obviously disrupted (pl. 7) in the vicinity of the Charleston-Nebo thrust zone (Crittenden, 1961) within the Sevier fold and thrust belt. Upper plate thicknesses are between 5,500 and 6,000 feet, whereas subjacent lower plate thicknesses nowhere exceed 3,300 feet. Southwestward along the Sevier orogenic belt in southwestern Utah, isopach offsets are conjectural but seem necessary to explain the distribution of the sparse thickness data available. Utah locality 251 in the southeastern corner of Millard County has a total thickness of only 273 feet within the Sevier thrust belt. This anomalously thin section overrides an autochthonous sequence with dashed isopachs of about 1,000 feet. Farther south in Beaver County, Utah, upper plate thicknesses of 2,500 feet in the Wah Wah Mountains and 1,240 feet in the Star Range override autochthonous lower plate thicknesses of less than 1,000 feet.

Thickness contrasts, isopach offsets, and facies differences occur along numerous other thrust boundaries in Nevada, western Utah, and southeastern California. Only a few of the major thrusts have been listed under the discussion of structural setting. Because of the low density of control points, limited exposure of the faults, and small scale of the map (pl. 7), most of them have not been shown.

Isopachs are bent, broken, or offset in the vicinity of the Nevada-California boundary along zones of complex thrusting, right-lateral strike-slip faults, and oroflexural bends.

An area of anomalously thick Mississippian rocks (2,000 feet or more) is present in southwestern San Bernardino County, Calif., in the vicinity of the poorly known metamorphosed Furnace Limestone. F. G. Poole (oral commun., 1972; Stewart and Poole, 1975) believed that the thickness of Mississippian strata in the Furnace may be significantly less than 2,000 feet. Thinner sequences occur adjacent to this area farther north and east. Mississippian rocks thin to a few hundred feet near the California-Arizona boundary.

Several thousand feet of Mississippian rocks may be present in metamorphosed volcanic and sedimentary rocks of northern California, but stratigraphic and structural complexities preclude isopaching (Yates, chap. Q).

PALEOTECTONIC IMPLICATIONS

The total isopach map of the Mississippian System in the Great Basin region largely reflects the distribution of Upper Mississippian rocks — that is, those assigned to intervals C and D. The thin, dominantly carbonate terranes of intervals A and B were buried beneath imposing detrital aprons of the succeeding intervals.

Crustal instability in the Great Basin region during the Mississippian Period is indicated by a pattern of repeated basin accumulation and highland denudation. The major positive feature — the Antler highland — extended in an arcuate pattern south-southwest from north-central Nevada to west-southwest into central California. This highland received no sediment during the Mississippian. Andesite and interbedded chert and slate are present west of the highlands at Taylorsville in northern California. Basaltic and andesitic volcanic rocks on the Antler highland in north-central Nevada are allochthonous and may have been thrust a great distance from their site of deposition. The volcanics are inferred to have been deposited in an oceanic environment somewhere northwest or west of the highland and moved into their present position in post-Jurassic –pre-

Tertiary time. Basic volcanic rocks along the northeast edge of the highland, however, are interbedded with orogenic sediments that may have been derived from the highland and are believed to record local Mississippian volcanic activity near the east side of the Antler orogenic belt.

East of the Antler highland, two sinuous troughs or foreland basins developed — the Chainman–Diamond Peak trough on the north and the Eleana trough on the south. Rocks in the Chainman–Diamond Peak trough, which are thicker and better known than those in the trough to the south, contain evidence that the trough was an area of intermittent subsidence throughout Mississippian time. Rocks in the Eleana trough are less well dated but also contain some evidence that this area was one of subsidence in all Mississippian intervals. Lower Mississippian shale and siltstone of limited areal distribution occur in both troughs and indicate that the Antler highland was an emerging feature of low to moderate relief during Early Mississippian time. During Late Mississippian time, however, the Antler was an actively rising positive area which contributed more than 10,000 feet of flysch — conglomerate, sandstone, and shale — to the Chainman–Diamond Peak trough, and more than 7,700 feet of sandstone, shale, and conglomerate to the Eleana trough to the south.

Unstable crust extended east of major trough development, where a north-northeast structural lineament recorded by thin Mississippian rocks extends from northwestern Utah southwestward into east-central Nevada. This elongate area of thin strata includes the Wendover highland, which locally is devoid of Mississippian rocks, and the Ely arch, a major positive feature during interval C time. Large volumes of detritus were shed from the Wendover highland and deposited in surrounding areas, but the Ely arch seems to have been either a low positive area or a submerged area of thin sediment accumulation during most of Mississippian time.

East of the Wendover highland, Mississippian strata are relatively thick (about 7,000 feet) and consist largely of rocks assigned to interval D. The area in northwestern Utah, east of the Wendover highland, remained relatively stable until latest Mississippian time when it began to subside gradually. Subsidence accelerated during Pennsylvanian and Permian time, and as much as 25,000 feet of the Oquirrh Formation accumulated in the Oquirrh basin (Roberts and others, 1965, p. 1937).

