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RECENT ADVANCES IN MIDDLE CARBONIFEROUS BIOSTRATIGRAPHY — A SYMPOSIUM

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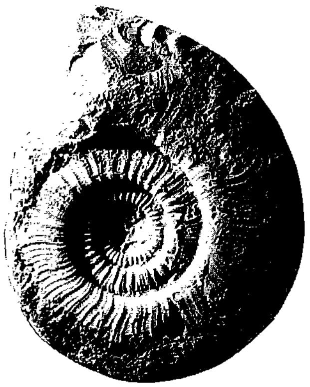

Proceedings of a symposium held March 5, 1990, during the 24th annual meeting of the Geological Society of America, South-Central Section, at Stillwater, Oklahoma, sponsored by the South-Central Section of the Paleontological Society and Oklahoma State University.

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Title-Page Illustration

Axinolobus quinni McCaleb and Furnish, 1964; lateral view of specimen (×1.3) collected from Gene Autry Shale Member, Golf Course Formation exposed along Sycamore Creek, Johnston County, Oklahoma (see Pl. 1, Fig. 7 of paper by Manger, Miller, and Mapes and their discussion).

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PREFACE

A symposium entitled "Recent Advances in Middle Carboniferous Biostratigraphy" was held at Stillwater, Oklahoma, on March 5, 1990, as part of the annual meeting of the South-Central Section, Geological Society of America. We organized this symposium under the aegis of the South-Central Section of the Paleontological Society. Fourteen invited papers were presented at the symposium by speakers representing institutions and organizations in both the United States and Canada. The symposium included a discussion session, open to all participants, on advances in Middle Carboniferous biostratigraphy to determine progress on the various series boundaries, divisions, and their correlations. This volume contains 12 revised contributions arising from the symposium as documentation of the current status of Middle Carboniferous biostratigraphy.

The past decade has seen formalization of a bipartite division of the Carboniferous into subsystems using the Madrid (1983) mid-Carboniferous boundary definition. Considerable progress has been made toward more precise and consistent recognition of the Morrowan–Atokan boundary since publication in 1984 of our previous symposium proceedings (Sutherland and Manger, Oklahoma Geological Survey Bulletin 136, now out of print). Attention has also begun to focus on recognition of the Atokan–Desmoinesian boundary and its correlation. The present volume provides contributions on a spectrum of topics related to refinement of Middle Carboniferous biostratigraphy. We are gratified by the continued interest in the Middle Carboniferous shown by our colleagues through their participation in our field trip held prior to this meeting, and also their support of the activities of the Subcommission on Carboniferous Stratigraphy (SCCS).

This symposium was an outgrowth of our work on the Atokan Series over the past several years. We acknowledge the continued support of the National Science Foundation through grants to Sutherland (EAR 8517591) and Manger (EAR 8520684). We thank Charles J. Mankin, Director of the Oklahoma Geological Survey, for his continued interest and encouragement of our research efforts on Middle Carboniferous stratigraphy in Oklahoma and adjacent areas. The School of Geology and Geophysics, University of Oklahoma, provided facilities for completion of the project. We are particularly indebted to Marjorie Starr, School of Geology and Geophysics, for her typing and editorial assistance and to Clifford Cuffey, School of Geology and Geophysics, for his help with figure drafting and modification.

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Middle and Upper Chesterian Brachiopod Biostratigraphy, Eastern Appalachians, Virginia and West Virginia

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ABSTRACT.—The Pennington Group and equivalent middle and upper parts of the Mauch Chunk Group in the eastern Appalachian basin range in age from Late Mississippian (middle Chesterian) to Early Pennsylvanian and consist of strata deposited in coastal-plain to shallow-marine environments. Both marine invertebrate and plant fossils are preserved. The Little Stone Gap Member near the middle of the Hinton Formation has a diverse fauna that is the basis for the *Orthotetes subglobosus–Martinia contracta* Brachiopod Assemblage Zone. This member correlates with the Fayetteville Shale in the Ozarks region and with the sequence from about the top of the Glen Dean Limestone to about the base of the Menard Limestone in the Chesterian "type" area. The upper shale member at the top of the Hinton Formation is characterized by a brachiopod assemblage dominated by *Productus*? n. sp. A and *Inflatia* n. sp. A. The upper shale member correlates with the lower to middle parts of the Pitkin Limestone of the Ozarks and probably with the Menard Limestone of the Illinois basin. The Pride Shale Member at the base of the Bluestone Formation has a restricted marine, mollusk-dominated fauna, and a sparse brachiopod fauna characterized by *Leiorhynchoidea carbonifera* and *Crurithyris fayettevillensis*. This association has more ecologic significance than biostratigraphic importance, although poorly preserved Pride Shale goniatites are upper Chesterian forms. Near the top of the Bluestone Formation, the widespread Bramwell Member contains a moderately diverse brachiopod fauna: the *Orthotetes* n. sp. B-*Diaphragmus* n. sp. A Brachiopod Assemblage Zone. These brachiopods support correlation of the Bramwell with the Imo Formation of northern Arkansas, and, possibly, the upper part of the Pitkin Limestone. Conodonts from the Bramwell, characteristic of the *Adetognathus unicornis* Zone, also support correlation with the lower and middle parts of the Imo Formation and with the Grove Church Shale of the Illinois basin, but not knowingly with the uppermost part of the Imo Formation, which is apparently unfossiliferous.

INTRODUCTION

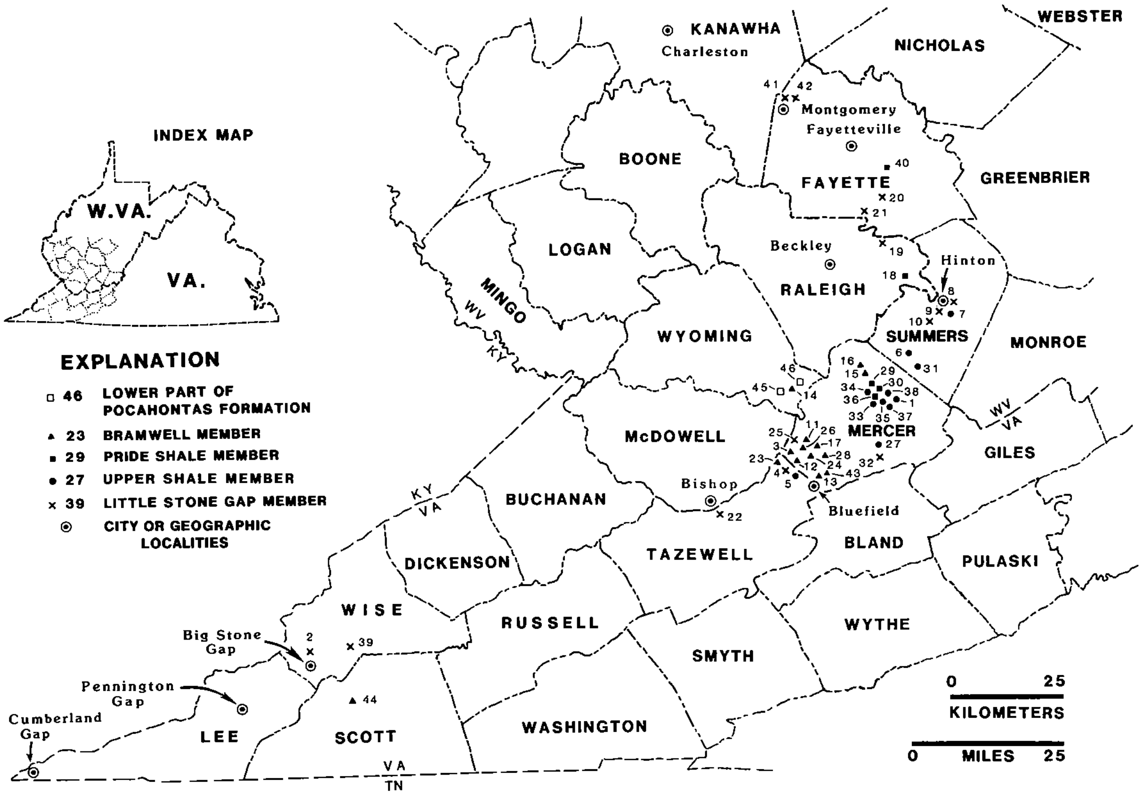
Scope of Investigation

This paper summarizes the biostratigraphic framework developed from studies of brachiopod macrofaunas from middle and upper Chesterian marine bands in the Pennington Group of southwestern Virginia and the coeval middle and upper parts of the Mauch Chunk Group of southern West Virginia. Pertinent biostratigraphic information from other fossil groups is included, where available. A much more comprehensive study of these same brachiopod faunas is nearing completion. Because many of the brachiopod species from these strata have not been described, open nomenclature is freely employed here for these taxa.

Approximately 5,300 brachiopods were examined from 82 collections of eastern Appalachian invertebrate fossils that contain brachiopods. Each of the 46 locality numbers in Text-figure 1 represents one or more collections of invertebrate fossils taken from a given bed or member at a discrete geographic site. Corresponding USGS "PC" (Permo–Carboniferous) locality register numbers and abbreviated stratigraphic and geographic information are given in the Appendix.

Study Area and Geologic Setting

The study area (Text-fig. 1) is located primarily in the outcrop belt of the Pennington Group and equivalent parts of the Mauch Chunk Group along the eastern structural margin of the Appalachian basin in southwestern Virginia and southern West Virginia, respectively. The study area also extends northward in gently northwest-dipping strata from Hinton, West Virginia, down the New River Gorge to the vicinity of Fayetteville, West Virginia, where the Pennington extends into the subsurface. Two core samples were obtained from coal-test boreholes in extreme western Fayette County (locs. 41 and 42 of Text-fig. 1), and one other core sample



Text-figure 1. Locality map showing brachiopod collection sites in southwestern Virginia and southern West Virginia. Numbers correspond to sets of USGS collection localities, presented in the Appendix.

was examined from northern Scott County, Virginia (loc. 44 of Text-fig. 1).

During the Late Mississippian and Early Pennsylvanian, the axis of part of the Appalachian foreland basin extended northeasterly through roughly Washington, Smith, Bland, and Giles Counties, Virginia (Text-fig. 1). We consider that the trough area of this sedimentary basin also included Tazewell, Mercer, and Summers Counties. This trough closely paralleled a large landmass (Appalachian positive element) that lay somewhat farther to the southeast and that contributed large volumes of terrigenous clastics to the foreland basin. Approximately 4,250 ft of Upper Mississippian rocks and ~1,700 ft of Lower Pennsylvanian rocks occur in the trough area, where sedimentation was virtually continuous across the Mississippian–Pennsylvanian boundary.

Acknowledgments

We gratefully appreciate the efforts of K. J. Englund, C. B. Cecil, G. D. Stricker, C. R. Meissner, Jr., C. E. Mason, W. H. Gillespie, and H. W. Pfefferkorn for assisting us with various aspects of this study, particularly with field work and collecting marine fossils. K. J. Englund, C. B. Cecil, J. T. Dutro, Jr., and P. K. Sutherland reviewed early drafts of this manuscript.

LITHOSTRATIGRAPHY

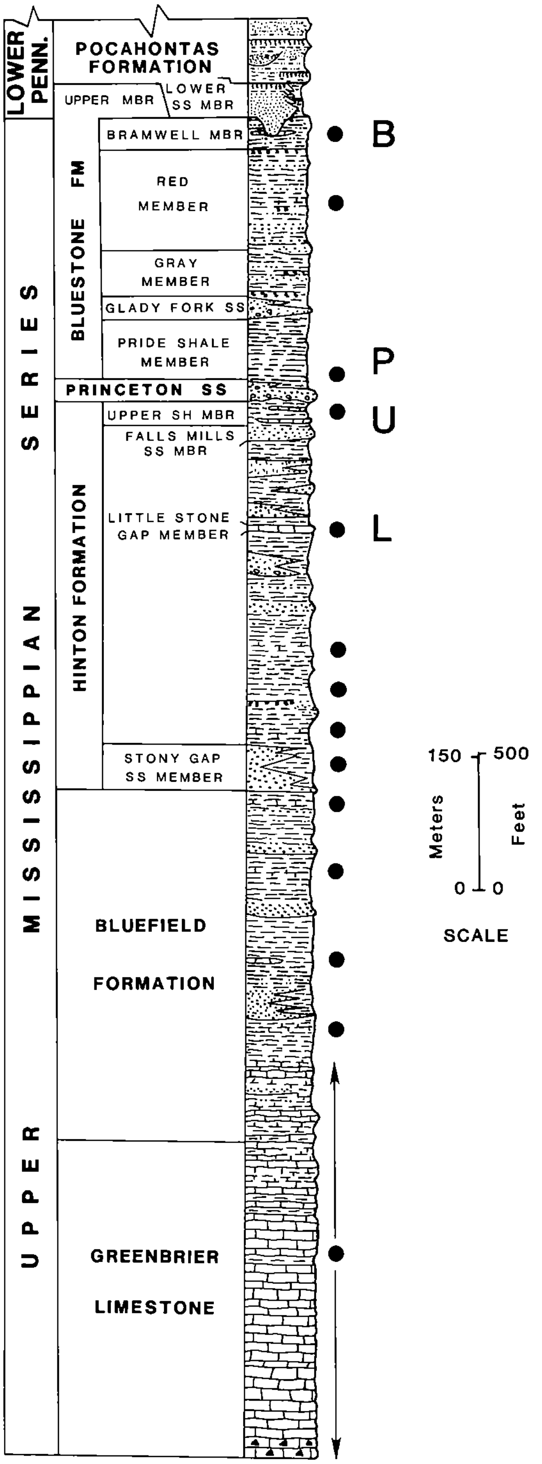
Extensive discussions of the lithostratigraphy and depositional environments of the Upper Mississippian sequence in the eastern Appalachians are presented by Englund (1979) and Arkle and others (1979). The nomenclature outlined herein primarily follows those authors.

The Upper Mississippian succession (Text-fig. 2) in the study area extends from the base of the Greenbrier Limestone (Meramecian and lower Chesterian), upward through the Bluefield Formation (lower and middle? Chesterian), and through most of the Pennington Group. In southwestern Virginia, the name Pennington Group is applied to the sequence encompassing the Hinton Formation, the Princeton Sandstone, and the Bluestone Formation. In adjacent parts of southern West Virginia, the name Mauch Chunk Group is used for the strata from the base of the Bluefield Formation through the Bluestone Formation. In the trough area, the boundary between the Mississippian and Pennsylvanian Systems is placed within the upper part of the Bluestone Formation. The Bluestone is generally conformable with the overlying coal-bearing Pocahontas Formation which, in turn, is succeeded by the New River Formation (both Lower Pennsylvanian). To the northwest away from the trough, an unconformity near the base of the New River Formation progressively truncated the lowermost part of the New River Formation, the Pocahontas Formation, and some of the underlying Upper Mississippian sequence. In Wise County, the Lee Formation (the lateral equivalent of the New River Formation) rests directly on strata of the middle to upper parts of the Bluestone Formation.

Middle and Upper Chesterian Stratigraghy and Marine Beds

The name Pennington was proposed as a formation by Campbell (1893) for the succession at Pennington Gap, Virginia (Text-fig. 1), and the exposures at Big Stone Gap were designated the type section by Wilpolt and Marden (1959). The Pennington was raised to group rank in Virginia by Harris and Miller (1958) to include the Hinton Formation, Princeton Sandstone, and Bluestone Formation, each defined from outcrops in the trough area in Tazewell, Mercer, and Summers Counties. In its type section, the Pennington is unconformably overlain by the Lee Formation. There, upper Bluestone Formation equivalents were thought to be missing beneath this unconformity until Englund and others (1983b,1989) demonstrated their presence in both the Pennington Gap and Big Stone Gap sections and elsewhere in southwestern Virginia. A more serious problem in application of Pennington as a group name involves the Princeton Sandstone. That unit has been miscorrelated to the southwest of its type area in Mercer County and is known to pinch out long before reaching the Pennington type area (see Englund, 1979, p. 11). In extreme southwestern Virginia, it is difficult to subdivide the Pennington into formations based on the same criteria used in the trough area. Nevertheless, we follow the criteria of deWitt and McGrew (1979), Englund and Randall (1981), and Englund and Henry (1984) in referring this set of strata to the Pennington Group throughout southwestern Virginia.

The name Mauch Chunk was extended from the southern anthracite field in eastern Pennsylvania (Arkle and others, 1979) but is less easily applicable as a group name than Pennington in our study area. Usage of the Mauch Chunk Group persists for the



Text-figure 2. Stratigraphic column for Upper Mississippian and part of Lower Pennsylvanian section in trough area of Appalachian basin in southwestern Virginia and southern West Virginia. Figure modified from Englund (1979). Dots to right of stratigraphic column depict positions from which marine or marginal marine invertebrate fossils were collected. Marine bands discussed in paper indicated by letters B (Bramwell Member of Bluestone Formation), P (Pride Shale Member of Bluestone Formation), U (upper shale member of Hinton Formation), and L (Little Stone Gap Member of Hinton Formation).

Bluefield through Bluestone section in southern West Virginia (Arkle and others, 1979).

The strata of the Pennington Group and equivalent parts of the Mauch Chunk Group are ~2,200 ft thick in the trough area near Bluefield, Virginia–West Virginia. This sequence consists of alternating nonmarine coastal-plain and shallow marine deposits characterized by variegated mudrock, fine- to coarse-grained sandstone, quartzose conglomeratic sandstone, fossiliferous argillaceous limestone, gray shale, thin carbonaceous shale, and coal beds underlain by rooted underclay. Its predominant grayish-red and greenish-gray color is in sharp contrast to the dominantly gray colors of the underlying Bluefield and overlying Pocahontas Formations. Marine invertebrate fossils occur in at least 10 stratigraphic positions within the Pennington Group and equivalent middle and upper Mauch Chunk strata (marked by solid dots on Text-fig. 2), but most of these are marginal-marine to brackish-water molluscan and/or ostracode associations that do not contain brachiopods. Only the four major marine units that bear brachiopods are considered here; these are also marked by capital letters on Text-figure 2.

Hinton Formation

The Hinton Formation conformably overlies the Bluefield Formation throughout the study area. Near Bluefield, the Hinton Formation is ~1,320 ft thick and contains two marine beds. The Little Stone Gap Member of the Hinton Formation, in about the middle of the formation, is the thickest and most persistent of the Pennington marine units in the eastern Appalachians. It extends from extreme eastern Tennessee (Englund and others, 1989) northward along the Appalachian front to east-central West Virginia and is present in boreholes in western Fayette County. It attains a maximum thickness of ~100 ft in the Bluefield area. Throughout most of the study area, this member is a medium-light-gray, highly fossiliferous, nodular limestone but becomes almost a marl in its southwestern exposures.

The upper shale member of Englund (1968) forms the upper subdivision of the Hinton Formation. This member is dominated by dark-gray shale and mudstone and attains its maximum thickness of ~145 ft near Bluefield. At its base, a carbonaceous shale is reported to be widely traceable in the subsurface with geophysical logs (Schalla, 1984). However, Englund (personal communication, 1990) asserts that Schalla's contention is overstated and that his "hot shale marker" is neither continuous nor easily traced. The upper shale member also contains several thin, argillaceous, fossiliferous limestone beds or lentils. The dark-gray calcareous shale and mudstone between and associated with these impure limestones are highly fossiliferous in many places.

Princeton Sandstone

The Princeton Sandstone is a conglomeratic unit that is ~60 ft thick in its type area in Mercer County. It has been widely miscorrelated throughout much of the study area (see Englund, 1979, fig. 3 and p. 11 for discussion). According to Englund (1979), it does not extend much farther southwest than west-central Tazewell County. It also pinches out in the New River Gorge in northern Raleigh County (Englund and others, 1977). The Princeton does not appear to contain indigenous marine faunas. Instead, marine fossils, where present, consist entirely of abraded "free" specimens and those in limestone pebbles and cobbles in the lower part of the formation clearly reworked from the underlying upper shale member of the Hinton Formation.

Bluestone Formation

The Bluestone Formation attains a maximum thickness of ~850 ft along the Bluestone River in its type area near Bluefield. This formation contains two relatively widespread marine units, the distinctive Pride Shale Member at its base and the Bramwell Member near its top.

The Pride Shale Member consists principally of regularly laminated, very dark-gray shale and silty shale that grades laterally into siltstone. The lower part of the member commonly contains a very thin coal bed and several rooted underclay horizons. Sparse marine fossils occur in a very thin, dark-gray calcareous mudstone and in dolomitic nodules in this part of the Pride section. The bulk of the member does not contain invertebrate body fossils. The Pride Shale Member is ~250 ft thick in Mercer County, and, although thinner elsewhere, it can be recognized throughout most of the study area, even where the Princeton Sandstone has pinched out.

The other marine bed is the Bramwell Member, the uppermost unit of Mississippian age in the Bluestone Formation. In its type section near Bluefield, the Bramwell attains a maximum thickness of 120 ft. It consists of a thin carbonaceous shale at the base. This is overlain by medium-gray shale that grades upward into poorly laminated siltstone and intensely bioturbated mudstone (forming most of the member) that, in turn, coarsens upward into fine-grained sandstone at the top. Included within this member are fine-grained, calcareous sandstone bodies that exhibit weak bidirectional flaser bedding and that contain brachiopods, pelecypods, and other marine invertebrates. Depositionally, the basal carbonaceous shale of the Bramwell Member represents tidal marsh sediments. They were succeeded by a tidal-flat sequence, and the coarser grained calcareous sandstone bodies within the Bramwell appear to be have been deposited in tidal creeks or guts.