Relatively thin Mississippian strata in southwestern Utah, southeastern Nevada, and southeastern California indicate slow carbonate deposition on a stable shelf. Recent work in northern Mexico and Baja California indicates that thin Mississippian carbonate rocks probably occur near the international border, even though they are metamorphosed in part (López-Ramos, 1969, p. 2409), and such an occurrence implies that a stable shelf extended over a large part of southernmost California.

GEOLOGIC UNITS DIRECTLY ABOVE MISSISSIPPIAN SYSTEM

UNITS OVERLYING MISSISSIPPIAN SYSTEM

In most of the Great Basin region, rocks of Pennsylvanian interval A (Morrow) overlie Mississippian strata (pl. 8). These Lower Pennsylvanian rocks include: in northwestern Utah, the upper part of the Manning Canyon Shale, Oquirrh Formation, Ely Limestone, and the Diamond Peak of the northern Silver Island Range (Schaeffer and Anderson, 1959, p. 1786); in northeastern Nevada, the Ely Limestone and the Moleen Formation of the Ely Group; in southwestern Utah, the Ely Limestone, Oquirrh Formation, Topache Limestone, and Callville Limestone; in southern Nevada, the Ely Limestone, Callville Limestone, Tippipah Limestone, and Bird Spring Formation; and in southern California, the Bird Spring Formation and upper part of the Furnace Limestone. Locally, the basal beds of these formations include rocks of latest Mississippian age.

Middle Pennsylvanian rocks of interval B (Atoka) — including the Ely Limestone in northeastern Nevada and the Keeler Canyon Formation, Tihvipah Limestone, and Bird Spring Formation in southeastern California — overlie Mississippian rocks around the depositional margins of Pennsylvanian interval A (pl. 8). Probably the Keeler Canyon and Tihvipah overlie Lower Pennsylvanian beds of the Lee Flat Limestone and Rest Spring Shale here assigned to Mississippian interval D. The Battle Formation of Middle Pennsylvanian age (interval B — Atoka), locally overlies the Goughs Canyon Formation with a fault contact in southeastern Humboldt County at the west edge of the Antler highland. Elsewhere, the Goughs Canyon is overlain by Quaternary alluvium or Tertiary volcanic rocks and locally is thrust over rocks as young as Jurassic (Willden, 1964, p. 18). In Pershing County the Inskip Formation is in thrust contact with the allochthonous Triassic Koipato Formation (Roberts and others, 1958, p. 2849; Silberling and Roberts, 1962, p. 15), but elsewhere it is overlain by Quaternary sediments (Ferguson and others, 1951).

Upper Pennsylvanian (intervals D and E) rocks of the Strathearn Formation unconformably overlie Mississippian and Lower Pennsylvanian rocks of the Diamond Peak Formation which are assigned to Mississippian interval D in northeastern Nevada, in

Elko County, in parts of the Pinon Range, Adobe Range, Pequop Mountains, and Windermere Hills. Farther south, Upper Pennsylvanian (intervals D and E) rocks of the Brock Canyon Formation unconformably overlie Mississippian rocks in southern Eureka County and adjacent north-central Nye County near the Antler highland.

Undivided Permian rocks, largely carbonate conglomerate, overlie a thin Mississippian limestone unit in the northern Silver Island Range in northwestern Utah (Anderson, 1960). The Permian *Pequop Formation* as used by Thorman (1962, p. 72) also unconformably overlies Mississippian Diamond Peak Formation or the *Donner Member* of the Chainman locally in the northern Pequop Mountains of eastern Elko County, Nev. Permian rocks of the Carbon Ridge Formation and associated strata unconformably overlie Mississippian rocks in the western part of the Eureka district (Nolan, 1962), in the southwesternmost Pinon Range (J. F. Smith, Jr., written commun., 1968), and in the northern Adobe Range, Nev. (Ketner, 1970a). Elsewhere in the Adobe Range and Pequop Mountains, rocks of Pennsylvanian and Permian age generally are in fault contact with the Diamond Peak Formation.

Tertiary volcanic rocks unconformably overlie the *Schoonover Formation* in northwestern Elko County, Nev.

Widespread Tertiary and Quaternary volcanic and volcanic-rich sedimentary rocks unconformably overlie Mississippian rocks in many places in eastern Nevada along the eastern edge of the Antler highland and in many areas east of the highland in west-central Utah in the Dugway Range, northern Sheeprock Mountains, Tintic Junction area, and Gilson Mountains.

The top of the *Meiklejohn Formation* in southern Nevada has not been described because it is either faulted or covered by Cenozoic deposits (Cornwall and Kleinhampl, 1961, 1964).

The Peale Formation at Taylorsville in northern California is overlain by the *Goodhue Formation* of McMath (1966, p. 176) of Permian age.

PALEOTECTONIC IMPLICATIONS

Lower Pennsylvanian rocks gradationally overlie Upper Mississippian interval D rocks over much of the central part of the area of known Mississippian deposition in the Great Basin. Deposition seems to have continued uninterrupted across the Mississippian-Pennsylvanian systemic boundary in a large part of the Great Basin. The stratigraphic record of the Mississippian System may be more complete in the Cordilleran geosynclinal area than it is in the platform area of the midcontinent region, where a depostional break occurs at the systemic boundary.

The record of Mississippian deposition is incomplete locally on the continental shelf or miogeosynclinal area where crustal instability has been documented at or near the close of the Mississippian. In the area of the Wendover highland, Pennsylvanian(?) erosion is believed to have beveled locally the quartzose clastic rocks of the undifferentiated Chainman–Diamond Peak sequence (Anderson, 1957, 1960; Schaeffer, 1960; Schaeffer and Anderson, 1959). This sequence, though generally fine grained, contains subangular quartzite-pebble conglomerate beds near the top that may indicate renewed uplift and erosion of a small part of the Wendover highland in latest Chester time (Sadlick, 1965, p. 150; Anderson, 1957). Pre-Ely erosion also has been suggested to explain the thin Chainman Shale sequence (350 feet) in the Ferguson Mountains (Nev. loc. 2), but possibly complex faulting has thinned the sequence (Berge, 1960).

Intermittent post-Mississippian uplift and erosion in the vicinity of the Antler highland is indicated by the overlap of Mississippian or older rocks by Upper Pennsylvanian and Permian rocks from the Adobe Range south to an area west of Eureka in central Nevada. Further evidence for late Paleozoic erosion of Mississippian rocks in the vicinity of the Antler highland has been documented at Jett Canyon in the Toiyabe Range (Nev. loc. 155) in south-central Nevada where cobbles of limestone containing Early and possibly Late Mississippian fossils occur in conglomerate beds within the Permian Diablo Formation above beds containing indigenous Permian fossils (Helen Duncan, written commun., 1959).

In the central Monitor Range south of Clear Creek Canyon, north-central Nye County, Nev., conglomerates that unconformably overlie Devonian and Ordovician rocks (Nev. loc. 329) are tentatively assigned to the Mississippian Eleana Formation, but they may instead belong to the Permian Diablo Formation, according to J. I. Ziony (oral commun., 1967).

The unconformity that suggested Lower Pennsylvanian interval B (Atoka) rocks resting on Upper Mississippian rocks in northern Inyo County in southeastern California (pl. 8) may not be real if the underlying formations (Rest Spring Shale and Lee Flat Limestone) assigned to Mississippian interval D contain Lower Pennsylvanian rocks. This area was near the edge of the carbonate shelf and may have been relatively stable during the transition from latest Mississippian into earliest Pennsylvanian time.

A disconformity representing early and middle Morrow time has been described by Gordon and Poole (1968) at the top of the Mississippian Eleana Formation and Chainman Shale equivalent, or *Indian Springs Formation* (type locality), for a large part of southern Nevada and adjacent southeastern California. This

period of Early Pennsylvanian epeirogenic uplift and erosion is reflected in the sparsity of interval D (Chester) strata over a large part of southern California and northern Arizona where later Pennsylvanian rocks disconformably overlap interval C (Meramec) strata.

Crustal instability and attendant uplift and erosion, at the end of the Mississippian, were most prevalent along the western margin of the craton in southern Nevada and southeastern California and on the shelf in the Wendover highland area in northwestern Utah. Areas of trough and shelf sedimentation in Nevada and southeastern California were relatively constant, and, though parts of the Mississippian Antler highland were inundated later in Early and Middle Pennsylvanian time, no significant changes have been recorded at the systemic boundary in this area.

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Pacific Northwest Region

By ROBERT G. YATES

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

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1010–Q

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

PACIFIC NORTHWEST REGION

By ROBERT G. YATES

ABSTRACT

The Pacific Northwest region, which consists of all of Washington and Oregon and the Klamath Mountains of California, contains not more than four widely separated areas of rocks identifiably of Mississippian age. The close association and similarity of these rocks to rocks of Devonian and Pennsylvanian ages suggest that conditions that existed before and after the Mississippian continued through this period without change or regional hiatuses. All rocks, except those at one locality in northeastern Washington, were deposited in the Cordilleran eugeosyncline and, accordingly, reflect the tectonic instability of that environment. The great lack of data prohibits graphical representation either of intervals or of total thicknesses and lithofacies; nevertheless, the available samples — poor as they are — are an adequate basis for interpreting the general character of the region during Mississippian time. The close spatial association of volcanic rocks, conglomerates, cherts, and coral-bearing limestones indicates and is compatible with an environment where water depth, local provenance, and sediment transport patterns were subject to sudden and extreme change.

INTRODUCTION

The Pacific Northwest region consists of all of Washington and Oregon and the Klamath Mountains of northwestern California. More than 99 percent of this region is covered by post-Permian rocks through which no drill hole has penetrated into rocks that are identifiably of Mississippian age. Rocks of Mississippian age — or possible Mississippian age — crop out in three widely separated areas (fig. 78; pl. 1): (1) the Klamath Mountains of northwestern California, (2) the Suplee area of east-central Oregon, and (3) northern Washington which includes two subareas: (3a) the San Juan Islands and northwestern Cascade Range in northwestern Washington, and (3b) northeastern Washington.