The upper member of the Bluestone Formation is primarily a sequence of grayish-red and greenish-gray nonmarine mudrock. The lower sandstone member of the Pocahontas Formation consists of a series of distributary sandstone bodies that cut through the upper member of the Bluestone into the Bramwell Member in many places in the vicinity of Bluefield (Englund, 1974; Englund and others, 1981). Laterally, away from the axes of the distributary channels, the upper member of the Bluestone Formation interfingers with the lower sandstone member of the Pocahontas Formation.

BRACHIOPOD ASSEMBLAGES AND BIOSTRATIGRAPHY

A moderately diverse brachiopod fauna consisting of 50 species is present in the marine bands of the Hinton and Bluestone Formations in the eastern Appalachians. Eleven new and undescribed species belong to the following genera: *Lingula, Oehlertella*?, *Orthotetes* (two new species), *Inflatia, Productus* (two species—one definite assignment, one questionable), *Diaphragmus, Echinoconchus, Ovatia*, and *Fluctuaria*. Most of these new species are known only from the eastern Appalachians, although several occur in the Chesterian section of the Ozarks. Of the remaining taxa, 27 are assigned to named Upper Mississippian species described and/or identified from the upper Mississippi valley or from the Ozarks region of the North American Midcontinent, and the remaining two are identified merely to genus because a lack of sufficient specimens and/or inadequate preservation. Typical species of each of the four marine bands discussed in the preceding section are illustrated in Plates 1–4.

The stratigraphic occurrences of the brachiopod taxa making up the eastern Appalachian faunas are plotted in the range chart (Table 1). The majority of the fifty species have restricted stratigraphic ranges and do not occur in all of the marine bands studied. Twenty-nine species are restricted to a single member in this interval, and an additional 14 occur only in two members. Only four species are identified or tentatively identified in all four marine bands. Thus, with the exception of the ecologically restricted fauna of the Pride Shale Member, the individual marine bands can be distinguished faunally from one another. The long-ranging species in the Pennington Group are *Orbiculoidea* sp. A and the widely reported Chesterian spiriferids, *Reticulariina spinosa* (Norwood and Pratten), *Eumetria costata* (Hall), and *Composita subquadrata* (Hall).

Brachiopod Faunas and Biostratigraphy of the Hinton Formation

A diverse fauna of 39 brachiopod species occurs in the two major marine bands of the Hinton Formation. Eleven of these taxa (depicted by asterisks in Table 1) are known to range upward from the subjacent Bluefield Formation, and an additional three are suspected to do so. Three more species—*Rugosochonetes chesterensis* (Weller), *Punctospirifer transversus* (McChesney), and *Girtyella indianensis* (Girty)—appear to extend upward from the upper and middle parts of the Greenbrier Limestone. Seven species definitely range upward into the Bluestone Formation, and two additional species may do so.

Orthotetes subglobosus–Martinia contracta Assemblage Zone

The Little Stone Gap Member contains the most consistently diverse brachiopod fauna of the Pennington Group. Twenty-nine species are identified from this member (Table 1). *Adectorhynchus suspectum* (Girty), *Productus* n. sp. A, *Inflatia cherokeensis* (Drake), *Echinoconchus* n. sp. A, *Nucleospira* n. sp., *Eumetria* cf. *E. acuticosta* S. Weller, and *Hustedia* n. sp. A are stratigraphically restricted to this member. Of these, only *Productus* n. sp. A is especially common. *Petrocrania chesterensis* (Miller and Gurley), *Composita acinus* Girty, *Anthracospirifer* sp. A, and *Skelidorygma ozarkensis* (Snider), which occur downsection, have their highest known occurrences in the eastern Appalachian faunas in the Little Stone Gap Member.

The most characteristic brachiopods of the Little Stone Gap Member, in addition to *Productus* n. sp. A, are the longer-ranging forms *Orthotetes subglobosus* Girty, *Orthotetes* n. sp. A, *Diaphragmus cestriensis* (Worthen), *Ovatia elongata* Muir-Wood and Cooper, *Coledium cestriensis* (Snider), *Composita subquadrata* (Hall), *Martinia contracta* (Meek and Worthen), and *Eumetria costata* (Hall). The two species of *Orthotetes, Productus* n. sp. A, *D. cestriensis*, and *Ov. elongata* dominate brachiopod assemblages in the limestone facies that characterizes this member throughout most of the study area. In contrast, *M. contracta* is especially common in the marly facies in Wise County, Virginia, where, in the area around Big Stone Gap, Virginia, the overall invertebrate macrofauna attained its maximum diversity.

The assemblage discussed above and the additional species listed in Table 1 from the Little Stone Gap Member constitute the *Orthotetes subglobosus–Martinia contracta* Brachiopod Assemblage Zone. The more common brachiopods from this assemblage zone are illustrated in Plate 1.

Productus? n. sp. B–*Inflatia* n. sp. A Assemblage Zone

Intensive collecting produced a diverse fauna of 27 brachiopod taxa (Table 1) from the upper shale member of the Hinton Formation. Characteristic forms are illustrated in Plate 2. Seven species of brachiopods are restricted to this member: *Inflatia* n. sp. A, *Productus*? n. sp. B, *Pustula* sp., a small indeterminate buxtonid, *Fluctuaria* n. sp. A, *Pugnoides*? sp., and *Dieslasma*? sp. Of these, *Productus*? n. sp. B and *Inflatia* n. sp. A are abundant in much of the outcrop area studied. Many of these productid specimens are in growth position with their long

TABLE 1.—RANGE CHART FOR BRACHIOPOD TAXA FROM EASTERN APPALACHIANS

	Brachiopod taxon	Member			
?	*Petrocrania chesterensis* (Miller & Gurley)	L	-	-	-
	Adectorhynchus suspectum (Girty)	L	-	-	-
	Productus n. sp. A	L	-	-	-
	Inflatia cherokeensis (Drake)	L	-	-	-
	Echinoconchus n. sp. A	L	-	-	-
*	*Composita acinus* Girty	L	-	-	-
	Nucleospira n. sp.	L	-	-	-
*	*Anthracospirifer* sp. A	L	-	-	-
*	*Skelidorygma ozarkensis* (Snider)	L	-	-	-
	Eumetria cf. *E. acuticosta* S. Weller	L	-	-	-
	Hustedia n. sp. A	L	-	-	-
	Orthotetes subglobosus Girty	L	U	-	-
	Orthotetes n. sp. A	L	U	-	-
*	*Rugosochonetes chesterensis* (S. Weller)?	L	U	-	-
	Heterolosia sp.	L	U	-	-
*	*Punctospirifer transversus* (McChesney)	L	U	-	-
	Diaphragmus cestriensis (Worthen)	L	U	-	-
	Flexaria cf. *F. arkansana* (Girty)	L	U	-	-
?	*Ovatia elongata* Muir-Wood & Cooper	L	U	-	-
?	*Coledium cestriense* (Snider)	L	U	-	-
	Cleiothyridina atrypoides Girty	L	U	-	-
*	*Martinia contracta* (Meek & Worthen)	L	U	-	-
*	*Torynifer setiger* (Hall)	L	U	P	-
	Marginovatia minor (Snider)	L	-	P?	-
?	*Girtyella indianensis* (Girty)	L	U	P?	-
	Orbiculoidea sp. A	L	U	P	B
*	*Reticulariina spinosa* (Norwood & Pratten)	L	U	P	B
	Eumetria costata (Hall)	L	U	P?	B
*	*Composita subquadrata* (Hall)	L	U	P?	B
	Inflatia n. sp. A	-	U	-	-
	Productus? n. sp. B	-	U	-	-
	Pustula sp.	-	U	-	-
	Buxtonid, gen. and sp. indet.	-	U	-	-
	Fluctuaria n. sp. A	-	U	-	-
	Pugnoides? sp.	-	U	-	-
	Dielasma? sp.	-	U	-	-
	Buxtonia semicircularis (Sutton & Wagner)	-	U	P	-
	Anthracospirifer leidyi (Norwood & Pratten)	-	U	P?	B
	Oehlertella? n. sp. A	-	U	-	B?
	Leiorhynchoidea carbonifera (Girty)	-	-	P	-
	Crurithyris fayettevillensis (Girty)	-	-	P	-
	Torynifer? sp.	-	-	P	-
	Hustedia sp. B	-	-	P	-
	Lingula n. sp. A	-	-	-	B
	Lingula sp. B	-	-	-	B
	Schizophoria sp.	-	-	-	B
	Orthotetes n. sp. B aff. *O. kaskaskiensis* (McChesney)	-	-	-	B
	Diaphragmus n. sp. B aff. *D. cestriensis* (Worthen)	-	-	-	B
	Ovatia n. sp. A. aff. *O. elongata* Muir-Wood & Cooper	-	-	-	B
	Composita aff. *C. deflecta* Mather	-	-	-	B

NOTE: Taxa arranged in order of stratigraphic first appearance. Abbreviations for members as follows: Hinton Formation— L = Little Stone Gap Member, U = upper shale member. Bluestone Formation—P = Pride Shale Member, B = Bramwell Member. Species marked with asterisk (*) on left side of table extend upsection from the underlying Bluefield Formation; those marked with query (?) may occur in the Bluefield. Query (?) in Member column indicates questionable identification.

body spines extending several centimeters outward into the sediment in which they were living. Other common species are *Orthotetes* n. sp. A, *Diaphragmus cestriensis* (Worthen), *Ovatia elongata* Muir-Wood and Cooper, *Coledium cestriense* (Snider), the long-ranging *Composita subquadrata* (Hall), *Martinia contracta* (Meek and Worthen), and *Eumetria costata* (Hall). *Anthracospirifer leidyi* (Norwood and Pratten), which becomes common in higher Chesterian strata, first appears in the upper shale member.

This brachiopod assemblage, plus the additional species listed in Table 1, are referred to as the *Productus*? n. sp. B–*Inflatia* n. sp. A Zone.

Faunas of the Bluestone Formation

Until recently, marine faunas were almost unreported from the Bluestone Formation (Henry and Gordon, 1979; Gordon and Henry, 1981), and latest Chesterian brachiopod faunas are poorly known. The Pride Shale Member has a depauperate and restricted brachiopod fauna. The Bramwell Member has yielded a respectable brachiopod assemblage. Both faunas are mollusk-dominated associations. Eighteen brachiopod species are definitely recorded from the Bluestone Formation, and an additional three taxa are questionably identified.

Leiorhynchoidea carbonifera–Crurithyris fayettevillensis Association

The Pride Shale Member has produced by far the sparsest and least diverse brachiopod fauna in the Pennington Group (Table 1; Pl. 3, Figs. 1–11). The bulk of this member characteristically contains no body fossils. Most of the brachiopod fauna was collected from a thin calcareous shale bed and from scattered carbonate (dolomite) nodules near the base of the member from a limited geographic area in Mercer County, West Virginia (locs. 30 and 36 of Text-fig. 1). Mollusks, including 19 genera of pelecypods and three goniatite taxa, are present in the thin shale bed over a slightly wider area. The ammonoids include *Anthracoceras, Stenoglaphyrites*?, and *Alloceras*, the preservation of which precludes specific identification, but all are late Chesterian genera.

Thirteen brachiopod taxa are present in our collections from the Pride Shale Member (Table 1); however, four of them are questionably identified because of the sparseness of collections and poor preservation. Of the 13 species, only four are restricted to the Pride Shale Member. These are: *Leiorhynchoidea carbonifera* (Girty), *Crurithyris fayettevillensis* Girty, *Torynifer*? sp. (a large form that clearly is not conspecific with *T. setiger*), and *Hustedia* sp. B (a very narrow, elongate species). The remaining nine species, including *Anthracospirifer leidyi* (Norwood and Pratten), do not appear to have closely restricted ranges. We refer to this assemblage as the *Leiorhynchoidea carbonifera–Crurithyris fayettevillensis* association.

The two species described by Girty probably have more ecologic significance than biostratigraphic utility. The type specimens of *Crurithyris fayettevillensis* are from Chesterian strata from the Ozarks. *Leiorhynchoidea carbonifera* ranges downsection in the Midcontinent into undisputed Meramecian strata, but it occurs almost exclusively in dark-colored shales, mudrocks, and limestones. Crurithyrids are generally found in similar lithic types deposited in what are interpreted as stressed ecologic conditions and in "deeper water" settings. However, because of the paleogeographic position of the Pride and the proximity of thin, rooted underclays just above and below the fossiliferous beds of the Pride, we do not favor a "deeper water" depositional interpretation for the Pride Shale Member.

Orthotetes n. sp. B–*Diaphragmus* n. sp. A Assemblage Zone

Like the Pride Shale Member, the Bramwell Member is dominated by mollusks. Thirteen species of brachiopods are identified from this member; characteristic taxa are illustrated in Plate 3 (Figs. 12–21) and Plate 4. Five of these brachiopods are new species; these are *Lingula* n. sp. A, *Oehlertella*? n. sp. A, the large and abundant *Orthotetes* sp. B aff. *O. kaskaskiensis* (McChesney), the common *Diaphragmus* n. sp. A aff. *D. cestriensis* (Worthen) (a coarsely costellate form), and *Ovatia* n. sp. A aff. *Ov. elongata* Muir-Wood and Cooper. In addition, the species referred to as *Composita* aff. *C. deflecta* Mather probably represents a hitherto unrecognized species that may be a precursor to Mather's Morrowan form. In addition to the new brachiopods, the Bramwell has yielded the long-ranging *Composita subquadrata* (Hall), abundant *Anthracospirifer leidyi* (Norwood and Pratten), *Reticulariina spinosa* (Norwood and Pratten), and *Eumetria costata* (Hall). Of the Bramwell species, the brachiopods are dominated by *Orthotetes* n. sp. B aff. *O. kaskaskiensis, Ovatia* n. sp. A aff. *Ov. elongata*, and *A. leidyi*.

The small, narrow *Lingula* n. sp. A (Pl. 4, Figs. 1–5) occurs by the thousands in a widespread, thin carbonaceous shale that marks the basalmost bed of the Bramwell Member. This taxon differs from the *Lingula* that has been collected from the lower part of the Pocahontas Formation at localities 45 and 46 of Text-figure 2. Also contained in the basal carbonaceous shale of the Bramwell is the "fresh-water" ostracode *Darwinula* and the "nonmarine" myalinids *Anthraconaia* and *Curvirimula*(?).

From a very large Bramwell sample, Repetski and Henry (1983) identified the conodonts *Adetognathus unicornis* (Rexroad and Burton), *Cavusgnathus convexus* Rexroad, *C. naviculus* (Hinde), *Gnathodus bilineatus* (Roundy) morphotype delta, *Hibbardella* cf. *H. milleri* Rexroad (*sensu lato*), and other ramiforms assignable to several form taxa. These are typical forms of the *Adetognathus unicornis* Conodont Zone.

CORRELATION WITH "TYPE" CHESTERIAN AND OZARKS SEQUENCES

Our eastern Appalachian brachiopod faunas have reasonably restricted stratigraphic ranges, and some are similar at the species level to those of the North American Midcontinent. Therefore, they are useful in correlating the Appalachian marine bands with the sections in the Ozarks region and in the Eastern Interior basin. Our correlations are illustrated on Text-figure 3.

The strongest correlation using the brachiopod faunas involves the *Orthotetes subglobosus–Martinia contracta* Assemblage Zone of the Little Stone Gap Member of the Hinton Formation. These strata correlate with the middle and upper parts of the Fayetteville Shale in the Ozarks region and with the sequence from about the top of the Glen Dean Limestone to the base of the Menard Limestone in the "type" Chesterian in the Illinois basin.

The *Productus*? n. sp. A–*Inflatia* n. sp. A Brachiopod Assemblage Zone, diagnostic of the upper shale member of the Bluestone Formation, correlates with the lower to middle parts of the Pitkin Limestone of the Ozarks and, probably, with the Menard Limestone of the Illinois basin.

The Pride Shale Member has a restricted marine fauna characterized by *Leiorhynchoidea carbonifera* and *Crurithyris fayettevillensis*. This association has more ecologic significance than biostratigraphic importance, although sparse, poorly preserved goniatites support a late Chesterian age. These strata probably correlate with the upper part of the Pitkin Limestone, although the evidence is not strong.

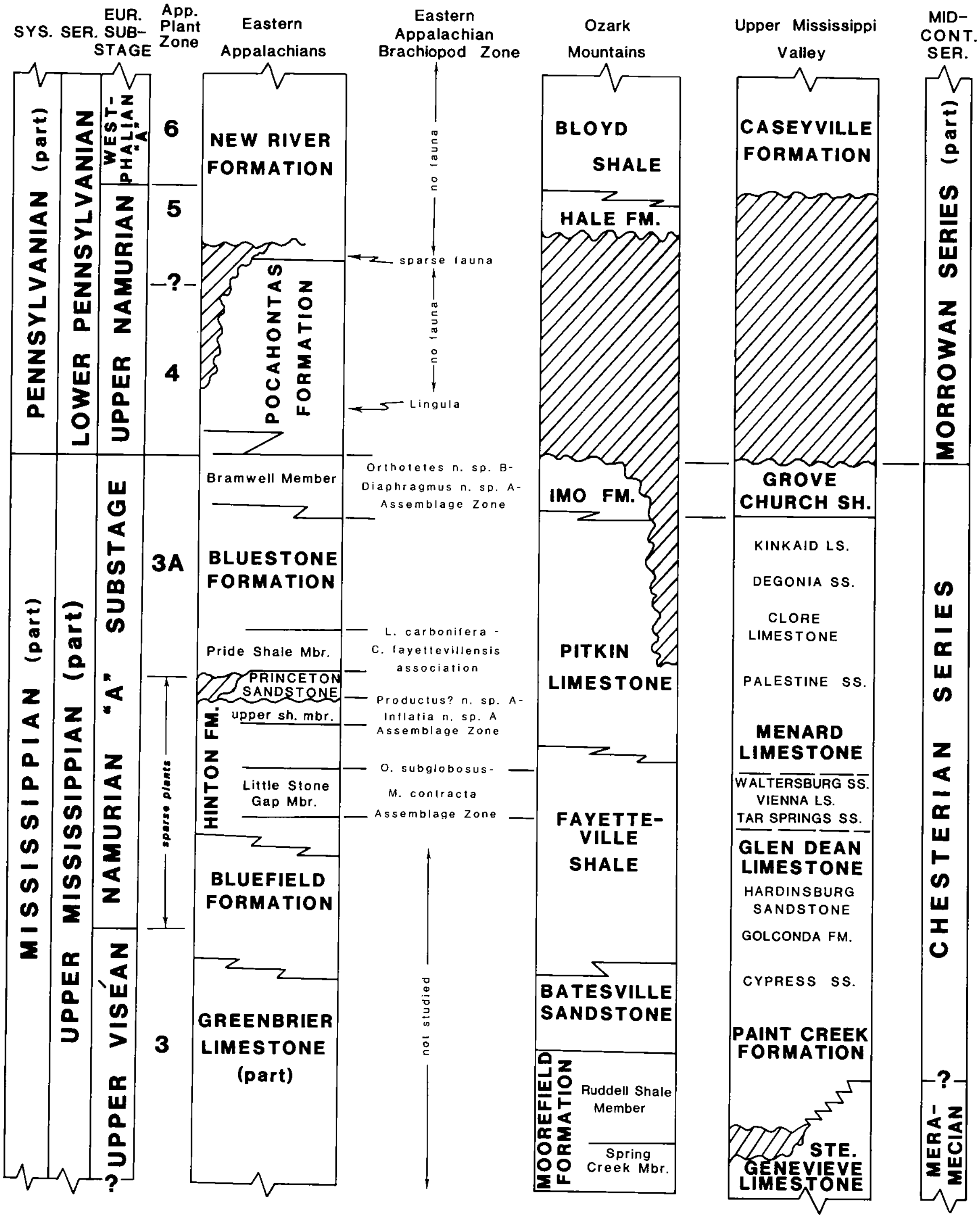
The Bramwell Member near the top of the Bluestone Formation is characterized by the *Orthotetes* n. sp. B–*Diaphragmus* n. sp. A Brachiopod Assemblage Zone and the *Adetognathus unicornis* Conodont Zone. The brachiopods permit correlation of the Bramwell with the Imo Formation of northern Arkansas and, possibly, with the upper part of the Pitkin Limestone in northern Arkansas as well. However, the more diagnostic conodonts support correlation of the Bramwell with the Grove Church Shale at the top of the Chesterian Series in the Illinois basin and with the lower and middle parts of the Imo Formation. The upper part of the Imo is not known to be fossiliferous.

MISSISSIPPIAN–PENNSYLVANIAN BOUNDARY BIOSTRATIGRAPHY

In the trough area of the eastern Appalachians, a Late Mississippian compression and impression macroflora characterized by *Sphenopteris elegans* (Brongnairt) Sternberg, *Stigmaria stellata* Goeppert, and *Sphenophyllum tenerrimum* Ettingshausen (Floral Zone 3A of Pfefferkorn and Gillespie, 1981) ranges from the base of the Bluestone Formation through the red member of the Bluestone. The Bramwell Member, which overlies the red member, has not yielded a diagnostic macroflora, although it contains the Late Mississippian marine invertebrate assemblage discussed above. Floral Zone 4 of Read and Mamay (1964), distinguished by *Neuropteris pocahontas* D. White and *Mariopteris eremopteroides* D. White, first appears in the upper member of the Bluestone Formation and in the basal part of the contemporaneously deposited lower sandstone member of the Pocahontas Formation. Only a sparse marginal-marine *Lingula* association has been collected from the Pocahontas Formation (locs. 45 and 46 of Text-fig. 1), and no conodonts or other diagnostic marine fossils have been recovered from Pocahontas strata in spite of extensive and intensive searching.