All the Paleozoic rocks in the Pacific Northwest region, except some in northeastern Washington, were deposited in the Cordilleran eugeosyncline and, accordingly, are subject to the severe limitations that this tectonic province inflicts on the precision of age determinations, lithologic correlations, boundary relations, thickness values, provenances, and projections under covered areas. Because of these limitations and the smallness of the available samples, it is impossible to pigeonhole meaningful stratigraphic units into the four time intervals and describe them with isopach and lithofacies maps. Doubtless, all Mississippian time is fragmentally represented throughout the region, but the rocks exposed at any one place lack fossils necessary to demonstrate or divide off any definitive part of the spectrum.

PALEOGEOLOGY

UNITS UNDERLYING MISSISSIPPIAN

In California, Oregon, and Washington, Devonian rocks occur either stratigraphically below or in close geographic association with Mississippian rocks (pl. 2). At no place in the region is there positive evidence of a major disturbance between the Devonian and Mississippian, although in Oregon and Washington structural complexities and meager faunal control make it difficult to establish uninterrupted deposition. In California the Middle Devonian Kennett Formation of the Klamath Mountains (pl. 15, col. 129) is separated by an unconformity from the overlying Bragdon Formation (Albers and Robertson, 1961, p. 16). The Bragdon contains Mississippian fossils in its upper part; but these are not diagnostic of any particular Mississippian interval, and it is conceivable that the lower part of the Bragdon is Late Devonian in age and that the unconformity falls within Devonian time. At the one Mississippian locality in Oregon, the Suplee area, Middle Devonian limestone (Kleweno and Jeffords, 1962) crops out within 2 miles of the fossiliferous Mississippian rocks, but exposures are too poor and structural relations are too uncertain to establish the nature of the boundary between the rocks of the two ages. In northwestern