The U.S. Geological Survey has recommended establishing the stratotype for the Mississippian–Pennsylvanian boundary at the conformable contact at the base of the lower sandstone member of the Pocahontas Formation near Bluefield, Virginia–West Virginia (loc. 13 of Text-fig. 1) (Englund and others, 1983a,1985,1986; Gillespie and Englund, 1989). This horizon coincides with the lowest known occurrence of Floral Zone 4, and, based on the macroflora, also coincides closely with the Namurian "A" and the Namurian "B" boundary of western Europe.

In contrast, the working definition of the "Mid-Carboniferous" or Mississippian–Pennsylvanian boundary, as recommended by the I.U.G.S. Subcommission on Carboniferous Stratigraphy (Lane and Manger, 1985), is the first occurrence of *Declinognathodus noduliferus* (Ellison and Graves). In the western United States, the *Rhachistognathus muricatus* Conodont Zone succeeds the *Adetognathus unicornis* Conodont Zone, precedes the *Declinognathodus noduliferus* Zone, and is taken to represent the uppermost Mississippian biostratigraphic subdivision (Lane and others, 1985). However, the *Rhachistognathus muricatus* Conodont Zone is differentiated from the underlying *A. unicornis* Zone only by the presence of *Rh. muricatus* (Dunn) together with *A. unicornis* (Rexroad and Burton). Moreover, *Rhachistognathus muricatus* has never been reported from either the eastern or southern Appalachians, from the Illinois basin, or from the southern Midcontinent (Manger and Sutherland, 1984; Manger and Sutherland, this volume). Given the following four points, it is possible that the "Mid-Carboniferous" boundary might indeed occur somewhere within the lower part of the Pocahontas Formation rather than its base. These points are: (1) the absence of *Rh. muricatus* in the Bramwell Member; (2) the lack of diagnostic marine invertebrates, including conodonts, in the upper member of the Bluestone and in the lower strata of the Pocahontas Formation of the eastern Appalachians; (3)



Text-figure 3. Correlation of Upper Mississippian and Lower Pennsylvanian section from the eastern Appalachian basin with the Ozark Mountains section of northern Arkansas and with the Mississippian "type" section in the southern Illinois basin. European substage correlations based on macrofloral plant zones of Read and Mamay (1964), as modified by Pfefferkorn and Gillespie (1981). Brachiopod zones developed from eastern Appalachian basin. Figure not to scale.

the assumption that the *Rh. muricatus* Zone is a viable biostratigraphic entity; and (4) the acceptance of the recommended I.U.G.S. "standard." However, we are unwilling to concede the veracity of points 3 and 4 above.

REFERENCES CITED

Arkle, T., Jr.; Beissel, D. R.; Larese, R. E.; and others, 1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—West Virginia and Maryland: U.S. Geological Survey Professional Paper 1110-D, 35 p.

Campbell, M. R., 1893, Geology of the Big Stone Gap coal field of Virginia and Kentucky: U.S. Geological Survey Bulletin 111, 106 p.

deWitt, Wallace, Jr.; and McGrew, L. W., 1979, The Appalachian basin region, *in* Craig, L. C.; and Conner, C. W. (eds.), Paleotectonic investigations of the Mississippian System in the United States: U.S. Geological Survey Professional Paper 1010, pt. 1, p. 13–48.

Englund, K. J., 1968, Geologic map of the Bramwell Quadrangle, West Virginia–Virginia: U.S. Geological Survey Geologic Quadrangle Map GQ-745, scale 1:24,000.

______1974, Sandstone distribution patterns in the Pocahontas Formation of southwest Virginia and southern West Virginia: Geological Society of America Special Paper 148, p. 31–45.

______1979, The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Virginia: U.S. Geological Survey Professional Paper 1110-C, 21 p.

Englund, K. J.; and Henry, T. W., 1984, The Mississippian–Pennsylvanian boundary in the central Appalachian basin: Compte rendu, Neuvième Congrès International de Stratigraphie et de Géologie du Carbonifère, Urbana, Illinois, U.S.A., v. 2, p. 330–336.

Englund, K. J.; and Randall, A. H., III, 1981, Stratigraphy of the Upper Mississippian and Lower Pennsylvanian Series in the east-central Appalachians, *in* Roberts, T. G. (ed.), Stratigraphy, sedimentology, *vol. 1 of* Geological Society of America Cincinnati 1981, Field Trip Guidebooks: American Geological Institute, Washington, D.C., p. 154–158.

Englund, K. J.; Arndt, H. H.; Henry, T. W.; Meissner, C. R., Jr.; Windolph, J. F., Jr.; and Warlow, R. C., 1977, Geologic map of the New River Gorge area, Fayette, Raleigh, and Summers Counties, West Virginia: U.S. Geological Survey Open-File Report OF-77-76 Map A, scale 1:50,000.

Englund, K. J.; Henry, T. W.; and Cecil, C. B., 1981, Upper Mississippian and Lower Pennsylvanian depositional environments, southwestern Virginia and southern West Virginia, *in* Roberts, T. G. (ed.), Stratigraphy, sedimentology, *vol. 1 of* Geological Society of America Cincinnati 1981, Field Trip Guidebooks: American Geological Institute, Washington, D.C., p. 171–175.

Englund, K. J.; Henry, T. W.; Gillespie, W. H.; Pfefferkorn, H. W.; and Gordon, Mackenzie, Jr., 1983a, Boundary stratotype for the base of the Pennsylvanian System, east-central Appalachian basin, U.S.A.: Resumenes X Congreso Internacional de Estratigrafia y Geologica del Carbonifero, Madrid, p. 355. [Also published in International Union of Geological Sciences Subcommission on Carboniferous Stratigraphy, Newsletter on Carboniferous Stratigraphy, no. 4, p. 13–14.]

Englund, K. J.; Windolph, J. F., Jr.; Weber, J. C.; and Thomas, R. E., 1983b, Lithologic descriptions and geophysical logs of coreholes in the southwestern Virginia coal field, Dickenson, Lee, Scott, and Wise Counties, Virginia, *part 1 of* Test drilling for coal in 1982–83 in the Jefferson National Forest, Virginia: U.S. Geological Survey Open-File Report 83-628, 374 p.

Englund, K. J.; Henry, T. W.; Gillespie, W. H.; Pfefferkorn, H. W.; and Gordon, Mackenzie, Jr., 1985, Boundary stratotype for the base of the Pennsylvanian System, east-central Appalachian basin, U.S.A.: Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, Compte rendu, v. 4, p. 371–382.

Englund, K. J.; Arndt, H. H.; Schweinfurth, S. P.; and Gillespie, W. H., 1986, Pennsylvanian System stratotype sections, West Virginia: Geological Society of America Centennial Field Guide, Southeastern Section, p. 59–68.

Englund, K. J.; Gillespie, W. H.; Cecil, C. B.; Crawford, T. J.; Windolph, J. F., Jr.; Blake, B. M., Jr.; and Fedorko, Nicholas, 1989, Field Guide to Mississippian and Lower Pennsylvanian coal-bearing rocks in the Appalachian basin, *in* Englund, K. J. (ed.), Characteristics of the Mid-Carboniferous boundary and associated coal-bearing rocks in the central and southern Appalachian basin: I.G.C. Field Trip T352, American Geophysical Union, Washington, D.C., p. 73–118.

Gillespie, W. H.; and Englund, K. J., 1989, New observations on the Mississippian–Pennsylvanian boundary in the Appalachian basin, U.S.A.: Proceedings of the Provo [Utah] Meeting of the International Union of Geological Sciences Subcommission on Carboniferous Stratigraphy, Part 2, Upper Carboniferous, 2 p.

Gordon, Mackenzie, Jr.; and Henry, T. W., 1981, Late Mississippian and Early Pennsylvanian invertebrate faunas, east-central Appalachians—a preliminary report, *in* Roberts, T. G. (ed.), Stratigraphy, sedimentology, *vol. 1 of* Geological Society of America Cincinnati 1981, Field Trip Guidebooks: American Geological Institute, Washington, D.C., p. 165–171.

Harris, L. D.; and Miller, R. L., 1958, Geology of the Duffield Quadrangle, Virginia: U.S. Geological Survey Geologic Quadrangle Map GQ-111, scale 1:24,000.

Henry, T. W.; and Gordon, Mackenzie, Jr., 1979, Late Devonian through Early Permian(?) faunas in area of proposed Pennsylvanian System stratotype, *in* Englund, K. J.; Arndt, H. H.; and Henry, T. W. (eds.), Proposed Pennsylvanian System stratotype, Virginia and West Virginia: American Geological Institute, Selected Guidebook Series, no. 1, p. 97–103.

Lane, H. R.; and Manger, W. L., 1985, The basis for a Mid-Carboniferous boundary: Episodes, v. 8, no. 2, p. 112–115.

Lane, H. R.; Baesemann, J. F.; Brenckle, P. L.; and West, R. R., 1985, Arrow Canyon—a potential Mid-Carboniferous boundary stratotype: Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, Compte rendu, v. 4, p. 429–439.

Manger, W. L.; and Sutherland, P. K., 1984, The Mississippian–Pennsylvanian boundary in the southern Midcontinent, United States: Neuvième Congrès International de Stratigraphie et de Géologie du Carbonifère, Compte rendu, v. 2, p. 369–376.

Pfefferkorn, H. W.; and Gillespie, W. H., 1981, Biostratigraphic significance of plant megafossils near the Mississippian–Pennsylvanian boundary in southern West Virginia and southwestern Virginia, *in* Roberts, T. G. (ed.), Stratigraphy, sedimentology, *vol. 1 of* Geological Society of America Cincinnati 1981, Field Trip Guidebooks: American Geological Institute, Washington, D.C., p. 159–164.

Read, C. B.; and Mamay, S. H., 1964, Upper Paleozoic floral zones and floral provinces of the United States: U.S. Geological Survey Professional Paper 454-K, 35 p.

Reger, D. B., 1924, Mercer, Monroe, and Summers Counties [West Virginia]: West Virginia Geological Survey [County Reports], 963 p.

Repetski, J. E.; and Henry, T. W., 1983, A Late Mississippian conodont faunule from area of proposed Pennsylvanian System stratotype, eastern Appalachians: Fossils and Strata, no. 15, p. 169–170.

Schalla, R. A., 1984, Deltaic deposits of the Upper Mississippian Ravencliff Member of the Hinton Formation, southern West Virginia: Southeastern Geology, v. 25, no. 1, p. 1–12.

Wilpolt, R. H.; and Marden, D. W., 1959, Geology and oil and gas possibilities of Upper Mississippian rocks of southwestern Virginia, southern West Virginia, and eastern Kentucky: U.S. Geological Survey Bulletin 1072-K, p. 587–656.

APPENDIX: USGS Locality and Collection Numbers

Loc.	Mbr.	County	State	7.5' Quadrangle	USGS Collection Number(s)
1.	U	Mercer	WV	Athens	GR702.
2.	L	Wise	VA	Appalachia	3151, 3157, 3173, 5638, 27894, 27895, 27896, 27904, 27905, 27983, 28580.
3.	L	Mercer	WV	Bramwell	3217.
4	L	Tazewell	VA	Bramwell	5513.
5.	U	Tazewell	VA	Bramwell	5514.
6.	U	Summers	WV	Pipestem	5558.
7.	U	Summers	WV	Talcott	5563.
8.	L	Summers	WV	Talcott	5571.
9.	L	Summers	WV	Talcott	5572, 5573.
10.	L	Summers	WV	Pipestem	5577.
11.	B	Mercer	WV	Bramwell	22500.
12.	B	Mercer	WV	Bramwell	22754, 26309, 26676, 26789.
13.	B	Mercer	WV	Bramwell	25730, 26787, 27886.
14.	B	Wyoming	WV	Crumpler	25731.
15.	B	Mercer	WV	Flat Top	25732, 25732A, 25732B.
16.	B	Mercer	WV	Flat Top	25733.
17.	B	Mercer	WV	Bramwell	25734.
18.	P	Raleigh	WV	Hinton	25735, 26724.
19.	L	Raleigh	WV	Prince	26677, 26678, 26720, 26721, 26723.
20.	L	Fayette	WV	Thurmond	26679, 26718.
21.	L	Fayette	WV	Thurmond	26717.
22.	L	Tazewell	VA	Tazewell North	26784, 27918.
23.	B	Tazewell	VA	Bramwell	26788.
24.	B	Mercer	WV	Bramwell	26790, 27853, 27885.
25.	B	Mercer	WV	Bramwell	27811.
26.	B	Mercer	WV	Bramwell	27847.
27.	U	Mercer	WV	Princeton	27850.
28.	B	Mercer	WV	Bramwell	27888, 27889.
29.	P	Mercer	WV	Athens	27911.
30.	P	Mercer	WV	Athens	27912.
31.	U	Summers	WV	Pipestem	27915.
32.	L	Mercer	WV	Princeton	27947, 27948.
33.	U	Mercer	WV	Athens	27984, 27985.
34.	U	Mercer	WV	Athens	28712.
35.	U	Mercer	WV	Athens	28714, 28715, 28716, 28717, 28718.
36.	P	Mercer	WV	Athens	28719, 28720.
37.	U	Mercer	WV	Athens	28721.
38.	U	Mercer	WV	Athens	28722.
39.	L	Wise	VA	Norton	27848, 27903.
40.	P	Fayette	WV	Fayetteville	17970.
41.	L	Fayette	WV	Montgomery	31107.
42.	L	Fayette	WV	Montgomery	31108.
43.	B	Mercer	WV	Bluefield	30665.
44.	B	Scott	VA	East Stone Gap	31180.
45.	X	Wyoming	WV	Crumpler	26701.
46.	X	Wyoming	WV	Crumpler	*

NOTE: The locality numbers correspond to those on Text-figure 2. All USGS collection numbers are from the "blue" upper Paleozoic (-PC) register, unless preceded by the letters "GR," which are from from the old "green" register. Abbreviations for members are as follows: Pocahontas Formation—X = lower part; Bluestone Formation—B = Bramwell Member, P = Pride Shale Member; Hinton Formation—U = upper shale member, L = Little Stone Gap Member ("Avis limestone" of Reger, 1924). Asterisk (*) denotes material noted in outcrop but not collected (see Englund and others, 1986, p. 62; bed ~100 ft below Pocahontas No. 1 coal bed).

Plates

PLATE 1

Brachiopods of Little Stone Gap Member of Hinton Formation; all illustrations ×1, unless otherwise indicated; all specimens are hypotypes; locs. refer to Text-figure 1.

Figures 1–3.—*Adectorhynchus suspectum* (Girty); USNM 369079, USGS colln. 26717-PC (loc. 21); typical mature, slightly broken, articulated specimen; ventral, lateral, posterior views, respectively.

Figures 4–7.—*Orthotetes subglobosus* Girty; USNM 314030, USGS colln. 26718-PC (loc. 20); mature specimen; ventral, dorsal, lateral, posterior (brachial valve up) views, respectively.

Figures 8–10.—*Orthotetes* n. sp. A; USNM 314029, USGS colln. 5638-PC (loc. 2); slightly crushed mature, articulated specimen; lateral, ventral, dorsal views, respectively.

Figures 11,12.—*Echinoconchus* n. sp. A; USNM 369191, USGS colln. 27905-PC (loc. 2); disarticulated, presumably mature specimen; lateral, ventral views, respectively.

Figures 13–16.—*Productus* n. sp. A. **13–15**, USNM 314032, USGS colln. 26717-PC (loc. 21); ventral, anterior, lateral views, respectively, of typical mature specimen. **16**, dorsal view of disarticulated brachial valve exterior, showing band marking edge of visceral disc.

Figures 17–19.—*Ovatia elongata* Muir-Wood and Cooper; USNM 369213, USGS colln. 27848-PC (loc. 39); ventral, anterior, and lateral views, respectively, of mature pedicle valve.

Figures 20,21.—*Marginovatia minor* (Snider); USNM 36924, USGS colln. 26784-PC (loc. 22); ×2, ventral, lateral views, respectively, of mature pedicle valve.

Figures 22,23.—*Diaphragmus cestriensis* (Worthen); USNM 314031, USGS colln. 27848-PC (loc. 39); ×2, ventral, anterior views, respectively, of mature specimen.

Figures 24–27.—*Inflatia cherokeensis* (Drake); USNM 369203, USGS colln. 26718-PC (loc. 20); ventral, anterior, dorsal, lateral views, respectively, of complete, mature specimen.

Figures 28,29.—*Martinia contracta* (Meek and Worthen); USNM 442036, USGS colln. 27905-PC (loc. 2); ×1.5, ventral, lateral views, respectively, of mature articulated specimen.

Figure 30.—*Composita subquadrata* (Hall); USNM 314037, USGS colln. 27918-PC (loc. 22); ×1.5, ventral view of mature pedicle valve.

Figure 31.—*Anthracospirifer* sp. A; USNM 441978, USGS colln. 27905-PC (loc. 2); ×2, ventral view of small broken pedicle valve.

Figure 32.—*Eumetria costata* (Hall); USNM 442069, USGS colln. 27905-PC (loc. 2); ×2, ventral view of small pedicle valve.

Figures 33,34.—*Coledium cestriense* (Snider); USNM 441893, USGS colln. 27905-PC (loc. 2); ×2, posterior, ventral views of mature specimen.

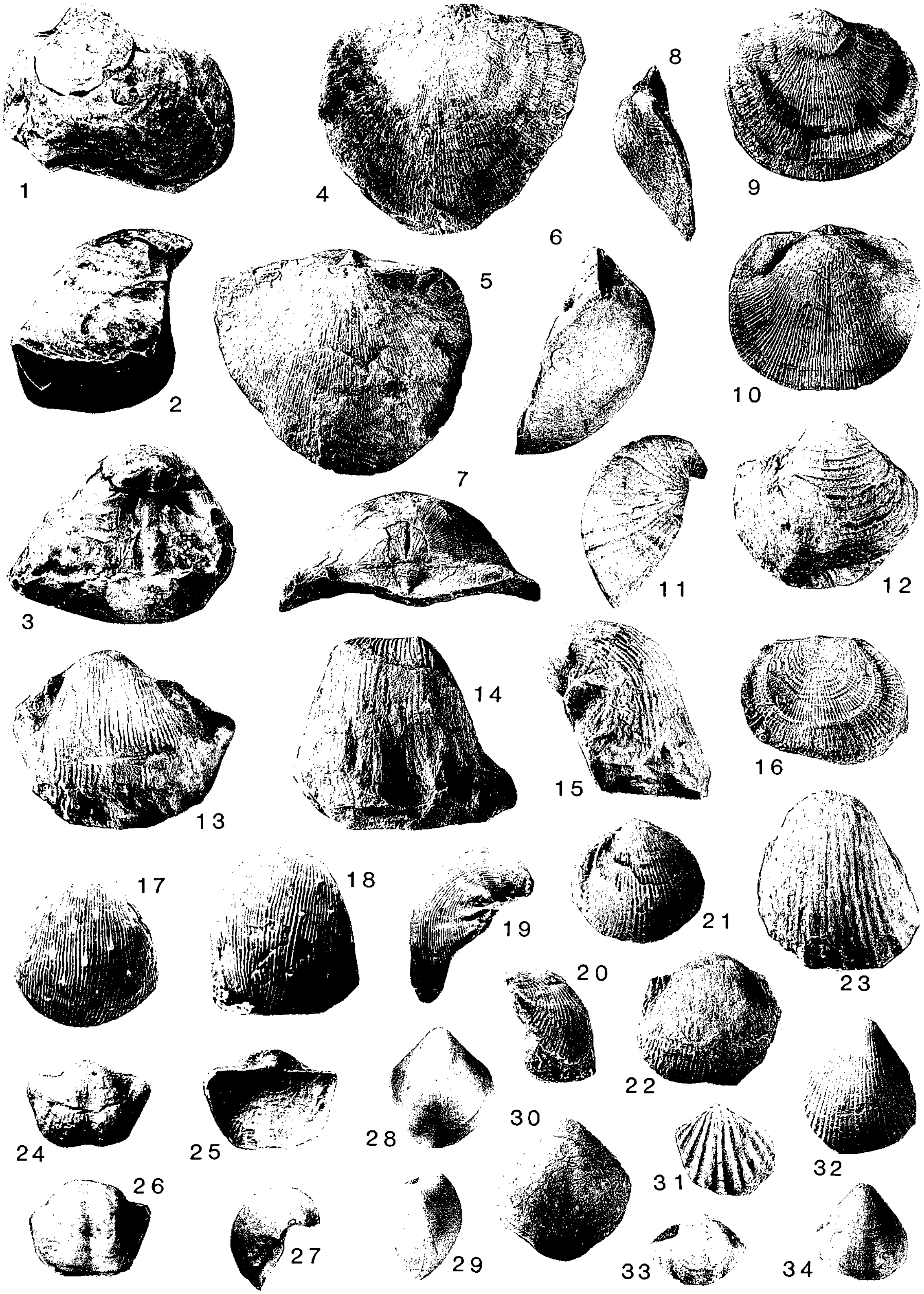

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PLATE 2

Brachiopods of upper shale member of Hinton Formation; all illustrations ×1, unless otherwise indicated; all specimens are hypotypes; locs. refer to Text-figure 1.

Figure 1.—*Orthotetes* n. sp. A; USNM 369100, USGS colln. 28715-PC (loc. 35); ventral view of mature, disarticulated pedicle valve.

Figures 2,3.—*Diaphragmus cestriensis* (Worthen); USNM 369173, USGS colln. 28712-PC (loc. 34); ventral, anterior views, respectively, of typical specimen.

Figures 4–6.—*Productus*? n. sp. B; USNM 369155, USGS colln. 27984-PC (loc. 33); ventral, anterior, lateral views respectively of mature specimen.

Figures 7–9.—*Inflatia* n. sp. A; USGS colln. 27985-PC (loc. 33). **7**, USNM 369120; ventral view of decorticated pedicle valve exterior, showing body spines on anterolateral margin. **8**, USNM 369203; ventral view of internal cast of pedicle valve, showing body spines. **9**, USNM 369138; ×1.5, ventral view of brachial valve interior.

Figures 10–15.—*Ovatia elongata* Muir-Wood and Cooper; USGS colln. 28717-PC (loc. 35). **10–12**, USNM 369209; ventral, lateral, anterior views, respectively, of mature pedicle valve. **13–15**, USNM 369210; anterior, ventral, lateral views, respectively, of mature specimen with slightly crushed trail.

Figure 16.—*Flexaria arkansana* (Girty); USNM 369214, USGS colln. 5563-PC (loc. 7); ventral view of small, well-preserved pedicle valve exterior.

Figures 17,18.—*Fluctuaria* n. sp. A; USNM 369218, USGS colln. 5558-PC; ×2, ventral, lateral views of disarticulated pedicle valve exterior.

Figure 19.—*Composita subquadrata* (Hall); USNM 441979, USGS colln. 5514-PC (loc. 5); ×1.5, ventral view of internal cast of pedicle valve.

Figure 20.—*Punctospirifer transversus* (McChesney); USNM 442056, USGS colln. 26723-PC (loc. 19); dorsal view of broken, disarticulated brachial valve.

Figure 21.—*Eumetria costata* (Hall); USNM 442065, USGS colln. 5514-PC; ventral view of internal cast of slightly crushed pedicle valve.

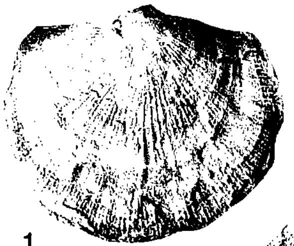
Figure 22.—*Torynifer setiger* (Hall); USNM 442029, USGS colln. 28718-PC (loc. 35); dorsal view of disarticulated brachial valve.

Figures 23–25.—*Cleiothyridina atrypoides* Girty; USNM 441937, USGS colln. 27985-PC (loc. 33); ventral, lateral, anterior views, respectively, of typical specimen.

Figures 26–28.—*Coledium cestriense* (Snider); USNM 441900, USGS colln. 27984-PC (loc. 33); ×2, ventral, anterior, lateral views, respectively, of slightly crushed specimen.

Figures 29,30.—*Anthracospirifer leidyi* (Norwood and Pratten). **29**, USNM 369216, USGS colln. 28712-PC (loc. 34); ventral view of internal cast of pedicle valve. **30**, USNM 442049, USGS colln. 28718-PC (loc. 35); ventral view of badly decorticated pedicle valve exterior.

Figure 31.—*Reticulariina spinosa* (Norwood and Pratten); USNM 442062, USGS colln. 28712-PC (loc. 34); ventral view of crushed brachial valve exterior.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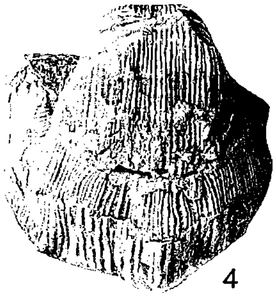
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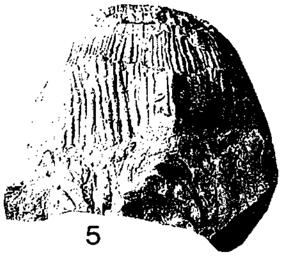
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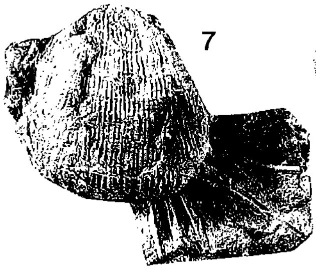
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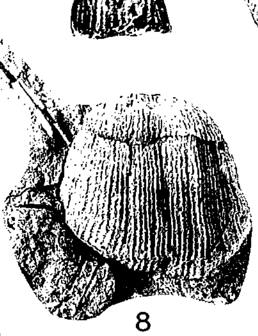
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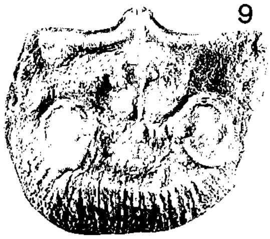
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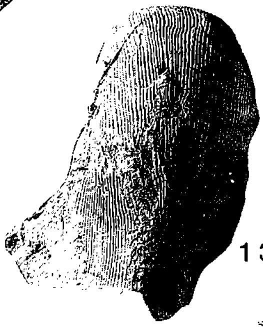
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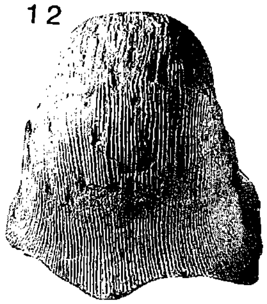
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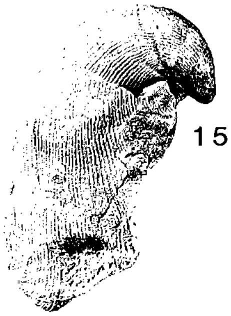
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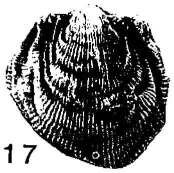
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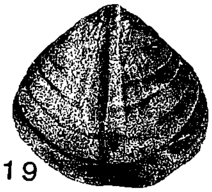
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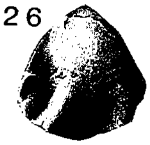
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PLATE 3

Brachiopods of Pride Shale Member (above line) and Bramwell Member of Bluestone Formation (below line); all illustrations ×1, unless otherwise indicated; all specimens are hypotypes; locs. refer to Text-figure 1.

Figures 1–6.—*Leiorhynchoidea carbonifera* (Girty). **1–5**, USNM 441912, USGS colln. 28720-PC (loc. 36); ventral, posterior, anterior, dorsal, and lateral views, respectively, of mature specimen. **6**, USGS colln. 27911-PC (loc. 29); specimens in approximate growth position from within large, very dark-gray dolomitic carbonate nodule; spiraling clockwise from largest specimen near center to smallest specimen in lower left corner, USNM 441902-441908; all specimens except USNM 441902 and 441908 have pedicle valve oriented toward observer.

Figures 7–10.—*Crurithyris fayettevillensis* Girty; USGS colln. 28720-PC (loc. 36). **7–9**, USNM 441929; ×4, lateral, ventral, anterior views of articulated specimen. **10**, USNM 441928; ×4, dorsal view of decorticated brachial valve interior.

Figure 11.—*Buxtonia semicircularis* (Sutton and Wagner); USNM 369199, USGS colln. 27820-PC (loc. 29); ventral view of fragment of pedicle valve.

Figures 12–16.—*Orthotetes* n. sp. B, aff. *O. kaskaskiensis* (McChesney); all USGS colln. 26789-PC (loc. 12), unless otherwise indicated. **12**, USNM 314046; dorsal view of pedicle valve interior. **13**, USNM 369111; ventral view of internal mold of brachial valve. **14**, USNM 369103; ×4, dorsal view of cardinal process of immature specimen. **15**, USNM 369112, USGS colln. 22500-PC (loc. 11); dorsal view of decorticated pedicle valve interior. **16**, USNM 314048; ventral view of immature brachial valve interior.

Figures 17–21.—*Ovatia* n. sp. A, aff. *Ov. elongata* Muir-Wood and Cooper, n. sp. A. **17,18**, USNM 369228, USGS colln. 26788-PC (loc. 23); posterolateral view of hinge area showing spines, ×4; ventral view of immature pedicle valve. **19,20**, USNM 369224, USGS colln. 26789-PC (loc. 12); ventral, lateral views, respectively, of broken pedicle valve. **21**, USNM 369230, USGS colln. 26788-PC (loc. 23); ventral view of almost complete brachial valve interior.


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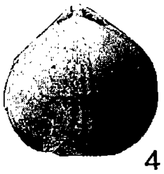
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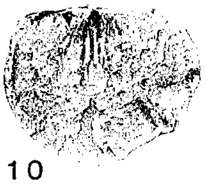
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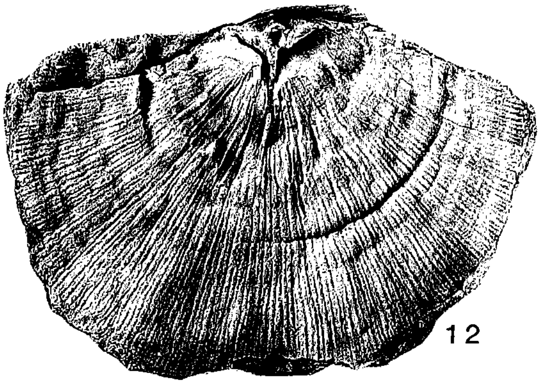
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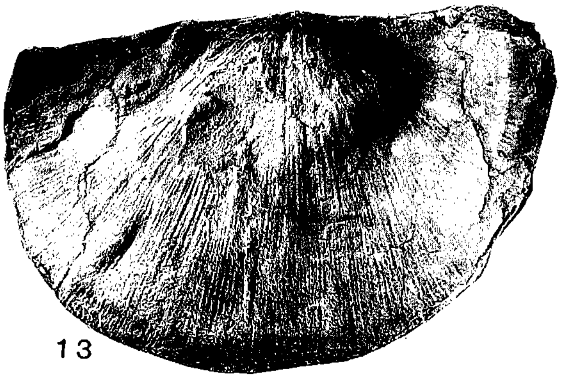
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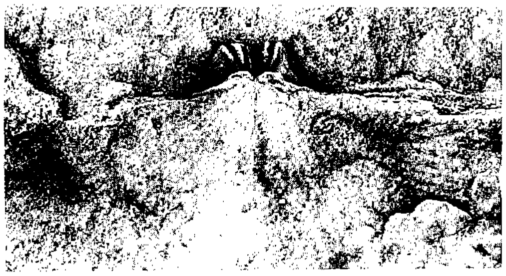
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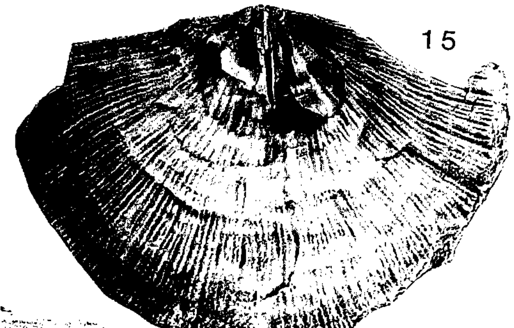
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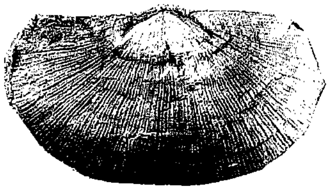
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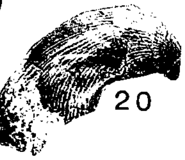
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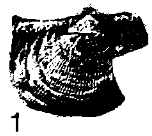
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PLATE 4

Brachiopods of Bramwell Member of Bluestone Formation; all illustrations ×2, unless otherwise indicated; all specimens are hypotypes; locs. refer to Text-figure 1.

Figures 1–5.—*Lingula* n. sp. A; USGS colln. 27889-PC (loc. 28). **1,** USNM 314044. **2,** USNM 314193. **3,** USNM 314194, ×3. **4,** USNM 314191. **5,** USNM 314190.

Figures 6–16.—*Diaphragmus* n. sp. A, aff. *D. cestriensis* (Worthen). **6,** USNM 369181, USGS colln. 26789-PC, (loc. 12); anterior view of mature pedicle valve, ×1. **7–11,** USNM 314050, USGS colln. 22754-PC (loc. 12); views of mature pedicle valve, ventral view, ×1; ventral, anterior, lateral, posterior views, respectively. **12,** USNM 369189, USGS colln. 26788-PC (loc. 23); dorsal view of brachial valve exterior. **13,** USNM 369183, USGS colln. 26789-PC (loc. 12); anterior view of large pedicle valve. **14,** USNM 369178, USGS colln. 26789-PC (loc. 12); ventral view of disarticulated pedicle valve. **15,** USNM 369187, USGS colln. 26788-PC (loc. 23); ventral view of latex cast of mold of brachial valve interior. **16,** USNM 369186, USGS colln. 26788-PC (loc. 23); ventral view of latex cast of mold of brachial valve interior.

Figures 17–19.—*Schizophoria* sp.; USGS colln. 26790-PC (loc. 24). **17,18,** USNM 369078; ventral views of mold of brachial valve interior, ×1, ×3, respectively. **19,** dorsal view of mold of broken pedicle valve interior, ×3.

Figures 20–22.—*Eumetria costata* (Hall and Clarke). **20,** USNM 442064, USGS colln. 22500-PC (loc. 11); ×1.5 dorsal view of internal mold of brachial valve interior of broken specimen. **21,** USNM 442063, USGS colln. 22500-PC (loc. 11); ×1.5, dorsal view of internal mold of brachial valve interior. **22,** USNM 314052, USGS colln. 26789-PC (loc. 12); ×1, ventral view of internal mold of pedicle valve interior.

Figure 23.—*Composita subquadrata* (Hall); USNM 314054, USGS colln. 26789-PC (loc. 12); ×1.5, ventral view of pedicle valve exterior.

Figures 24–27.—*Anthracospirifer leidyi* (Norwood and Pratten). **24,** USNM 314053, USGS colln. 26788-PC (loc. 23); ventral view of pedicle valve exterior. **25,** USNM 442048, USGS colln. 26788-PC (loc. 23); ventral view of pedicle valve exterior. **26,** USNM 442044, USGS colln. 26790-PC (loc. 24); ×1, ventral view of pedicle valve exterior. **27,** USNM 44243, USGS colln. 26788-PC (loc. 23); ×1.5, dorsal view of brachial valve exterior.


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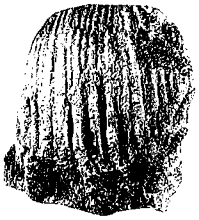
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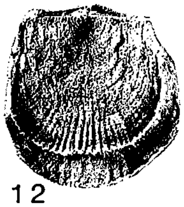
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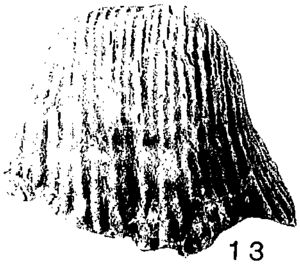
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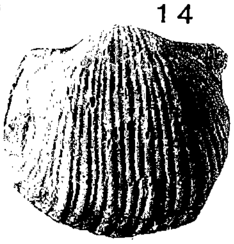
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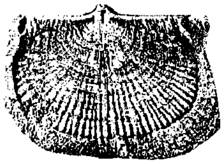
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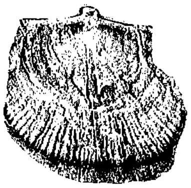
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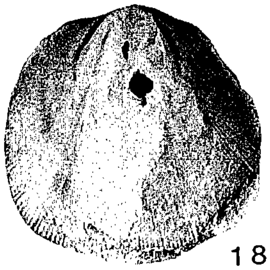
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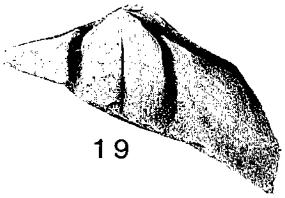
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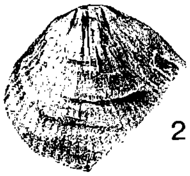
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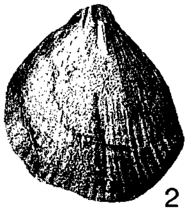
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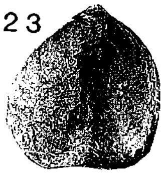
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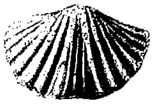
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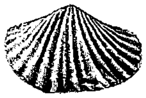
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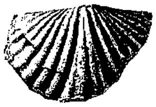
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Oklahoma Geological Survey Circular 94, 1992

Intraspecific Variability in Axial Morphology of *Archimedes*, Pitkin Limestone (Chesterian), Arkansas

Frank K. McKinney
Appalachian State University
Boone, North Carolina

ABSTRACT.—Two species of *Archimedes*, *A. lunatus* and *A. magnus*, occur in the Pitkin Limestone (Chesterian) of Arkansas. The two species differ in mesh and zooecial size, zooecial shape, skeletal microstructure, and the presence of stellate apertures in *A. lunatus* and of secondary nanozooecia in *A. magnus*. In addition, both species exhibit widely varying characteristics of the central axial support, including spacing between successive whorls and breadth of the axial screw, which encompasses the characters of several other Pitkin species of *Archimedes* regarded here as synonyms.

INTRODUCTION

Conspicuous, screw-shaped axes of the fenestrate bryozoan *Archimedes* are widespread geographically and are distributed, commonly in great abundance, through most carbonate facies and calcareous siliciclastic rocks of Chesterian (latest Viséan–Namurian A) and underlying Valmeyeran (remaining Viséan excepting V_{1a}) ages in eastern and south-central North America. The extraordinary conspicuousness of the fossil resulted in the limestones of this interval from the Mississippian Valley to the Appalachians being called the "*Archimedes* limestone," following the indication by Owen (1838, p. 13) that it is "... characterized by a fossil, described by Lesueur under the name of *Archimedes*, on account of its screw-like form." Lesueur never actually described *Archimedes*, so that Owen's introduction of the name led to a taxonomic problem that took a long time to settle (Easton and Duncan, 1953).

Shumard (1853) first recognized *Archimedes* in Arkansas (as *Archimedipora archimedes*). Shumard's illustrated specimen was considered unrecognizable by Easton and Duncan (1953), whereas Condra and Elias (1944, p. 14) surmised that it belonged to one of their new, but unspecified, species, and Horowitz (1977) considered it comparable to axes of *A. moorei* Condra and Elias, 1944. To the present, 25 species names have been applied to Chesterian *Archimedes* of northern Arkansas and eastern Oklahoma (Horowitz, 1977), 17 of which have been used for specimens from the Pitkin Limestone. Although found only in Chesterian Rocks of Arkansas and Oklahoma, *Archimedes* is known to range elsewhere from early Viséan (V_{1a}) through Artinskian (late Early Permian). Geologic and geographic ranges are well summarized by Condra and Elias (1944) and by Ross (1978,1981).

This paper presents a restudy of *Archimedes* from the Pitkin Limestone of northern Arkansas. It is based largely on collections made during the summer of 1977.

Lithologic and Environmental Distribution

The only specimens of either species collected *in situ* come from an algal-bryozoan boundstone mound at locality 1977–79 (see Appendix). These specimens of *Archimedes lunatus* retain complete sets of the delicate branch whorls and were oriented with the axis vertical and with the distal growing tip uppermost. Unidentified species of *Archimedes* have been reported *in situ* from other boundstone mounds and toppled into peripheral talus within the Pitkin (Tehan and Warmath, 1977; Webb, 1987).

All other specimens collected in this study are axial supports denuded of the delicate branch whorls and have been transported. *A. lunatus* and *A. magnus* are associated in some units; they also occur independently of one another, although in a similar range of lithologies. The grainstones and packstones that contain most of these denuded axial supports were deposited in shoals and were probably not the environments in which they lived. Elsewhere, *Archimedes* occur *in situ* most abundantly on mud bottoms leeward of shoals (McKinney, 1979; Whisonant and Scolaro, 1979; McKinney and Gault, 1980; Wulff, 1990). Populations in the Pitkin sea probably lived in the same, protected environment and were incorporated into the shoal deposits as they migrated across the region.

In both species,the smallest whorl spacing (<5 mm vs. >5 mm in other beds) is found in specimens in a skeletal ooid grainstone (locality 1977–79, unit 5) that apparently was deposited on or near the crest of a shoal. Although it is possible that such distribution results from sorting during transportation, the axial screws are so dense that they most likely were not transported great distances from the site in which they lived. It is more likely that populations living in relatively turbulent water closer to the shoal crests had more compact growth and closer whorl spacing and were more readily transported into shoal-crest deposits than were more distant populations.

Acknowledgments

I thank H. W. Gault, M. J. McKinney, and E. Fuchs for assistance in the field and specimen preparation, and A. H. Horowitz and R. J. Cuffey for critical reviews of the manuscript. R. Pabian of the Nebraska Geological Survey kindly supplied all that could be located of the pertinent type specimens from Condra and Elias's study of *Archimedes*. This work was supported by NSF Grant EAR-8419006. In addition, acknowledgment is made to the Donors of the Petroleum Research Fund, administered by the American Chemical Society, for support of this research.