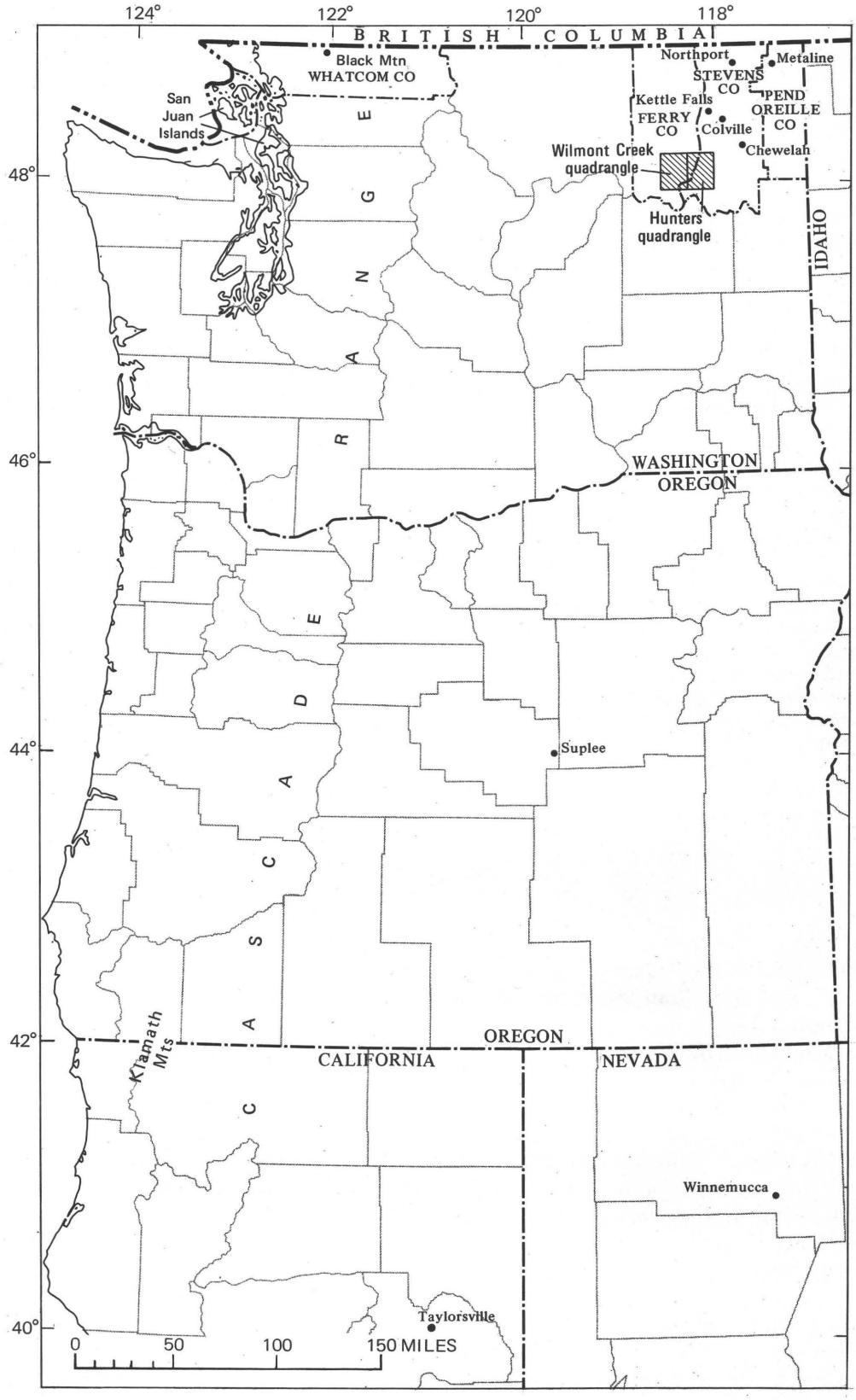

FIGURE 78. — Pacific Northwest region showing localities and areas mentioned in the text.

Washington, Mississippian rocks may or may not be present; the probability of this is discussed in a following section. If present, they are represented by at least some rocks Danner (1966, p. 65–67) considered Pennsylvanian in age, rocks which occur in association with Middle and Upper Devonian limestones. The stratigraphic problems are compounded not only by numerous faults but even more so by an unsystematic assemblage of graywacke, argillite, chert, and greenstone containing isolated lenses and pods of limestone that range in age from Devonian to Permian (Danner, 1966, p. 62–74; Misch, 1966, p. 116).

This assemblage is called the *Chilliwack Group* (Moen, 1962). Conceivably, limestone lenses and pods were forming more or less continuously at one place or another during *Chilliwack* time and now occur like plums scattered in a pudding. The age of the exposed limestone "plums" depends upon where erosion has "sliced the pudding."

In the Chewelah area of northeastern Washington, Mississippian limestone and dolomite are in fault contact with Cambrian and Precambrian rocks. Middle Devonian and possible Upper Devonian rocks occur in the area. An unfossiliferous unnamed argillite sequence, traceable for 50 miles southwestward from northernmost Pend Oreille and Stevens Counties, and the Covada Group (not shown on pl. 15), a eugeosynclinal assemblage that lies farther to the southwest in southern Stevens and Ferry Counties, probably embrace at least some of Mississippian time. The unfossiliferous argillite is underlain by Devonian rocks, and possibly the same relationship exists within the Covada Group.

INTERVALS A–D

The paucity of fossil control in the Pacific Northwest region prohibits the definition of interval boundaries. In Washington and Oregon there are no sections with two or more fossil horizons; in California the section has one reliable, closely dated Mississippian fossil horizon and one other horizon that defines little more than the period. In most places in California and Oregon the Mississippian age of the rocks is decided by the presence of the gigantoproductid, *Titanaria costellata* Muir-Wood and Cooper, and in Washington by the presence of a similar gigantoproductid (Danner, 1966, p. 66–67). This genus, as well as most corals and brachiopods found in West Coast eugeosynclinal faunas, is not found in the miogeosynclinal and shelf facies to the east, but all these fossils have their equivalents or close relatives in Asia or Europe.

FORMATIONS INCLUDED

Mississippian time is faunally represented in California by the Bragdon and Baird Formations, in Oregon by the Coffee Creek Formation, and in the Chewelah area of northeastern Washington by unnamed limestone and dolomite (pl. 15, cols. 129–131). Mississippian time is possibly represented in northeastern Washington in the Metaline area by the black argillites of Russian Creek — which are considered by Dings and Whitebread (1965, p. 26) to be Ordovician rocks — in the Northport area by the *Grass Mountain sequence*, and in southern Ferry and adjacent Stevens County by the Covada Group. The Mississippian is questionably represented in northwestern Washington by the *Red Mountain sequence* (Danner, 1966, p. 74) of the *Chilliwack Group*. The Covada Group, *Grass Mountain sequence*, argillites of Russian Creek, and supposedly related argillites will hereafter be referred to as the argillite sequence.

The Baird and Coffee Creek Formations have enough fossil elements in common to be considered — at least in part — time equivalents; the *Red Mountain sequence* has affinities with both Coffee Creek and Baird through the mutual presence of gigantoproductids. The Bragdon Formation, underlying the Baird, presumably was deposited in the Early Mississippian, representing all or only part of Kinderhook and Osage time. Although the fossils that Diller (1906, p. 3) found in the Bragdon Formation confirm the Mississippian age, they do not define the series represented. Diller's fossil locality has never been recovered for additional collecting.

The *Red Mountain sequence*, which was considered, with reservations, by Danner (1966) to be Early Pennsylvanian in age, is included here in the Mississippian because of the controversial nature of this assignment. In Danner's words (1966, p. 66): "The dating of the limestone is somewhat doubtful, as they contain brachiopods strongly resembling those of the Visean (Mississippian) of Asia and Europe, a few Foraminifera and a stromatoporoid(?) that seem to be characteristic of the Early Pennsylvanian in many parts of the world, and a coral fauna of distinctive Asiatic type, which appears to have strong affinities to the Permian." Assuming that the fossils are correctly identified, the problem becomes either a case of reworked fossils in clastic limestones, or a widely accepted but fundamental misconception of the time range of Asian genera found this side of the Pacific Ocean. Danner (1966, p. 67) interpreted the faunal evidence as favoring the intermediate age, Pennsylvanian, probably because the Permian corals can be most easily equated to it. However, if

the corals are disregarded, the evidence strongly favors a Mississippian assignment. As long as the age of this fossil assemblage is unresolved it is better to tentatively consider these limestones and closely associated rocks as possibly Mississippian in age and include them here, as well as in the Pennsylvanian folio (W. P. Irwin in McKee, Crosby, and others, 1975, p. 329–332).