SYSTEMATIC PALEONTOLOGY

Phylum Bryozoa
Class Stenolaemata
Order Fenestrida
Genus **Archimedes**

Archimedes lunatus Condra and Elias, 1944
Text-figure 1A; Plates 1 and 2

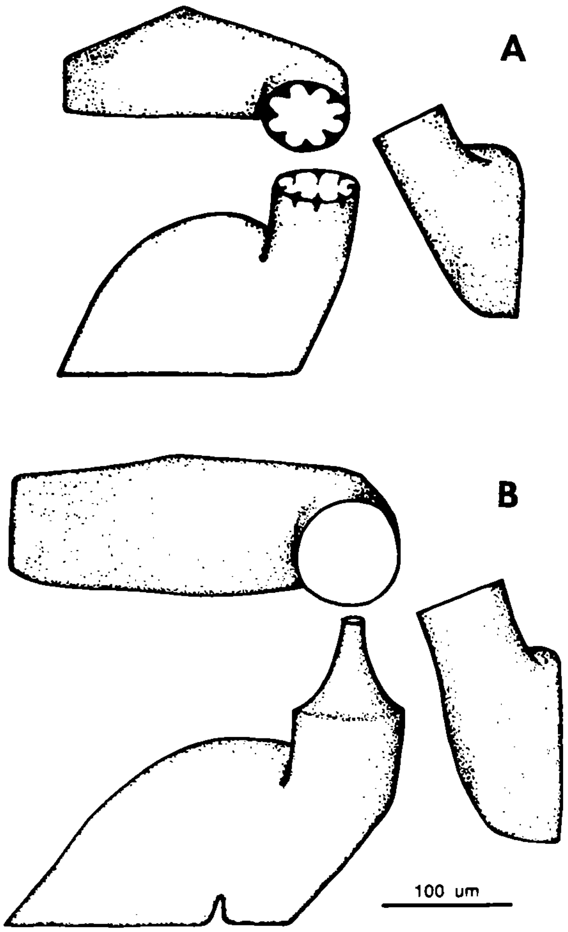
Archimedes lunatus CONDRA and ELIAS, 1944, p. 132–133, pl. 1, fig. 2; pl. 11, figs. 4–7; pl. 26, figs. 14–16.
Archimedes fosteri CONDRA and ELIAS, 1944, p. 113–115, pl. 23, figs. 1–4.
Archimedes moorei CONDRA and ELIAS, 1944, p. 123–125, pl. 20, figs. 6,7; pl. 24, figs. 6,7.
Archimedes pitkinensis ELIAS, 1957, p. 418–419, pl. 47, figs. 1–5.
Archimedes magnus CONDRA and ELIAS, 1944, p. 154–157 (in part: nonillustrated paratype, Nebraska Geological Survey 243b).

Diagnosis.—Distinguished from similar species by relatively fine meshwork characteristics, presence of "trident" keel spines and stellate apertures.

Description.—Zoarial robustness as reflected in axial structure varies from moderately delicate to moderately robust (Pl. 1, Figs. 1,2); whorl spacing varies from ~3 mm to ~6.5 mm. Diameter of axial core structure varies from 2 mm to 11 mm, and that of axis including flanges varies from 4 to 20 mm. Whorls of branches diverge from the axis at a relatively constant angle, averaging 65°.

The fenestrate meshwork is relatively delicate, with closely spaced, narrow branches and dissepiments (Pl. 1, Figs. 5,6; Pl. 2, Figs. 4,6). Beyond the axial region, where skeletons are thick and fenestrules are infilled, fenestrules are slightly narrower than branches and occupy ~28% of the mesh area.

Zooecia alternate from side-to-side along branches (Pl. 2, Fig. 6) and have a pentagonal to rarely triangular cross-sectional shape at their bases, associated with a strongly zigzag median wall that is expressed along the frontal surface as a corresponding, but more subdued, sinuous keel (Pl. 2, Fig. 4). The reverse zooecial walls are transversely curved so that zooecia are deepest along the axial wall (Pl. 1, Fig. 3); the frontal surface of zooecial chambers is gently convex, with a superior hemiseptum located at the junction of the inflated chamber and the distal tube (Pl. 2, Fig. 1). Maximum chamber width averages slightly less than half zooecial length as measured parallel with the branch axis, and maximum chamber depth averages slightly less than two-thirds zooecial length. The vertical wall between zooecia in a row diverges



Text-figure 1. Zooecial chamber shapes in (A) *Archimedes lunatus* and (B) *A. magnus*, drawn from frontal (top), distal (right) and lateral (lower) perspectives.

from the reverse wall at a steep angle, averaging 67°. The distal tube extends obliquely outward from the disto-lateral corner of the inflated chamber, at an ~30° angle with respect to the branch midplane. Diameter of the distal tube equals about one-third chamber length. There commonly are eight scallops with intervening projections pointing toward the axis of the distal tube at its outer end, producing stellate apertures (Text-fig. 1; Pl. 1, Fig. 6; Pl. 2, Fig. 7). Where each inward projection originates from the circular zooecial aperture, a small spine protrudes (Pl. 1, Figs. 3,4; Pl. 2, Fig. 7), with a core of granular skeleton that averages 9 μm in diameter.

Zooecial walls consist of thick granular wall on the reverse and lateral sides, and thin—locally absent—granular wall on the frontal side and surrounding the distal tubes. Reverse and lateral granular walls have thick longitudinal ridges on their outer sides (Pl. 1, Fig. 3; Pl. 2, Fig. 2). A thin inner lining of laminar wall is commonly present in older zooecia.

Two types of heterozooecia are present. Giant zooecia are found in all colonies and constitute the inner margin of the branch systems in the center of the massive axial screw (Pl. 1, Fig. 5). Typically the giant zooecia are ~50% greater than autozooecia in their linear chamber dimensions, although some may be more than twice the length of autozooecia. The second type of heterozooecium is a single zooecium immediately below a branch bifurcation, where three autozooecia would be expected. The single zooecium has short longitudinal walls extending from the proximal transverse wall that would have divided the space into three autozooecia had they been complete. This single enormous zooecium may be the skeleton of a gonozooid.

A low median keel extends along the frontal surface and contains a slightly sinuous core of granular skeleton. Within the axial skeleton, large spines extend from the median keel at ~0.15-mm spacing; ~0.20 mm above the zooecial chambers, the spines give rise to one or more lateral projections that curve frontally. The spines and their lateral projections produce a trident (or "polydent") pattern as seen in branch cross sections (Pl. 1, Fig. 3). They produce transverse strips of granular skeleton where sections tangential to branches cut through or just above the point of origin of the lateral projections (Pl. 1, Fig. 6; Pl. 2, Figs. 3,7). The trident spines extend ~0.5 mm above frontal surfaces of zooecia. Laminar skeleton deposited above the frontal side of zooecia contains tiny microstyles that have an average spacing of 10 μm.

Laminar skeleton deposited on the reverse side of branches contains tiny microstyles spaced ~12 μm apart; larger styles (macrostyles) with granular skeletal cores of 0.02-mm diameter are spaced on average 0.06 mm apart. Macrostyles extend through the entire thickness of laminar skeleton, either to the reverse branch surface or through up to several millimeters of skeleton in axial screws.

Discussion.—Initial discrimination of this group of specimens as the same species was based on measurements of zooecial chamber dimensions and spacing of branches and dissepiments. Visual inspection confirmed that these specimens were similar in skeletal microstructure, distribution and shape of the unique trident spines extending from the frontal keel, and common presence of the eight scallops and intervening inward projections at outer ends of distal tubes. Grouping of the specimens on zooecial, skeletal, and mesh characteristics results in a broad range of robustness in axial screws being included. Width and whorl spacing of axial screws are therefore interpreted as a variable characters.

Archimedes fosteri Condra and Elias and *A. moorei* Condra and Elias are here considered synonymous with *A. lunatus*. All three species have closely similar branch and dissepiment spacing, aperture spacing, and spacing of nodes along the frontal keel (Table 1). The primary differences between these three putative species as noted by Condra and Elias are in the robustness of the axial screw (axis slender with whorls 5–5.5 mm apart in *A. fosteri*, axis intermediate with whorls 4.5–5.5 mm apart in *A. lunatus*, and axis very robust with whorls 6.5–10 mm in *A. moorei*). Condra and Elias (1944, p. 123) noted presence of transverse "luniform bars" in *A. moorei* as well as in *A. lunatus* (p. 132, 133), but no mention of them was made for *A. fosteri* (p. 113–115). The differences in robustness of the axial screw are not considered here to be of taxonomic significance, but to be a plastic response to environmental conditions, because of the indistinguishable characteristics of zooecia, mesh, and skeleton across the range of robustness. Absence of trident spines in specimens assigned to *A. fosteri* by Condra and Elias is likely due to the slight calcification of the axial screw, or they may be present but inconspicuous or partially developed due to incomplete calcification of the axis.

Measurements.—See Table 2.

Material.—Locality 77–79, unit 3 (3 specimens), unit 5 (10 specimens), algal-bryozoan boundstone mound (2 specimens), float (4 specimens); locality 77–81, locality 77–82, unit 10 (1 specimen), unit 27 (8 specimens); locality 77–83, unit 1 (1 specimen), unit 5 (1 specimen),unit 6 (2 specimens), unit 7 (4 specimens), float (1 specimen); Nebraska Geological Survey (NGS) 242, holotype of *A. lunatus* Condra and Elias, 1944; NGS 240-1 (holotype) and 240-2 (paratype), probably fragments of same colony of *A. fosteri* Condra and Elias, 1944; NGS 245a, paratype of *A. moorei* Condra and Elias, 1944; NGS 243b, nonillustrated paratype of *A. magnus* Condra and Elias, 1944.

TABLE 1. — COMPARISON OF AVERAGES OF COUNTS OF FENESTRATE MESHWORK CHARACTERISTICS OF *ARCHIMEDES LUNATUS*, *A. FOSTERI*, AND *A. MOOREI* AS GIVEN BY CONDRA AND ELIAS (1944, p. 114, 124, 133), AND *A. PITKINENSIS* ELIAS (1957, p. 418) WITH THEIR CONVERSION TO LINEAR MEASUREMENTS

	A. lunatus	A. fosteri	A. moorei	A. pitkinensis
branch spacing	26/10 mm (0.38 mm)	24/10 mm (0.41 mm)	27/10 mm (0.37 mm)	28/10 mm (0.36 mm)
dissepiment spacing	20/10 mm (0.49 mm)	25/10 mm (0.39 mm)	24/10 mm (0.41 mm)	22/10 mm (0.44 mm)
aperture spacing	27/5 mm (0.19 mm)	27/5 mm (0.19 mm)	27/5 mm (0.19 mm)	27/5 mm (0.19 mm)
frontal node spacing	27/5 mm (0.19 mm)	36/5 mm (0.14 mm)	37/5 mm (0.13 mm)	40/5 mm (0.12 mm)

Archimedes magnus Condra and Elias, 1944
Text-figure 1B; Plates 3 and 4

Archimedes magnus CONDRA and ELIAS, 1944, p. 154–157, pl. 6, fig. 1; pl. 32, figs. 2–9 (not including nonillustrated paratype NGS 243b).

Archimedes proutanus CONDRA and ELIAS, 1944, p. 147–150, pl. 30, figs. 1–11 (not ULRICH, 1890, p. 576–577, pl. 63, figs. 3–3d).

?*Archimedes fragilis* CONDRA and ELIAS, 1944, p. 127–129, pl. 3, figs. 1,2; pl. 25, figs. 7,8.

Diagnosis.—Distinguished from similar species by relatively robust meshwork, presence of robust, but non-dichotomous, keel spines and styles on reverse side of branches, and secondary nanozooecia in axial screw.

Description.—Zoarial robustness as reflected in axial structure varies from delicate to robust (Pl. 3, Figs. 1–4); whorl spacing varies from ~3 mm to ~8.5 mm. Diameter of axial core structure varies from 1.6 to 12 mm, and that of the axial structure including flanges varies from 3.2 to 19 mm. Whorls of branches diverge from the axis at a relatively constant angle, averaging 68°.

The fenestrate meshwork is relatively robust (Pl. 4, Figs. 1,3,5), with branches and dissepiments more widely spaced and broader than in *A. lunatus*. Beyond the axial region, where skeletons are thick and fenestrules are infilled, fenestrules are slightly narrower than branches and occupy ~24% of the mesh area.

Zooecia alternate from side-to-side along branches and have a rectangular to slightly pentagonal cross-sectional shape at their bases, associated with a slightly zigzag median wall (Pl. 4, Fig. 3). The reverse zooecial walls are gently curved transversely so that zooecia are deepest along the axial wall (Pl. 3, Fig. 7); the frontal surface of zooecial chambers is gently convex. A superior hemiseptum is located at the junction of the inflated chamber and the distal tube, and a strongly developed inferior septum extends across the floor of the chamber, at about one-fourth the distance from the distal to the proximal transverse wall (Pl. 4, Fig. 3). Maximum chamber width averages slightly less than half zooecial length as measured parallel to the branch axis, and maximum chamber depth averages slightly less than three-fourths zooecial length. The vertical wall between zooecia in a row diverges from the reverse wall at a moderate angle, averaging 52°. The distal tube extends obliquely outward from the disto-lateral corner of the inflated chamber, at an angle ~18° with respect to the branch midplane (Pl. 3, Fig. 7). Diameter of the distal tube equals about one-third chamber length. Commonly, secondary nanozooecia are located within the axial screw, where they developed as a final stage before zooecia became engulfed by developing axial tissue. The secondary nanozooecia are characterized by an inverted funnel capping the distal tube (Text-fig. 1; Pl. 3, Fig. 7; Pl. 4, Fig. 2).

Zooecial walls consist of thin granular wall on the reverse, lateral, and frontal sides. Granular wall is very thin and appears to be locally absent around distal tubes. Reverse and lateral granular walls have thick longitudinal ridges on their outer sides (Pl. 3, Fig. 7; Pl. 4, Fig. 2). A thin inner lining of laminar wall is commonly present in older zooecia.

Zooecia vary in width along branches; they are broadest at dissepiments and narrowest midway between dissepiments. One unusually broad zooe-

TABLE 2. — MEASUREMENTS OF SPECIMENS OF *ARCHIMEDES LUNATUS* FROM THE PITKIN LIMESTONE OF ARKANSAS

	Mean	Range	Standard deviation	Number of	
				Specimens	Measurements
whorl spacing	4.8	3.1-6.4	1.0	18	36
angle, whorl - axis	64.5	48-72	5.3	7	30
axial diameter, including flange	9.2	4-20	4.6	19	19
branch spacing	0.347	.300-.430	0.033	28	242
branch width	0.183	.148-.224	0.020	28	179
dissepiment spacing	0.398	.302-.600	0.054	27	243
dissepiment width	0.135	.122-.149	0.014	4	35
aperture spacing along row	0.180	.152-.235	0.022	27	224
aperture spacing across branch	0.182	.135-.232	0.022	27	233
aperture spacing between branches	0.177	.161-.189	0.012	7	70
diameter of distal tube	0.061	.044-.073	0.008	26	211
chamber length parallel with base	0.179	.151-.234	0.022	27	224
maximum chamber width	0.085	.066-.100	0.010	28	258
maximum chamber depth	0.108	.083-.128	0.013	24	223
angle, reverse - transverse walls	68	60-80	5.4	7	49
angle, distal tube - branch midplane	29	22-38	5.3	3	6
diameter of keel nodes	0.040	0.32-.040	0.001	6	57
spacing of keel nodes	0.148	.128-.185	0.017	4	40

NOTE: Linear measurements are in millimeters, and angles are measured in degrees.

cium at a branch bifurcation and beside a dissepiment may be the skeleton of a gonozooid (Pl. 4, Fig. 3), or its unusual size and shape may be microenvironmentally determined. Distinct heterozooecia include the secondary nanozooecia and the highly enlarged zooecia that constitute the inner margin of the branch systems in the centers of the axial screws (Pl. 3, Figs. 5,6). Typically the giant zooecia are ~50% greater than autozooecia in their linear chamber dimensions, although some may be about twice the length of autozooecia.

A low median keel extends along the frontal surface but a narrow core of granular skeleton is only locally developed (Pl. 4, Fig. 1). Large spines extend from the median keel at ~0.18-mm intervals (Pl. 3, Fig. 7; Pl. 4, Fig. 1). The spines have a thick core of granular skeleton from which granular cores of microstyles arise in profusion, giving the spine a hirsute appearance (Pl. 4, Fig. 6). The spines may extend for several millimeters through the skeleton of the axial screws, and, where the surface is well-preserved, the spines may project at least 0.25 mm beyond the general skeletal surface. Laminar skeleton deposited above the frontal side of zooecia contains a range of styles penetrating it (Pl. 4, Fig. 4); larger styles have a diameter of ~19 μm and are spaced ~46 μm apart.

Laminar skeleton deposited on the reverse side of branches contains macrostyles with cores that average 37 μm in diameter, but are somewhat smaller near their point of origin along ridges on the reverse granular zooecial walls and larger as they extend away (Pl. 3, Figs. 5–8; Pl. 4, Fig. 5). The macrostyles are spaced an average of 107 μm from center to center and, as do the frontal spines, give rise to a plethora of tiny microstyles. Both sizes of styles may extend for several millimeters away from reverse sides of branches through thick skeleton of axial screws.

Discussion.—Initial discrimination of this group of specimens as the same species was based on measurements of zooecial chamber dimensions and spacing of branches and dissepiments. Visual inspection of specimens confirmed that they were similar in skeletal microstructure, distribution and shape of the hirsute keel spines and reverse-side macrostyles, and common presence of secondary nanozooecia. Grouping of the specimens on zooecial, skeletal, and mesh characteristics results in a broad range of robustness in axial screws being included. Width and whorl spacing of axial screws are, therefore, interpreted for this species, as for *A. lunatus*, as variable characters.

Archimedes fragilis Condra and Elias is here considered a probable synonym with *A. magnus*. Both species have closely similar branch, dissepiment, and aperture spacing, although the type specimens

TABLE 3. — COMPARISON OF AVERAGES OF COUNTS OF FENESTRATE MESHWORK CHARACTERISTICS OF *ARCHIMEDES MAGNUS* AND *A. FRAGILIS* AS GIVEN BY CONDRA AND ELIAS (1944, p. 128, 156) WITH THEIR CONVERSION TO LINEAR MEASUREMENTS

	A. magnus	A. fragilis
branch spacing	23/10 mm (0.44 mm)	20/10 mm (0.49 mm)
dissepiment spacing	15/10 mm (0.66 mm)	16/10 mm (0.62 mm)
aperture spacing	24/5 mm (0.21 mm)	25/5 mm (0.20 mm)
frontal node spacing	32/5 mm (0.16 mm)	17/5 mm (0.29 mm)

TABLE 4. — MEASUREMENTS OF SPECIMENS OF *ARCHIMEDES MAGNUS* FROM THE PITKIN LIMESTONE OF ARKANSAS

	Mean	Range	Standard deviation	Number of Specimens	Number of Measurements
whorl spacing	5.3	2.5-8.9	1.5	33	66
angle, whorl - axis	69.7	60-75	3.8	4	20
axial diameter, including flange	8.9	3.2-19.3	3.6	24	24
branch spacing	0.415	.337-.507	0.039	29	210
branch width	0.277	.178-.277	0.025	29	190
dissepiment spacing	0.573	.421-.676	0.070	29	231
dissepiment width	0.174	.148-.191	0.019	4	40
aperture spacing along row	0.215	.170-.248	0.018	28	248
aperture spacing across branch	0.213	.170-.252	0.019	28	252
aperture spacing between branches	0.237	.224-.256	0.011	9	90
diameter of distal tube	0.072	.061-.090	0.006	28	246
chamber length parallel with base	0.214	.169-.247	0.018	28	248
maximum chamber width	0.101	.082-.138	0.011	28	249
maximum chamber depth	0.145	.116-.186	0.019	23	166
angle, reverse - transverse walls	52	35-60	5.1	5	47
angle, distal tube - branch midplane	18	11-27	4.5	2	15
diameter of keel nodes	0.043	.037-.051	0.004	8	67
spacing of keel nodes	0.175	.129-.223	0.025	13	98

NOTE: Linear measurements are in millimeters, and angles are measured in degrees.

differ in spacing of nodes along the frontal keel (Table 3). The primary differences between these two putative species as noted by Condra and Elias are in the robustness of the axial screw (axis very slender with whorls 2–2.5 mm apart in *A. fragilis* and axis quite robust with whorls 6.5–9 mm apart in *A. magnus*). Condra and Elias (1944, p. 154) reported a curved axial screw of *A. magnus* with a delicate proximal portion in which the core of the axis has a diameter <1 mm and whorl spacing of 5 mm, which is much closer to that of typical *A. fragilis* than typical *A. magnus.*

A. magnus was reported (Condra and Elias, 1944, p. 155) to have a "dotlike" structure in the center of some peristomes, which was interpreted as an aggregate of dark brown bodies. Many *Archimedes* specimens have dark-brown aggregates preserved in peristomes, which, as Condra and Elias inferred for *A. magnus,* likely represent original brown bodies (i.e., decayed remnants of polypides). However, the structures that drew their attention in *A. magnus* were the inverted funnel caps of secondary nanozooecia atop distal tubes.

Measurements.—See Table 4.

Material.—Locality 77–79, unit 5 (4 specimens), unit 10 (1 specimen), float (15 specimens); locality 77–82, unit 9 (2 specimens), unit 10 (1 specimen), unit 27 (6 specimens), float (1 specimen); locality 77–83, unit 2 (1 specimen), unit 5 (5 specimens), unit 6 (4 specimens), unit 7 (6 specimens); NGS 243a, paratype of *A. magnus* Condra and Elias, 1944; NGS 254, 255, hypotypes of Pitkin *A. proutanus* in Condra and Elias, 1944.

REFERENCES CITED

Condra, G. E.; and Elias, M. K., 1944, Study and revision of *Archimedes* (Hall): Geological Society of America Special Papers 53, 243 p.

Easton, W. H.; and Duncan, H., 1953, *Archimedes* and its genotype: Journal of Paleontology, v. 27, p. 737–741.

Elias, M. K., 1957, Late Mississippian fauna from the Redoak Hollow Formation of southern Oklahoma: Journal of Paleontology, v. 31, p. 370–427.

Horowitz, A. S., 1977, Late Mississippian and Early Pennsylvanian bryozoan faunas of Arkansas and Oklahoma: a review, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, p. 101–105.

McKinney, F. K., 1979, Some paleoenvironments of the coiled fenestrate bryozoan *Archimedes*, *in* Abbott, M. B.; and Larwood, G. P. (eds.), Advances in Bryozoology: Academic Press, London, p. 321–335.

McKinney, F. K.; and Gault, H. W., 1980, Paleoenvironment of Late Mississippian fenestrate bryozoans, eastern United States: Lethaia, v. 13, p. 127–146.

Owen, D. D., 1838, Report of a geological reconnaissance of the state of Indiana; made in the year 1837: Bolton and Livingston, Indianapolis, 34 p.

Ross, J. R. P., 1978, Biogeography of Permian ectoproct Bryozoa: Palaeontology, v. 21, p. 341–356.

_____1981, Biogeography of Carboniferous ectoproct Bryozoa: Palaeontology, v. 24, p. 313–341.

Shumard, B. F., 1853, Descriptions of the species of Carboniferous and Cretaceous fossils collected, *in* U.S. War Department, Exploration of the Red River of Louisiana, in the year 1852; by Randolph B. Marcy . . . with reports on the natural history of the country: 32nd Congress, 2nd Session, Senate Document 54, Appendix E, Palaeontology, p. 197–211.

Tehan, R. E.; and Warmath, A. T., 1977, Lime-mud mounds of the Pitkin Formation (Chesterian), northwestern Arkansas, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, p. 49–54.

Ulrich, E. O., 1890, Palaeozoic Bryozoa: Illinois Geological Survey, v. 8, p. 283–688.

Webb, G. E., 1987, Late Mississippian thrombolite bioherms from the Pitkin Formation of northern Arkansas: Geological Society of America Bulletin, v. 99, p. 686–698.

Whisonant, R. C.; and Scolaro, R. J., 1979, Shore-related clastic and carbonate depositional environments in the Mississippian Bluefield Formation in eastern West Virginia: Geological Society of America Abstracts with Programs, v. 11, p. 217–218.

Wulff, J. L., 1990, Biostratinomic utility of *Archimedes* in environmental interpretation: PALAIOS, v. 5, p. 160–166.

APPENDIX: Localities

1977–79.—Pitkin Quarry, east of U.S. 71, 1.3 km south of Arkansas Highway 170, West Fork, Arkansas. NW¼SW¼ sec. 4, T. 14 N., R. 30 W., West Fork Quadrangle, Washington County. Unit 3, skeletal grainstone, (1.7–2.7 m) above base of section; unit 5, skeletal ooid grainstone, (3.5–7.0 m) above base of section; unit 10, intraclastic packstone, (11.3–12.8 m) above base of section; algal-bryozoan boundstone mound, (8.2–10.7 m) above base of section.

1977–82.—Along Arkansas Highway 263, 2.7–3.5 km south of Arkansas Highway 66, southwest of Timbo, Arkansas. SE¼NW¼ and NE¼SW¼ sec. 13, T. 14 N., R. 13 W., Fox Quadrangle, Stone County, Arkansas. Unit 9, skeletal packstone, (3.8–4.9 m) above base of formation; unit 10, skeletal wackestone, (4.9–8.7 m) above base of formation; unit 27, skeletal packstone, (17.7–21.7 m) above base of formation.

1977–83.—West side of Arkansas Highway 25, 1.4 km south of Arkansas Highway 14, Locust Grove, Arkansas. NE¼SE¼ sec. 32, T. 13 N., R. 7 W., Concord Quadrangle, Independence County, Arkansas. Unit 1, lime mudstone, (0–1.0 m) above base of section; unit 5, oncolitic wackestone, (4.6–5.3 m) above base of section; unit 6, skeletal packstone, (5.3–5.9 m) above base of section.

PLATE 1

Figures 1–6.—*Archimedes lunatus* Condra and Elias. **1**, exterior of robust axial screw, U.S. National Museum of Natural History (USNM) 450586, locality 1977–82, unit 27, ×2. **2**, exterior of more delicate axial screw, USNM 450587, locality 1977–83, unit 7, ×2. **3**, section transverse to branches within flange, showing "trident" spines, USNM 450588, locality 1977–79, unit 5, ×37.5. **4**, section longitudinal with respect to branch in flange, showing apertural spines and macrostyles in massive skeleton, USNM 450589, locality 1977–79, unit 5, ×37.5. **5**, section across axis, tangential with respect to branch system and passing below zooecial chambers in upper half of figure, USNM 450590, locality 1977–79, unit 5, ×37.5. **6**, section across axis, tangential with respect to branch system and passing through chambers along left border and above chambers in middle and right of figure, USNM 450591, locality 1977–79, unit 5, ×37.5.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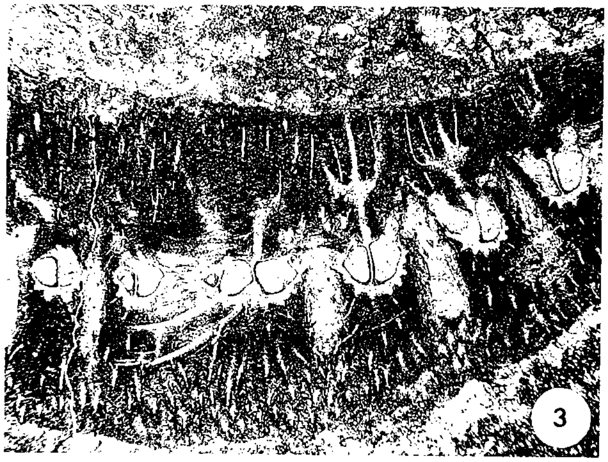
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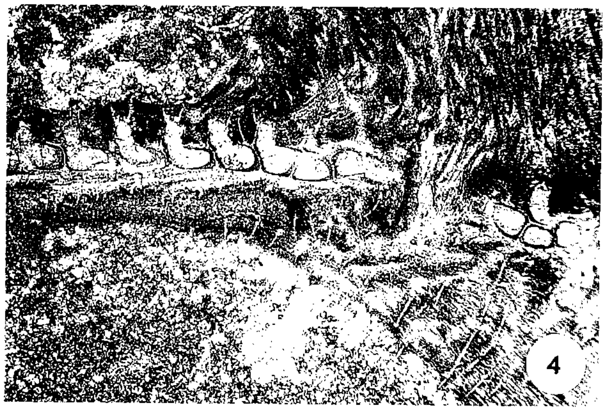
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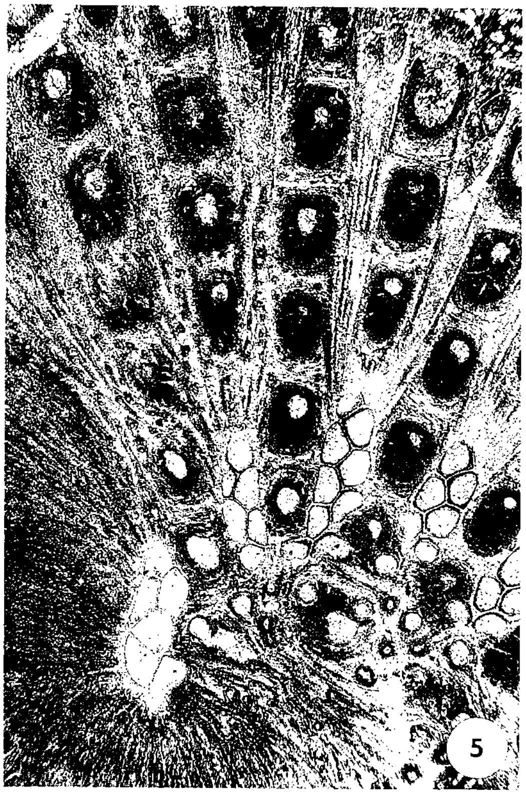
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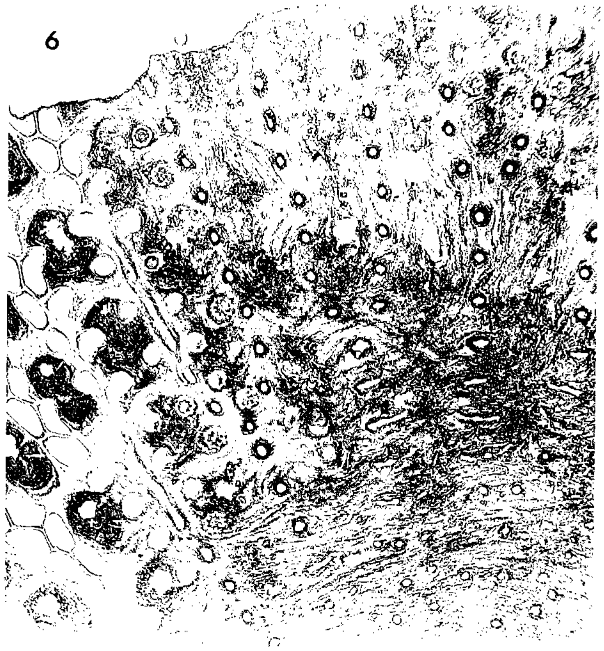
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PLATE 2

Figures 1–7.—*Archimedes lunatus* Condra and Elias. **1**, longitudinal section, USNM 450592, locality 1977–83, unit 6, ×37.5. **2**, transverse section, USNM 450593, locality 1977–83, unit 1, ×37.5. **3**, section across axis, tangential with respect to branch system but largely passing above branches, through region of "trident" spines, USNM 450590, ×37.5. **4**, shallow tangential section, USNM 450594, locality 1977–83, float, ×37.5. **5**, longitudinal section along center of axial screw, through giant zooecia of inner edge of branch system, USNM 450595, locality 1977–79, unit 5, ×37.5. **6**, deep tangential section, USNM 450594, ×37.5. **7**, tangential section through stellate apertures (left and top), spines above stellate apertures (right), and keel spines (left margin and right side), USNM 450596, locality 1977–79, unit 5, ×125.

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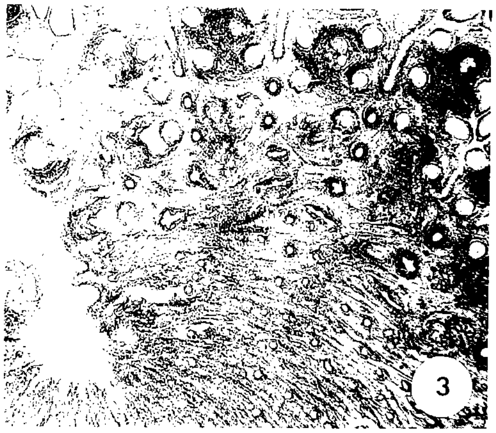

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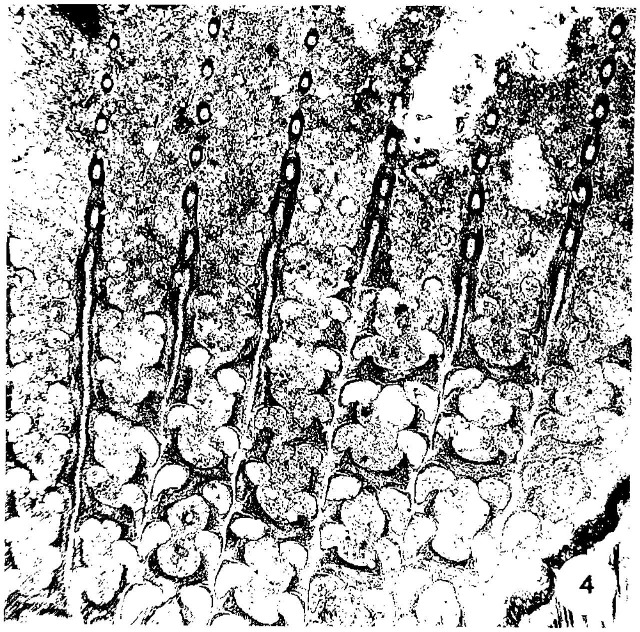

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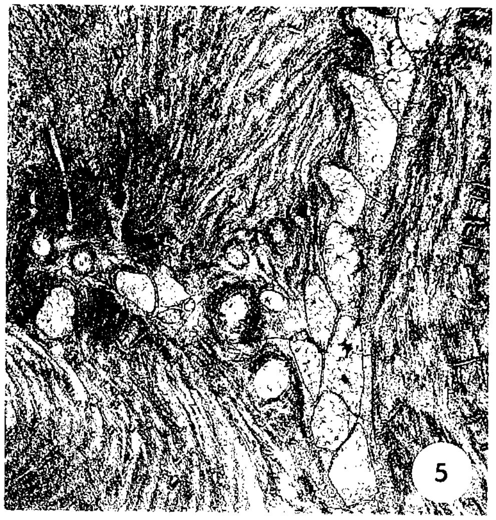

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PLATE 3

Figures 1–8.—*Archimedes magnus* Condra and Elias. **1–4**, axial screws of varying degrees of robustness, all ×2. **1**, USNM 450597, locality 1977–79, float; **2**, USNM 450598, 1977–83, unit 6; **3**, USNM 450599, 1977–79, unit 5; **4**, USNM 450600, 1977–79, float. **5**, longitudinal section along center of axial screw, through giant zooecia of inner edge of branch system, USNM 450601, locality 1977–79, unit 5, ×37.5. **6**, section across axis, through massive skeleton with macrostyles extending from reverse side of branches, USNM 450602, locality 1977–79, float, ×37.5. **7**, section transverse to branches within flange, through massive skeleton deposited above frontal side of branches and containing highly elongate, hirsute keel spines, USNM 450603, locality 1977–83, unit 7, ×37.5. **8**, section longitudinal with respect to branch in flange, showing macrostyles extending beyond general skeletal surface, USNM 450602, ×37.5.


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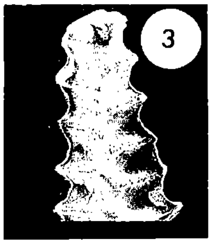
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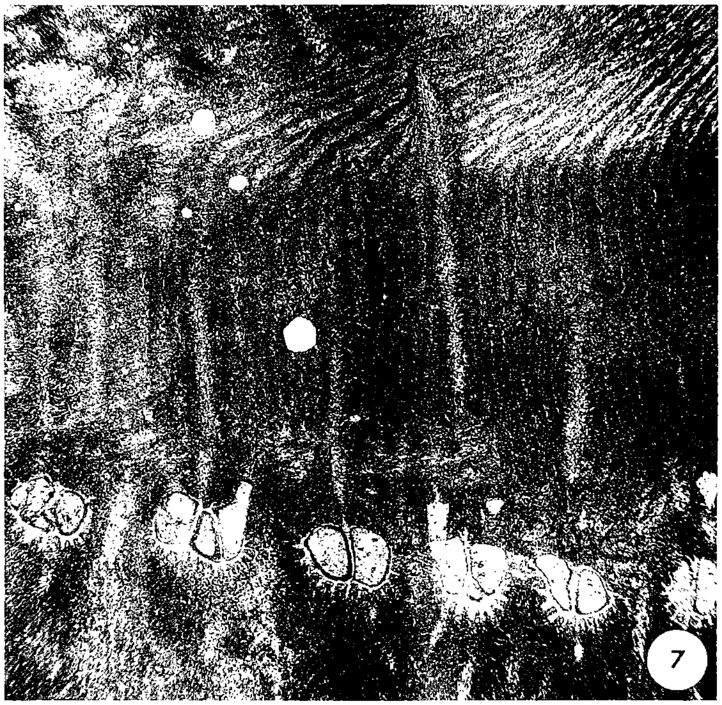
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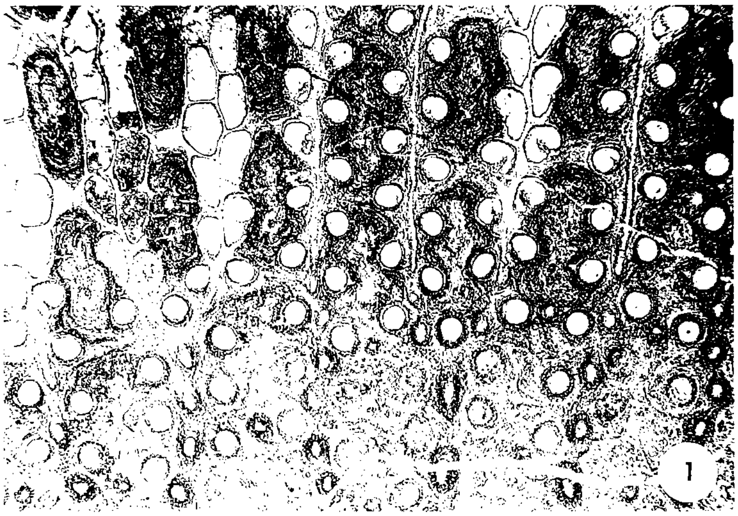
PLATE 4

Figures 1–6.—*Archimedes magnus* Condra and Elias. **1**, section through axis, tangential along frontal side of branches, USNM 450602, ×37.5. **2**, transverse section of branch within axial screw, with zooecium on left side transformed to secondary nanozooecium by funnel cap, USNM 450603, ×125. **3**, section through axis, tangential at depth of zooecial chambers with conspicuous inferior hemisepta, USNM 450604, locality 1977–83, unit 6, ×37.5. **4**, section through axis, cutting across extended axial spines above frontal side of branches, USNM 450602, ×37.5. **5**, section through flange of axial screw, tangential to branch system and cutting across robust macrostyles on reverse side of branches, USNM 450603, ×37.5. **6**, section through flange of axial screw, cutting elongate, "hirsute" spine extending from keel, above frontal side of branches; spine gives rise prolifically to microstyles, USNM 450602, ×125.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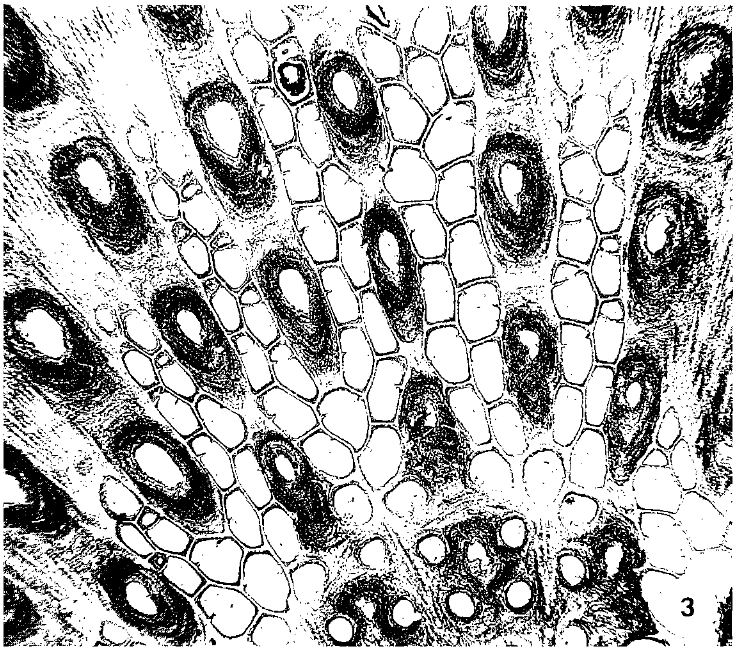
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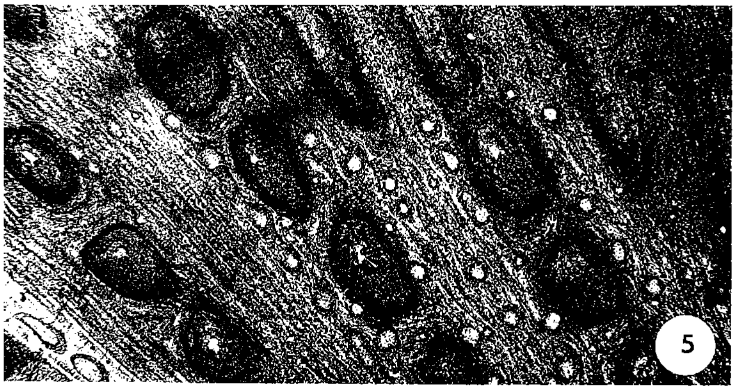
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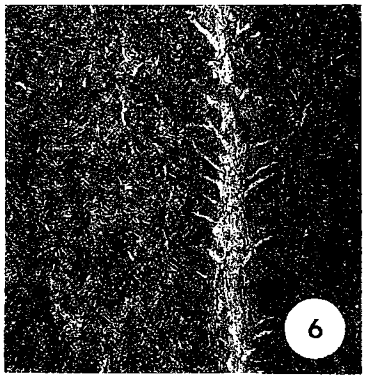
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Oklahoma Geological Survey Circular 94, 1992

Paleopedology and Conodont Biostratigraphy of the Mississippian–Pennsylvanian Boundary Interval, Type Grove Church Shale Area, Southern Illinois

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ABSTRACT.—Subaerial exposure features, mineralogy, paleosol horizons, and conodonts in a core drilled near the type section of the uppermost Mississippian Grove Church Shale indicate that the formation is bounded by disconformities and contains an intraformational disconformity. The lower Grove Church records a transgression, represented by prodeltaic mudstone, and subsequent regression and subaerial exposure, represented by a paleosol profile. The upper Grove Church is dominated by interbedded shale and limestone and was deposited in a low-energy, near-shore marine environment. The contact with the overlying basal Pennsylvanian Wayside Sandstone Member, Caseyville Formation, is abrupt. Distinct mineralogic differences at the sharp contact indicate an unconformity.