Rocks of questionable Mississippian age also are present in northeastern Washington. In this area the only rocks that contain Mississippian fossils are limestones and dolomites that occur in the Chewelah area. Fifty miles to the north in the Metaline area, black siliceous argillites that contain pods and lenses of conglomerates containing clasts of quartzite and chert overlie fossiliferous Middle Devonian rocks. These rocks can be recognized in northern Stevens County near Northport, where they are included in the *Grass Mountain sequence* (Yates, 1964) and are considered to be Carboniferous in age. Possible correlatives of this sequence can be traced at least 25 miles to the southwest to the vicinity of Kettle Falls, where they may be represented by a thick conglomerate containing chert clasts. Still farther south, in the Hunters and Wilmont Creek quadrangles (Campbell and Raup, 1964; Becraft, 1964), a eugeosynclinal assemblage, the Covada Group, could by inference lie above probable Devonian rocks. Quite possibly, Mississippian time is represented both in the lower part of this group and in the *Grass Mountain sequence*, although the assignment is tenuous.

UPPER BOUNDARY

At no place in the Pacific Northwest region are the relations between Mississippian and younger Paleozoic rocks clear and unequivocal. In northwestern Washington, at Black Mountain in Whatcom County, a limestone reef of Leonard (Permian) age overlies a conglomerate that in turn overlies a fossiliferous limestone of the *Red Mountain sequence*. The conglomerate, which contains plant fragments, represents a change from marine to nonmarine conditions and, although not directly datable, is probably Pennsylvanian in age regardless of whether the *Red Mountain sequence* is Mississippian or Early Pennsylvanian. In northeastern Washington the Mississippian in the Chewelah area is overlain by Pleistocene deposits. It is possible that Pennsylvanian as well as Mississippian rocks are present in both the *Grass Mountain sequence* and the Covada Group.

In Oregon the Coffee Creek Formation is overlain by a basal conglomerate of the Spotted Ridge Formation which contains clasts of limestone from the Coffee Creek. The Spotted Ridge Formation is dated as Pennsylvanian (Mamay and Read, 1956) from poorly preserved plant fossils found 1,200 feet above its base and 400 feet below the overlying Coyote Butte Limestone of late Wolfcamp age (Ogren, 1958, p. 23). Accordingly, the unconformity at the top of the Coffee Creek may represent erosion during Mississippian time or erosion during Early Pennsylvanian time. In the Klamath Mountains of California, the Upper Mississippian formation, the Baird, contains Visean brachiopods and corals in limestone lenses in the middle mudstone unit 400 feet below the upper greenstone unit, which is separated in most places from the overlying McCloud Limestone of Permian age by a sill of mafic quartz diorite (Albers and Robertson, 1961, p. 20–22). Farther north, J. W. Harbaugh reported (in Skinner and Wilde, 1965, p. 11) that silty shales strongly resembling parts of the Baird Formation contain Morrow fusulinids, which suggest that Mississippian deposition continued without interruption into the Pennsylvanian.

THICKNESS TRENDS

Meaningful thickness measurements of Mississippian rocks are available only from the Shasta area of northwestern California, and even within this small area, total thickness ranges from 3,000 to 8,000 feet (pl. 7). The Coffee Creek Formation of central Oregon consists of about 900 feet of limestone, but, because it is only a partial section, this thickness is almost meaningless as a measure of the total thickness of Mississippian rocks present. The same is true of the several hundred feet of unnamed limestone and dolomite in the Chewelah district of Washington. If some or all of the questionable Mississippian argillites (the *Grass Mountain sequence* and possible equivalents) in northeastern Washington are truly Mississippian, the thickness of Mississippian rocks there may be anything from tens to thousands of feet. In northwestern Washington, the structural and stratigraphic complexity permits speculation that the thickness of Mississippian rocks may range from hundreds to thousands of feet.

LITHOFACIES TRENDS

The Mississippian rocks of the Pacific Northwest region, except those in the Chewelah area in northeastern Washington, belong to the eugeosynclinal facies. Because the eugeosynclinal assemblage is a potpourri of lithologies, which individually have no regional significance, any regional lithofacies trend is that of the assemblage, or in other words, that of the eugeosyncline, whose eastern limit is the only linear element that can be even approximately fixed (pl. 10). The sum of lithologies from all areas of exposure probably represents a qualitative sample of rock types that are present, but the sample is not representative of relative abundance; for example, in Oregon, carbonate

rock is the only lithologic type that is exposed, and in California, conglomerate is of major abundance. Both rock types are of relative unimportance in the total accumulation in the eugeosyncline. Because of the difficulty of determining representative lithofacies symbols for the four intervals of the Mississippian System in the Pacific Northwest, the "lithofacies undetermined" symbol has been used at those localities on the interval lithofacies maps (pls. 3-B, 4-B, 5-B, 6-B).