Differences in conodont assemblages between the Kinkaid, Grove Church, and Wayside units suggest hiatuses at both the top and bottom of the Grove Church rather than paleoecological changes. Conodonts recovered from the Grove Church probably represent the late Chesterian *Adetognathus unicornis* Biozone. The Grove Church may also contain the *Rhachistognathus muricatus* Biozone (latest Chesterian), but the name-bearer is absent. *A. unicornis* and *Kladognathus* cf. *K. tenuis* dominate the fauna, with element abundances >50/kg in some samples. *Cavusgnathus naviculus*, *C. unicornis*, *Hindeodus cristula*, *H. minutus*, and *Vogelgnathus* n. sp. are also present in low to moderate numbers. The lower shale sequence of the overlying Wayside Sandstone contains rare *Lochriea commutata* and *Rhachistognathus* cf. *R. websteri*, which suggest an early Morrowan age. A morphotype of *A. lautus*, previously identified as *A. unicornis*, occurs in moderate abundances throughout the Wayside, but its biostratigraphic utility is unknown.

INTRODUCTION

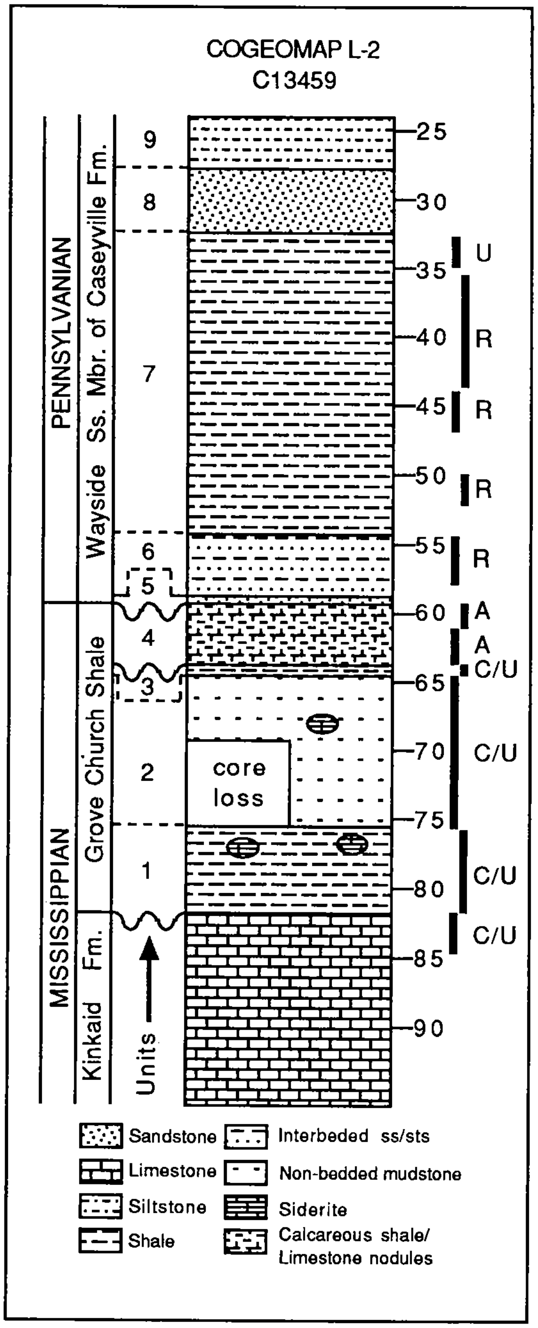
For more than 125 years, physical stratigraphers have recognized a disconformity between Mississippian and Pennsylvanian strata in the Illinois basin, based on recognition of a paleo-topographic surface at the boundary and truncation of pre-Pennsylvanian strata toward basin flanks. Biostratigraphic study of the boundary interval in southern Illinois, where lowest Pennsylvanian overlie uppermost Mississippian strata, has not corroborated the lithostratigraphic interpretation, primarily because basal Pennsylvanian strata are sparsely fossiliferous. Recently, conodont biostratigraphic data have been presented indicating the absence of a boundary unconformity. Thus, the question of whether the boundary in the southernmost part of the basin is conformable or unconformable remains controversial and inadequately documented.

The availability of a core, drilled recently through the Mississippian–Pennsylvanian boundary and close to the type section of the Grove Church Shale (youngest formation in the type Mississippian), allowed detailed analysis of the boundary interval, assessment of conformable or unconformable contacts, and determination of hiatuses, if any. The core from COGEOMAP L-2 (locality 1; see Appendix) was drilled ~260 m east–northeast of the Grove Church type section (locality 2), northwestern Johnson County. The core penetrated the Pennsylvanian Wayside Sandstone Member of the Caseyville Formation, and the Mississippian Grove Church Shale, Kinkaid Limestone, and upper Degonia Formation. After we studied the upper part of the core (Text-fig. 1), additional rock was collected from the type section (Text-fig. 2) to process for conodonts. At this exposure, the Wayside Sandstone is lithologically similar to rocks exposed at its type locality, ~1.3 km to the north–northeast, and match the description of the lower Wayside according to Lamar (1925).

Core descriptions, augmented by outcrop description, were used to distinguish boundaries between stratigraphic sequences from the uppermost Goreville Limestone Member, Kinkaid Limestone, to the top of the core. Petrographic and X-ray diffraction analyses provided additional data for our sedimentologic conclusions. Measurements are metric except for core depths, which were originally measured in feet.

Acknowledgments

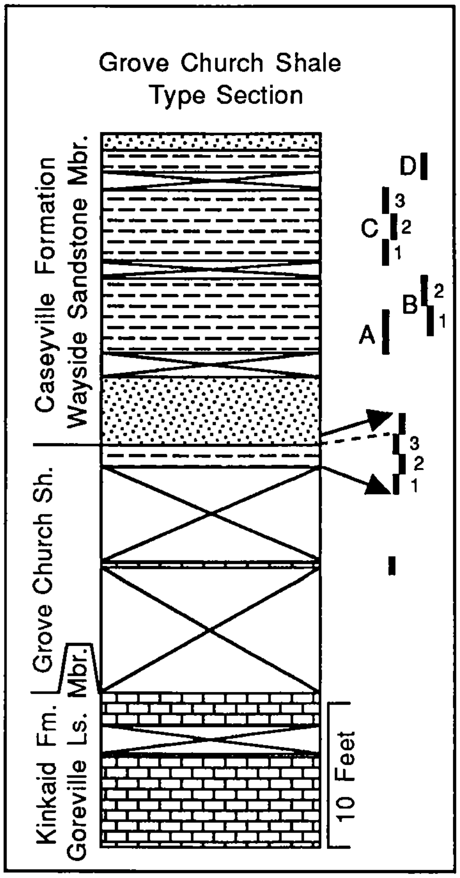
Work on this project was partially supported through a cooperative agreement with the U.S. Geological Survey under its Cooperative Geologic Mapping Program (COGEOMAP). We thank Drs. Leon R. Follmer and Randall E. Hughes of the Illinois State Geological Survey (ISGS) for their assistance in interpretation of soil structures and mineralogical analyses. We also wish to thank Joel Dexter of the ISGS for the photographs of the core samples, and Norm Peters and Rick Olson of the University of Illinois Center for Electron Microscopy, and Don Lowry of the ISGS for technical assistance in obtaining scanning electron micrographs of the conodonts. The final version of this paper was improved by the reviews of Walter Manger, University of Arkansas, and Frank Ettensohn, University of Kentucky.



Text-figure 1. Stratigraphic column of COGEOMAP L-2 core. Boundaries of stratigraphic sequences are on left side of column. Depth markers are in feet. Heavy vertical bars indicate sampling intervals. Relative number of conodonts recovered per sample interval indicated by R (rare), U (uncommon), C (common), and A (abundant).

PREVIOUS WORK

Worthen (1866) recognized the Pennsylvanian subcrop in the basin to be as old as the Ordovician St. Peter Sandstone and as young as the Mississippian Kinkaid Limestone. Lamar (1925) mapped the Carbondale 15′ Quadrangle, which encompasses the Grove Church type locality of Swann (1963) and



Text-figure 2. Stratigraphic column of the type section of the Grove Church Shale. Heavy vertical bars indicate sampling intervals.

the Wayside Sandstone type area. Lamar interpreted the erosional boundary surface to consist of buried valleys and divides, confirmed by the later mapping of Siever (1951), Wanless (1955), and Bristol and Howard (1971). Ostrom and Potter (1961) recognized a weathering profile on the Cave Hill Shale Member, Kinkaid Limestone, at the Mississippian–Pennsylvanian boundary in Crittenden County, Kentucky, where the contact is an erosional unconformity. This view remained unchallenged until Ethridge and others (1973) and Ethridge and Bailey (1976) suggested, with little substantiating data, that the systemic boundary may be conformable in southern Illinois.

Foraminifers, ostracodes, and conodonts were collected and described initially from the uppermost Kinkaid Limestone (= Grove Church) at the type locality of the Grove Church (Cooper, 1947). Rexroad and Burton (1961) and Collinson and others (1962,1971) documented the conodont *Streptognathodus unicornis* Assemblage Zone (= *Adetognathus unicornis* Biozone) as the highest zone of the standard Mississippian section in the Illinois basin, but postulated that the Chesterian Series was truncated by an erosional unconformity and was incomplete. No biostratigraphic evidence for this hiatus was included in those studies. Little attention was given to basal Pennsylvanian conodonts until preliminary reports from the Wayside Sandstone Member by Rexroad and Merrill (1979) and Merrill and Rexroad (1981), followed with supporting data by Rexroad and Merrill (1985). They concluded that the conodont *Adetognathus* and lithogenetic data demonstrated both an evolutionary and depositional continuum across the Mississippian–Pennsylvanian boundary in the southern Illinois basin. In contrast, Jennings and Fraunfelter (1986), in a study of macrofossils from the Grove Church and the Wayside, concluded that a "major" hiatus is present at this boundary.

STRATIGRAPHY, PALEOSOLS, AND DEPOSITIONAL HISTORY

Goreville Limestone Member

The Goreville Limestone Member, Kinkaid Limestone, records the last major Mississippian transgression, and is the highest, relatively thick, pre-Pennsylvanian carbonate deposit in the Illinois basin (Swann, 1963). The Goreville consists of thick-bedded, crinoid-dominated, sparite-cemented calcarenites and locally interbedded fossiliferous calcilutites that were deposited in a near wave-base, subtidal, marine environment. The top of the unit contains characteristics attributable to a pedologic origin, strongly suggesting that a period of subaerial exposure occurred before deposition of overlying Grove Church. This unconformity was first suggested by Rexroad and Burton (1961) based on conodonts.

A polished core slab reveals an obliquely bisected root cast near the top of the member, infilled with Goreville lithoclasts and shale of the overlying Grove Church (Text-fig. 3A). Smaller shale infillings of the Grove Church are also common near the top and are attributed to either small root casts or fracture infillings. The root trace provides good evidence that the Goreville Limestone was exposed subaerially with subsequent plant growth (Retallack and others, 1984; Retallack, 1988). The top 1 cm of the member is a breccia of Goreville Limestone clasts surrounded by fractures and voids filled with Grove Church Shale. The contact between the breccia zone and the underlying calcarenite varies from abruptly to rapidly (<0.5 cm) diffuse, and the surface is irregularly undulated. The breccia zone was caused by root penetration that fractured the limestone substrate (Klappa, 1980; Ettensohn and others, 1988), and the fractures and voids were infilled during the subsequent Grove Church transgression.

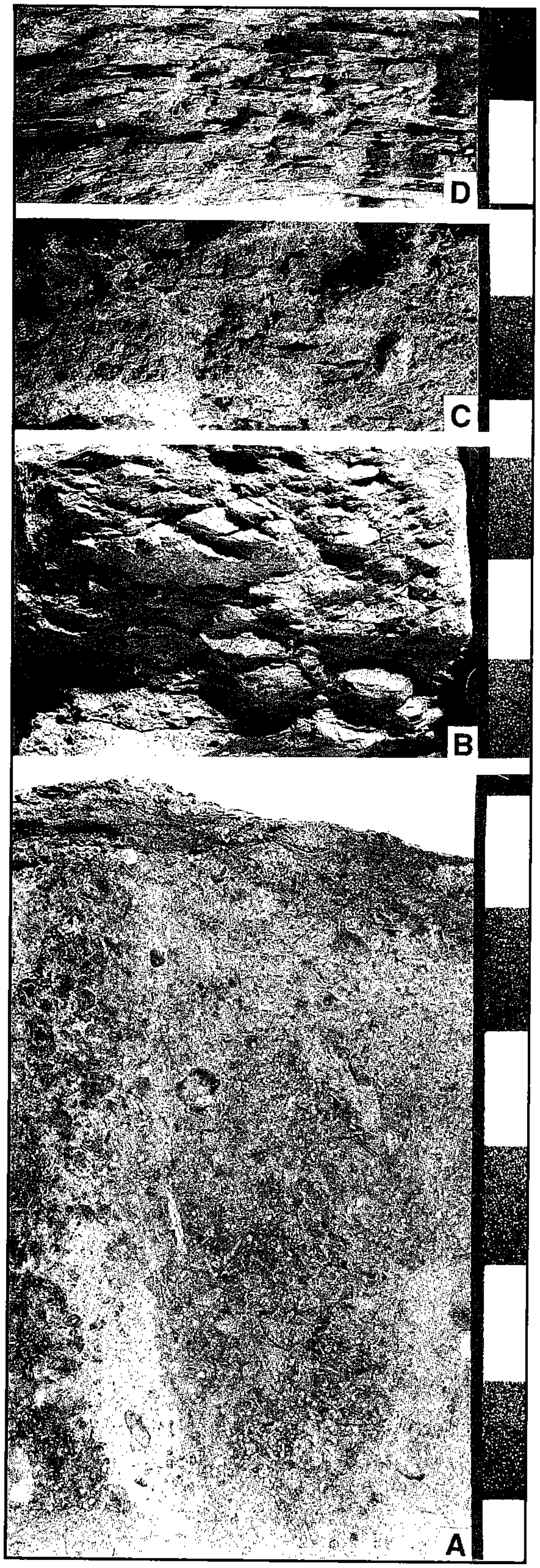
Grove Church Shale

The lower three of the four lithologic units in the Grove Church in the core (Text-fig. 1) comprise a stratigraphic sequence that records the first Grove Church transgression. The basal unit (1), from 75.7 to 81.8 ft, is a greenish-gray, weakly calcareous, massive shale that contains siderite nodules, common, but poorly preserved, casts of brachiopods and fenestrate bryozoans, and rare, flattened productid brachiopods. This unit is interpreted as a nearshore marine deposit, proximal to either an aggrading coastline or delta. The subsequent primary depositional record of the sequence has been extensively modified by pedogenesis. After deposition of units 2 and 3 (and possibly more for which there is no record), the seas regressed, exposing the strata, and allowing weathering and pedogenesis. We recognize the succeeding units, a mudstone (unit 2) and a thin shale (unit 3) as paleosol horizons. The soil profile overlies the unaltered bedrock (paleosol horizon R) of unit 1, but originally developed on, and modified, units 2 and 3.

Unit 2, a gray, noncalcareous mudstone, from 64.5 to 75.7 ft, includes a lost core interval from 69.5 to 75.7 ft. At the macroscale, the lithology appears uniform and massive, but at the mesoscale (mm), pedogenic features can be recognized. The silty clay of this unit is gray (N3.5–4/0 at top, grading down to N4–5/0) to greenish-gray (5GY6/1). The high silt content gives broken surfaces a smooth, powdery appearance. The fabric (soil structure) at the top is platy and weak slickensides are present. A siderite nodule (= a glaebule) occurs at 64.8–64.9 ft. Weak gray (N4/0) argillan (cutan) mottles are common from 65.0 to 69.2 ft, where the fabric is dominated by rounded, compressed, unorganized, pelletal aggregates (= peds) of silty clay in the 0.5–2.0 mm size range. Mean aggregate size increases with depth and the shape becomes rhomboid (Text-fig. 3B,C). Interaggregate porosity is very low and slickensided surfaces appear to increase with depth. Abundance of subordinate platy and angular forms increases with depth (67.2–69.2 ft).

We interpret this unit to be equivalent to a modern illuviation zone or the B horizon. Ped size steadily increases with depth, as in recent soil profiles (Follmer and others, 1979). Slickenside formation is caused by either shrinking/swelling of clay during dry/wet periods, or burial compression (Retallack, 1988). Slickensides in the upper part are not common, but apparently extend across the core (Text-fig. 3C), and are attributed to the former cause. Slickensides associated with the larger peds near the bottom are randomly oriented and of short length. We interpret these, as well as the rhomboid shapes, to be the result of the burial compression. We designate this paleosol horizon as Bg (gley) because of the pedogenic features described above and the dominantly neutral color. During pedogenesis, gleying is the result of stagnant water preventing oxidation and preserving the reducing state (Birkeland, 1984).

Unit 3, from 63.65 to 64.5 ft, is black (N3/0), very thin-bedded, weakly calcareous, silty, clay shale. Its fabric is strongly expressed as thin, platy (thin-bedded) forms with smooth, horizontal faces (Text-fig. 3D). Coal laminae and carbonized plant impressions of branches and leaves are common. The contact with the Bg horizon is diffuse. We interpret this unit to be equivalent to a modern elluviation zone. Differentiation between modern A and O horizons is based on ratio of organic and mineral contents. Since this ratio is unattainable in paleosols due to modification by compression and diagenesis, we



Text-figure 3 (left). Megascopic views of portions of the core. Footage is depth. Scale in cm. *A*—Goreville Limestone Member, Kinkaid Formation. Top at 81.8 ft, top of unit 1. Polished core slab. Obliquely bisected root cast in lower left, infilled with Goreville lithoclasts and shale from the overlying Grove Church. A smaller root cast in upper left. Micro-breccia at top composed of Goreville clasts and pyrite in Grove Church Shale matrix. Disconformity at top. *B*—Grove Church Shale. Top at 67.3 ft, unit 2. Split core. Sample composed of peds up to 0.5 cm in diameter. Rhomboid shape and numerous slickensided surfaces of short length formed during burial compression. *C*—Grove Church Shale. Top at 65.9 ft, unit 2. Split core. Sample composed of peds, <1 mm in diameter, in paleosol horizon Bg. Oblique trace, from upper left to lower right, is apparent slickensided surface that formed during moisture content fluctuation. *D*—Grove Church Shale. Top at 63.95 ft, unit 3. Split core. Thin, platy bedding of paleosol horizon AO. Coal laminae and carbonized leaf and branch impressions are common on bedding surfaces.

designate this unit as an AO horizon. The platy structure is relict bedding inherited from the deposition of organic matter and has been enhanced by compaction. The soil profile represents the regressive phase of the first Grove Church transgression. The top of the AO horizon is sharp, well defined, and indicates a sudden transition from subaerial pedogenesis to deposition of the succeeding marine sequence.

The second Grove Church stratigraphic sequence (unit 4), from 59.3 to 63.65 ft, consists of interbedded calcareous shale and limestone that contains abundant fossils. Limestone occurs as nodules, either isolated or concentrated in poorly defined beds of grain-supported to mud-supported biocalcarenite with a calcisiltite matrix. Pelmatozoan fragments are the dominant bioclasts, but brachiopods, bryozoans, endothyroid foraminifers, fish debris, ostracodes, rugose corals, and trilobites are also present, suggesting normal salinity. The fragmented, but diverse, fossil assemblage and lack of winnowing suggest deposition in a nearshore marine environment with low energy and shallow to moderate water depths during the second Grove Church transgression. Pedologic structures and paleosol horizons have not been identified within or at the top of this unit. The contact between the uppermost Grove Church shale (unit 4) and the basal Wayside sandstone (unit 5) is sharp and has a nonplanar surface. Furthermore, the mineralogical assemblages of shales from the Grove Church and Wayside contrast sharply. X-ray diffraction analysis of the shale at the top of unit 4 indicates that the illite to illite/smectite (I:I/S) ratio is roughly 1:1; siderite is present, feldspar is absent, and chlorite is either absent or occurs in a trace amount. In contrast, the lowest Wayside shale (unit 6) exhibits a I:I/S ratio of roughly 9:1; siderite is absent, feldspar is present, and chlorite is abundant. The Grove Church mineralogical assemblage is consistent with sediments derived from a chemically weathered soil, whereas the Wayside is more indicative of detritus from an unweathered source (i.e., mechanical origin) (Hughes and others, 1987). Ostrom and Potter (1961) presented similar mineralogical trends in their study of a Mississippian–Pennsylvanian boundary paleosol on the Cave Hill Shale Member in Crittenden County, Kentucky. On the basis of the abrupt change in the lithology at the contact, the abrupt change of shale mineralogies within the boundary interval, and a sharp, well-defined, nonplanar contact, we interpret the boundary interval to be a succession of two different depositional environments separated by a hiatus.

Wayside Sandstone Member

The basal unit (5), from 58.95 to 59.3 ft, of the lowest sequence in the core (Text-fig. 1) in the Wayside Sandstone Member, Caseyville Formation, is a thin, very fine-grained, weakly calcareous, quartzose sandstone. It is sparsely fossiliferous; while Rexroad and Merrill (1985) and Jennings and Fraunfelter (1986) reported fossil occurrences, we found only a few broken conodont elements. The reported marine fossils indicate deposition in a littoral environment. The sandstone is overlain by noncalcareous, thin-bedded to laminated, interbedded sandstone, siltstone, and shale with abundant coal laminae (unit 6) from 54.25 to 58.95 ft. Siltstone clasts are scattered throughout the sandstones. Lithologic contacts are often abrupt and loadcasts are common; however, evidence of subaerial exposure, pedogenesis, or erosion is absent. The abrupt lithologic changes, clasts, and loadcasts suggest episodic deposition in either an interdistributary bay or a lagoon. A thick, dark-gray, sparsely to moderately fossiliferous, noncalcareous shale (unit 7), from 32.8 to 54.25 ft, follows. Lack of plant fossils, silt, and larger grains indicates deposition in a low-energy, shallow, protected lagoon with restricted access to open seas without terrestrial influence. The contact with unit 8 is abrupt.