In northwestern California, a great variety of rocks is represented, including such diverse types as bioclastic limestone and mafic lava flows. They occur in two formations, the Bragdon and the Baird. The Bragdon Formation, the older unit, was described by Albers and Robertson (1961, p. 16) as "principally shale, mudstone, and conglomerate, but includes a minor amount of siltstone, sandstone (graywacke), tuff, and volcanic breccia." The conglomerate consists largely of chert fragments but also includes sparse fragments of greenstone and limestone, some of which contain Devonian fossils. The Baird Formation, according to Albers and Robertson (1961, p. 18), consists "chiefly of pyroclastic rocks, interbedded with tuffaceous sandstone, mudstone, and mafic flows in the lower part; siliceous mudstone, and minor amounts of limestone, chert, and tuff in the middle part; and greenstone in the upper part." The pyroclastic rocks include andesite tuff, keratophyre tuff, and quartz keratophyre tuff; the mudstone, which is interlayered with the pyroclastic rocks, consists chiefly of microcrystalline quartz.

The sole Mississippian formation in Oregon, the Coffee Creek Formation, has a fauna correlative to that of the Baird Formation but is lithologically quite different. It is a pure to argillaceous or sandy limestone in its upper part and predominantly calcareous sandstone in its lower part; its base is not exposed. According to Merriam and Berthiaume (1943, p. 149–151): "Patchy distribution of the pure limestone outcrops in the upper part of the formation suggests that they represent more or less extended lenses within arenaceous and argillaceous limestone facies."

In Washington the fossiliferous Mississippian rocks of the Chewelah area are a carbonate facies. The inferred but questionable Mississippian rocks of the Northport area, along with the equally questionable Covada Group rocks that occur farther to the south, are transitional and eugeosynclinal facies, and the questionable Mississippian of northwestern Washington is a eugeosynclinal unit. According to Fred K. Miller (written commun., 1967), the unnamed Mississippian rocks of the Chewelah area are fine- to coarse-grained medium-gray fetid limestone and dolomite, which are separated by about 100 feet of maroon slate from underlying nonfossiliferous massive dolomite that is locally cherty. This rock unit belongs to the miogeosynclinal facies, or possibly to the shelf facies. If not autochthonous, it was thrust in from the east. The *Grass Mountain sequence* of the Northport area and the unnamed probable equivalent that extends southward to the Colville–Kettle Falls area are predominantly a dark-gray to black siliceous argillite that contains a few beds of fine-grained limestone and rare pods of pebble conglomerate. The Covada Group is predominantly fine-grained graywacke and gray to black slate but includes greenstone and some chert. The best that can be said of its age is that it is probably post-Devonian and pre-Cretaceous. Its thickness of tens of thousands of feet could include Mississippian rocks, but whether the greenstone would occur in this part of the section is, of course, unknown. In northwestern Washington the limestones that contain the fossils of questionable Mississippian age are part of an assemblage of graywacke, dark argillite, chert, and mafic volcanic rocks.

GEOLOGIC UNITS ABOVE THE MISSISSIPPIAN SYSTEM

The units overlying the Missisippian System are shown on plate 8. In California the Mississippian Baird Formation apparently extends into Pennsylvanian (Morrow) time without lithologic change or hiatus (pl. 15, col. 129). In Oregon the Coffee Creek Formation is overlain by the basal conglomerate of the Spotted Ridge Formation, which contains poorly preserved Pennsylvanian plant fossils. In northwestern Washington the *Red Mountain sequence* is overlain by a conglomerate containing plant fragments, which may be inferred to be of Pennsylvanian age because the conglomerate is in turn overlain by a limestone containing Permian fossils. In northeastern Washington the carbonate rocks of the Chewelah area are overlain by Pleistocene glacial deposits, and the questionable Mississippian age of the argillite sequence is probably overlain by more argillite of Pennsylvanian age.

PALEOTECTONIC TRENDS AND IMPLICATIONS

Almost all Paleozoic rocks of the Pacific Northwest region belong to the eugeosynclinal facies and almost all fossils found in these rocks are in biohermal and reeflike limestones and probably indicate shallow-water conditions. Because limestone is a very minor component of the facies, we can infer that the environment that could both support life and preserve its fossil forms must have been limited. The close association of these shallow-water limestones with such diverse rocks as bedded chert, graywacke, greenstone, dark argillite, and conglomerate indicate frequent and extreme shifts

in water depth, provenance, or tectonic stability. Only from the aggregate sample of all the Paleozoic rocks in the eugeosyncline can one interpret the conditions that existed during the Mississippian. This larger sample has influced all interpretations here presented. Because of the impossibility of distinguishing the Mississippian intervals in the Pacific Northwest region, the same generalized interpretations have been repeated on each of the interval paleotectonic (pl. 10, figs. 1–4) and paleogeographic (pl. 12, figs. 1–4) maps.

Data are inadequate to define directly the trend of the eugeosyncline, but its course is inferred to parallel the course of the miogeosyncline, which extends roughly N. 15° E. from southern California to southeast Idaho (pl. 10). The eastern limit of the eugeosyncline falls in Nevada; the restored boundary drawn by Roberts, Hotz, Gilluly, and Ferguson (1958, fig. 4) lies 15 miles west of Winnemucca. In Nevada, Early Mississippian time was occupied by the Antler orogeny (Late Devonian–Early Mississippian) which culminated in the Roberts Mountains thrust fault, a decollement that moved eugeosynclinal rocks of the early Paleozoic eastward more than 100 miles (Roberts and others, 1958). Assuming that the upper plate did not separate into detached blocks, one can make a palinspastic restoration and can return the rocks to the hypothetical highland from which they slid. This reconstruction places the topographic axis of this positive area within 30 miles of Taylorsville, Calif. where marine rocks were accumulating during the Early Mississippian (McMath, 1966). A more conservative estimate is that of Gilluly and Gates (1965, p. 122) who spoke of the displacement on the Roberts Mountains thrust, as "more than 55 miles." This more modest — although minimal — estimate of displacement permits the axis of the hypothetical Antler highland to be considerably east of Taylorsville and more symmetric in position. The highland probably extended northward into Idaho, possibly into the area that is now occupied by the Idaho batholith.

There is no compelling evidence, or even suggestive evidence, that the Antler highland extends into Washington. Such a projection would place a positive area in north-central Washington, west of both limestone and black shale sequences, and east of the eugeosynclinal facies that existed in western Washington. The part of this area in north-central Washington that is not covered by Tertiary volcanic rocks is occupied by a small amount of Permian and Triassic eugeosynclinal deposits, the Colville batholith, and a high-grade metamorphic complex. The metamorphic complex is a southern extension of the Shuswap Complex of southern British Columbia and includes the only rocks that possibly could have formed a positive area during the Mississippian. The age of the Shuswap protoliths and the age of the metamorphism were most recently summarized by Wheeler (1966, p. 37) as follows. [Documentations omitted from quotation.]

> The rocks of the Shuswap Complex have been variously considered to be of Archean age * * *, to lie unconformably beneath the Permian * * *, and to include Mesozoic strata * * *. Recent work * * *, however, has not confirmed the unconformities except where the rocks involved may not belong either to the Shuswap or to the Permian and furthermore evidence to follow indicates that the earliest recognizable deformation took place in the Mesozoic. It is uncertain whether the granite gneiss in the cores of the domes represented old basement. The overlying metasediments may, however, range in age from late Precambrian to late Paleozoic and locally may even include the Mesozoic.
>
> The earliest recognized structures in the complex are warps and isoclinal, recumbent structures developed along east-west axes. Similar oriented structures may occur in low-grade metamorphic rocks of Triassic and Lower Jurrassic age just east of the complex at 50°10′ N, 117°45′ W and suggest therefore that the early structures in the Shuswap developed in post-Early Jurassic time.

The present state of knowledge, therefore favors a protolith of rocks ranging in age from Precambrian to Mesozoic, whose earliest structures and metamorphism are late Jurassic or younger. Accordingly, a hypothetical Mississippian positive area in north-central Washington would be an unmetamorphosed terrane, possibly capable of contributing sediments, such as the graywacke phase of the Covada Group, but certainly not capable of contributing the argillites of the *Grass Mountain sequence*. The point to be made is not that positive areas or islands did not exist in north-central Washington but that evidence of a major landmass and associated orogenic belt in this area is lacking.

The Mississippian Period, however, was not necessarily without tectonic events in Washington State. The suggestion of diastrophism introduced by the sudden appearance of coarse clastics of possible Mississippian age in an otherwise monotonous black lutite sequence suggests the appearance of a new provenance without disruption of the old. An epeirogenic emergence of lower Paleozoic sediments deposited to the east could supply clasts of both quartzite and chert to the conglomerate, as well as add to the great bulk of lutite that probably was furnished by deep longshore currents coming down from the north. Such a positive area might have been located in the panhandle of Idaho.

The "possible" Mississippian rocks of northwestern Washington, the *Red Mountain sequence*, are limestones, which represent bioclastic accumulations in shallow waters. However, associated with the limestones are unfossiliferous argillite, graywackes, cherts, and mafic lavas, which by associations are also "possible" Mississippian. These probably were deposited in the most westerly part of the eugeosyncline near the western limits of continent-derived sediments. Deposition was in a marine environment in-

terrupted by tectonic and volcanic islands. Detrital materials could be derived from rocks formed earlier in the eugeosyncline and brought up from depths as fault-block islands. By such a process of cannibalsim augmented by the contribution of volcanic debris from local sources and the introduction of a lutite fraction from the continent — perhaps distributed by deep longshore currents — the clastic rocks could have accumulated. The limestones may have been fringing reefs marginal to the tectonic and volcanic islands.

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