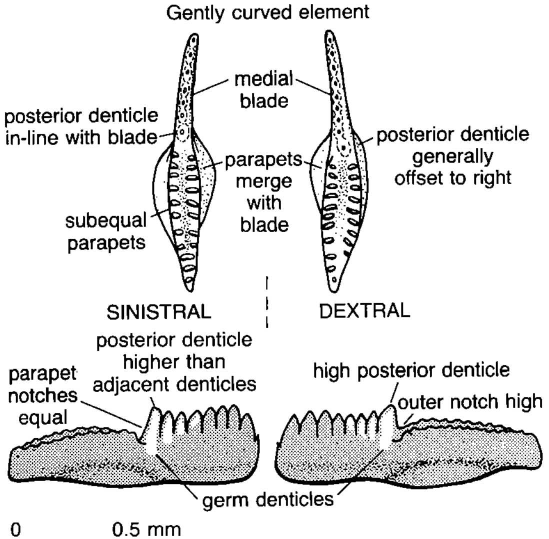
The second Wayside stratigraphic sequence in the core is incomplete and consists of a basal unit (unit 8), from 27.9 to 32.8 ft, a medium-grained quartzose sandstone containing siltstone interbeds, and a quartzose siltstone (unit 9), from 24.0 to 27.9 ft, that contains common carbonaceous plant fragments. We have not studied this sequence in detail.

CONODONT BIOSTRATIGRAPHY

Grove Church Fauna

Conodonts are well represented throughout the 22.2 ft of Grove Church Shale in the core (Text-fig. 1). Unit 4 yielded an abundant fauna (220 elements/kg) with no significant differences from that described previously (Cooper, 1947; Rexroad and Burton, 1961; Rexroad and Merrill, 1985). The fauna (Pl. 1) includes *Adetognathus unicornis* (Rexroad and Burton), *Kladognathus* cf. *K. tenuis* (Branson and Mehl), *Cavusgnathus naviculus* Youngquist and Miller, *Vogelgnathus* n. sp., *Hindeodus minutus* (Ellison), *H. cristula* (Youngquist and Miller), and *C. unicornis* Youngquist and Miller, in order of decreasing relative abundance. *Adetognathus unicornis* represents about a third of the fauna and is the key species. *A. unicornis* is somewhat atypical of the genus as the blade is central to subcentral on both dextral and sinistral Pa elements (Text-fig. 4). The Pa elements of typical Pennsylvanian *Adetognathus* species have blades that merge with either the left or the right parapet of the platform.

Most of the non-Pa elements of the various species within the cavusgnathoid group are very similar and are generally grouped as "vicariously shared" elements, although some subtle differences probably exist among the species. Vicarious sharing of elements coupled with the paucity of non-Pa cavusgnathoid elements within the Grove Church



Text-figure 4. Key characteristics of *Adetognathus unicornis* (Rexroad and Burton).

make apparatus reconstruction of *A. unicornis* difficult.

Large elements of *Kladognathus* and the larger Pa and non-Pa (Pb, M, Sa, Sb, and Sc) elements of platform-type species dominate the Grove Church fauna. Small Pa and non-Pa elements were generally not recovered, except for a few Pa elements of *Vogelgnathus* n. sp. Severe under-representation of non-Pa elements makes apparatus reconstruction very difficult, thus obscuring phylogenetic and ontogenetic developments. Under-representation of non-Pa elements generally is attributed to winnowing action that reduces the smaller, more delicate elements to pieces that either cannot be identified or are not recovered. However, the general sedimentology, in both the core and outcrop, does not suggest significant winnowing action. Therefore, other factors, such as slow sedimentation rates, animal ingestion of conodonts, bioturbation, and percolation of acidic solutions through the sediment, might also contribute to a selective loss of smaller conodont elements. Additional well-known species, including *Vogelgnathus* n. sp. (previously illustrated by Norby and Rexroad, 1985, and as new genus n. sp. by Rexroad and Merrill, 1985), from the Grove Church are illustrated on Plate 1.

The conodont fauna of the first Grove Church sequence (units 1–3) differs from that of the underlying Kinkaid Limestone and reinforces the probability of a hiatus suggested by the exposure surface at the top of the Goreville Limestone Member. The principal faunal difference is the first appearance and abundance of *A. unicornis* in the Grove Church along with the first appearances of *Hindeodus minutus* and *Vogelgnathus* n. sp., whereas earlier *Kladognathus primus* Rexroad and *Lochriea commutata* (Branson and Mehl) are absent from the Grove Church. Several typical Chesterian species (e.g., *Cavusgnathus unicornis* and *H. cristula*) are greatly reduced in numbers.

Wayside Fauna

The Wayside conodont fauna is very distinct from the fauna in the underlying Grove Church. With the exception of *Lochriea commutata*, no other Chesterian species occurs in the Wayside. The important *Adetognathus unicornis* in the Grove Church is absent from the Wayside. *A. lautus* (Gunnell) morphotype A, a possible descendant of *A. unicornis* (reasons for this designation will be discussed below), first appears in the Wayside.

Only a few broken elements, including two Pa elements of *Adetognathus lautus* morphotype A, were recovered from Wayside intervals sampled in the core (Text-fig. 1). Our preliminary samples from the moderately well-exposed section immediately above the type Grove Church Shale (Text-fig. 2), yielded moderate abundances of conodonts (15/kg) except for the weakly calcareous, basal portion of the lowest sandstone (unit 5). This calcareous sandstone horizon produced a few broken elements, one of which appears to be an M element of *Lochriea commutata* (Pl. 2, Fig. 35). This species is moderately common in the Kinkaid but has not been reported from the Grove Church. It may be indigenous to the Wayside or, though unlikely, it could have been reworked from topographically higher Kinkaid strata.

In samples A through C3 from the outcrop equivalent of unit 7 (Text-fig. 2), the moderately abundant conodont fauna consists almost entirely of a highly morphologically variable cavusgnathoid plexus/species, which we have designated *Adetognathus lautus* morphotype A. In addition to *A. lautus* morphotype A, one Pa element each of *L. commutata* and *Rhachistognathus* cf. *R. websteri* Baesemann and Lane were recovered. Rexroad and Merrill (1985) also reported and illustrated several gnathodontoids that we did not recover in our samples.

Within *A. lautus* morphotype A, the Pa elements range from very small (0.6 mm) to medium-sized (1.1 mm). Most of the Pa elements have blades that merge either with the right side of the platform in dextral specimens or with the left side of the platform in sinistral specimens. A minor percentage of specimens (generally of smaller size) may have a more subcentral blade. It is a few of the latter specimens that superficially resemble *Adetognathus unicornis*.

The non-Pa elements of *A. lautus* morphotype A are abundant and show some size variation. All element positions are well represented (Table 1) permitting easy reconstruction of the apparatus. Most

TABLE 1. — ELEMENT ABUNDANCES OF *ADETOGNATHUS LAUTUS* MORPHOTYPE A FOR THREE SAMPLES WITHIN THE SHALE OF THE WAYSIDE SANDSTONE MEMBER AT THE TYPE GROVE CHURCH LOCALITY

		Elements				
Sample	Wt. (kg)	Pa	Pb	M	Sa	Sb-Sc
C3	11.85	33	24	19	7	39
C2	11.55	71	40	28	11	65
C1	10.95	84	47	65	19	109
Total	34.35	188	111	112	37	213

Numbers indicate an average Pa:Pb:M:Sa:Sb-Sc ratio of 2.0:1.2:1.2:0.4:2.4. This is not significantly less than the hypothetical whole animal ratio of 2.0:2.0:2.0:1.0:8.0, especially when breakage and non-recognition of very small fragments is considered. Unidentifiable fragments comprise 70% of these samples.

Pb elements (Pl. 2, Fig. 29), although invariably broken to some extent, show medium-sized discrete denticles on the anterior bar and alternating sets of smaller denticles on the posterior bar. This alternation on the posterior bar is more characteristic of Sc elements. M elements (Pl. 2, Figs. 30–32) are similar in morphology to the M elements (form species "*Neoprioniodus loxus*") of *Cavusgnathus unicornis*, but the anterior edge of the anticusp of the present elements have between 1 and 3 denticles which are generally absent on the latter species. The Sa, Sb, Sc elements (Pl. 2, Figs. 27,28,33,34) are typically cavusgnathoid, but the vertical height of the blade with denticles is much reduced over other cavusgnathoid elements. The Sc element has a variable anterolateral bar that suggests two possible forms of the Sc element. The size range and element representation strongly indicate that very little winnowing had occurred and observation supports the interpretation from lithologic and sedimentologic data that the environment is low energy and restricted.

Adetognathus in Boundary Strata

Rexroad and Merrill (1985) supported the somewhat controversial suggestion of Ethridge and others (1973) that the Mississippian–Pennsylvanian boundary is conformable in a small localized area of the Illinois basin. To support their contentions, Rexroad and Merrill (1985) presented sedimentologic and faunal data which suggested that environmental changes rather than a hiatus were responsible for changes in the conodont fauna.

Paramount to their case, Rexroad and Merrill (1985) reported a continuation of the late Chesterian species *Adetognathus unicornis* across the boundary into Pennsylvanian Wayside strata. We examined their figured and study collections of Grove Church and Wayside *Adetognathus* specimens to be certain that we had not overlooked any distinguishing morphologic criteria. A detailed examination of Rexroad and Merrill's Wayside specimens of "*A. unicornis*" indicates that they are not true *A. unicornis*. Only one of Rexroad and Merrill's (1985) figured specimens and a few specimens in their study collections approach *A. unicornis* in morphology, which parallels the findings in this study. In both studies, the specimens in question are generally small, but they still can be separated from *A. unicornis* on one or more morphologic criteria. We conclude that no specimens of *Adetognathus unicornis* were recovered from Wayside strata in either study. Therefore, the phylogenetic continuum of *A. unicornis*, as suggested by Rexroad and Merrill (1985), does not cross the Mississippian–Pennsylvanian boundary as defined in the southern Illinois area. If *A. unicornis* did give rise to *A. lautus* morphotype A, then the necessary morphologic changes would indicate at least a short interval of time. Thus, a hiatus at the Grove Church/Wayside contact is indicated, instead of only a change in environment as postulated by Rexroad and Merrill (1985).

We have categorized the Wayside *Adetognathus* plexus under *Adetognathus lautus* (Gunnell) *sensu lato*, as most of the forms are morphologically similar to *A. lautus*. To distinguish this plexus from other representatives of *A. lautus*, we have labeled these specimens as *A. lautus* morphotype A, which is a broader use of the informal morphotype notation. We tried several schemes to group and classify the Pa elements of this plexus and arrived at the same conclusion. These forms represent one species of *Adetognathus* that is highly variable in its outward morphology. Whether these forms should be placed entirely under one species is open to debate, but no logical basis exists for separating specimens with slight morphologic differences.

The Pa elements within the Wayside differ from most later forms of *A. lautus* in that the anterior end of the inner platform parapet generally ends near or just anterior to the posterior-most blade denticle. In the few larger specimens in this study, the parapet did extend farther anterior of the posterior-most denticle which is more similar to typical *A. lautus*. Therefore, the positional termination of the inner parapet may be related to ontogeny.

The Wayside specimens show some similarities to other Morrowan examples of *A. lautus*, but they also show differences. Therefore we would not place all Morrowan specimens of *A. lautus* under this morphotype. This morphotype and other early Pennsylvanian representatives of *A. lautus* may or may not be conspecific with the type that occurs in

Missourian strata (Rexroad and Merrill, 1985) and new species designations, at least for this morphotype, may be warranted after further study.

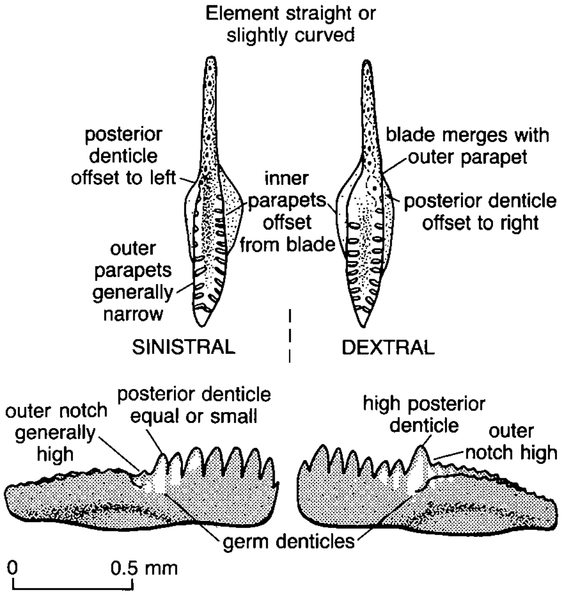
To facilitate their usage near the Mississippian–Pennsylvanian boundary, key distinguishing features of Pa elements of *A. unicornis* and *A. lautus* morphotype A are shown on Text-figures 4 and 5. General taxonomic distinctions among species of *Cavusgnathus, A. unicornis,* and *A. lautus* are discussed by Lane (1967,1968). Mature Pa elements of *A. unicornis* generally exhibit curvature of the element in upper view as compared with straight or slightly curved elements of *A. lautus* morphotype A. The blade is generally aligned with the platform trough on Pa elements of *A. unicornis,* but is offset in *A. lautus* morphotype A. The blade may occupy a more medial position on smaller elements of *A. lautus,* then apparently shifts toward the outer parapet during ontogeny. The posterior-most blade denticle is moderately large and in-line with the trough on sinistral Pa elements of *A. unicornis,* but is smaller and offset to the left on *A. lautus.* The parapets are nearly equal in width on *A. unicornis,* but the outer parapet is narrower than the inner parapet on both sinistral and dextral elements of *A. lautus.* Additional germ denticles are often present posterior to the posterior-most denticle in *A. lautus,* but are generally absent posterior to the blade in *A. unicornis.*

Age

Grove Church Shale

The presence of *Adetognathus unicornis* indicates that the Grove Church lies somewhere within the *A. unicornis* and *R. muricatus* Biozones (Text-fig. 6). Lack of *R. muricatus* in the Illinois basin generally has been held to reflect removal (or non-deposition) of latest Mississippian strata (i.e., *R. muricatus* Biozone) by the Mississippian–Pennsylvanian unconformity. Unfortunately, rhachistognathids generally have been found only in moderate- to high-energy shoaling environments (Krumhardt and Harris, 1990; Morrow and Webster, 1990), which are notably lacking in any boundary strata in the Illinois basin. Therefore, strata representing at least part of the *R. muricatus* Biozone could be present in the Grove Church without the name-bearer's presence.

No unconformity or time break was recognized previously between the Grove Church and the underlying Kinkaid, thus it was assumed that the *A. unicornis* Biozone succeeded the *Kladognathus–Cavusgnathus naviculus* Biozone (Collinson and others, 1962,1971) without a time break. Our data suggest that the Kinkaid–Grove Church disconformity and possibly the Grove Church intraformational disconformity could represent parts of either the *A. unicornis* or *Kladognathus–Cavusgnathus naviculus* Biozone.

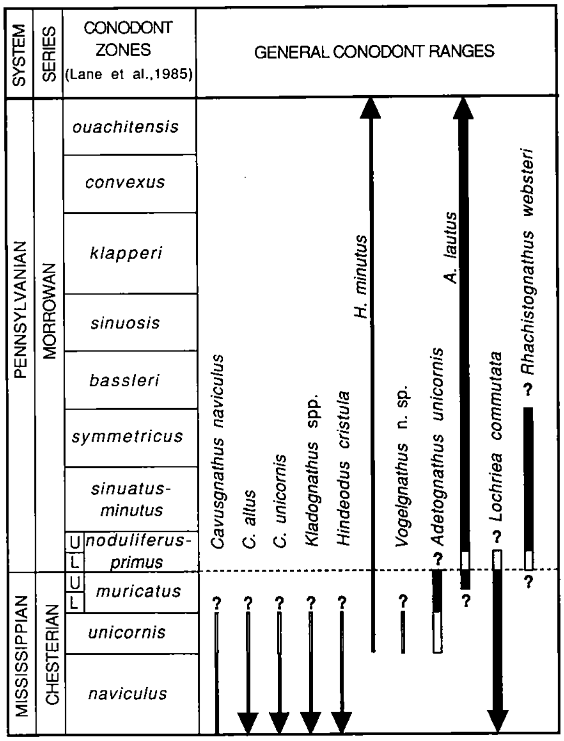


Text-figure 5. Key characteristics of *Adetognathus lautus* (Gunnell) morphotype A.

Wayside Sandstone Member

The biostratigraphic utility of *A. lautus* morphotype A is unknown; however, *A. lautus* has been reported from Mississippian rocks in some sections (Baesemann and Lane, 1985). The occurrence of the one small Pa element (Pl. 2, Fig. 37) of *Rhachistognathus* cf. *R. websteri* Baesemann and Lane (1985) in sample C3 from locality 2 indicates an early Morrowan age, if indigenous (Text-fig. 6). The known range of *R. websteri* is from the *Declinognathodus noduliferus–Rhachistognathus primus* Biozone through the *Neognathodus symmetricus* Biozone (Baesemann and Lane, 1985). *R. websteri* is a newly described species and, after additional study, its range could be extended upward, or less likely, downward. This small specimen also could be a variant of the shorter-ranging *R. primus* (Dunn). Our specimen shows close similarities to *R. primus* as illustrated by Lane and Straka (1974, fig. 35.21) from the Target Lentil, Springer Formation, Oklahoma. Neither of these rhachistognathids has been reported from the Mississippian. The possibility of reworking also exists; however, it logically would be from a post-Grove Church, pre-Wayside unit representing a nearshore, high-energy environment. The probability of finding additional specimens of *Rhachistognathus* is very low, as the shale of the Wayside indicates a low-energy environment which, as previously mentioned, is atypical for most rhachistognathids.

The presence of the Pa element of *Lochriea commutata* is more of an enigma. Although it may range higher in Europe, *L. commutata* has not been reliably reported from undisputed Pennsylvanian strata in



Text-figure 6. Biostratigraphic chart showing conodont zones for the latest Chesterian and Morrowan Series. Known conodont ranges of species encountered in this study are shown by vertical bars. Lower horizontal shaded area indicates approximate age of Grove Church. Upper shaded area indicates suspected age for lower shale interval of Wayside Sandstone Member. Range of *A. lautus* includes *A. lautus* morphotype A. Most Chesterian conodont species with "?" at top of their ranges probably range throughout the *R. muricatus* Biozone.

North America. Grayson and others (1985) did report abundant *Paragnathodus commutatus* (= *L. commutata*) from their "*Declinognathodus noduliferus* Zone" in Oklahoma. They indicated that these specimens and their enclosing strata may represent an interval between the traditional Mississippian and Pennsylvanian. The Wayside specimen could also be reworked, but from the Kinkaid Limestone or lower; it is not impossible but very unlikely. If this specimen is indeed indigenous, then the joint occurrence of *L. commutata* and *R.* cf. *R. websteri*, would represent a level within the *D. noduliferus–R. primus* Biozone and a very early Morrowan age (Text-fig. 6). The presence of the one Pa element of a Pennsylvanian species of *Rhachistognathus* places the lower part of the Wayside within the Pennsylvanian. Our study of the core and outcrops suggests that other unconformities exist within strata assigned to the Wayside. Higher portions of the Wayside may represent middle or later Morrowan zones as suggested by some conodont species illustrated by Rexroad and Merrill (1985). A closer examination of the gnathodontoids recovered by Rexroad and Merrill (1985) may also provide additional age information.

CONCLUSIONS

We have identified *three* disconformities within the Mississippian–Pennsylvanian boundary strata of southern Illinois: (1) Root casts and a breccia zone at the Kinkaid–Grove Church boundary provide evidence for an exposure surface. The sudden appearance of abundant elements of *Adetognathus unicornis* along with a general reduction of numbers and species of typical Chesterian conodonts supports the probability of a significant hiatus at this boundary. (2) A preserved soil profile provides evidence for an intraformational paleosol and associated hiatus within the Grove Church. (3) A sharp, well-defined contact and associated abrupt changes in lithology and shale mineralogy indicate a disconformity at the Mississippian–Pennsylvanian contact. The change from a moderately diverse *A. unicornis* assemblage in the Grove Church to an essentially monospecific assemblage of *A. lautus* morphotype A in the Wayside supports this hiatus. The presence of rare elements of *Lochriea commutata* and *Rhachistognathus* cf. *R. websteri* suggests, however, that this hiatus was of relatively short duration (probably less than two conodont zones). Sedimentologic and conodont data (Rexroad and Merrill, 1985) suggest additional hiatuses may be present higher within Wayside strata.

We conclude that deposition of late Mississippian and early Pennsylvanian strata at the southern edge of the Illinois basin was interrupted by several hiatuses rather than by only one, an interpretation having important implications in the recognition of sequence boundaries. The Mississippian–Pennsylvanian boundary separates the Kaskaskia and Absaroka sequences of Sloss (1963), who defined sequence boundaries as major regional unconformities. The transition of this boundary from a single, major unconformity in the northern and central parts of the basin to a number of smaller unconformities at the southern edge indicates that a correlative, conformable boundary could have occurred farther south, either in the part of the Illinois basin uplifted (and now eroded) by the Pascola arch, or in the Reelfoot rift.

REFERENCES CITED

Baesemann, J. F.; and Lane, H. R., 1985, Taxonomy and evolution of the genus *Rhachistognathus* Dunn (Conodonta; Late Mississippian to Early Middle Pennsylvanian), *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg, v. 74, p. 93–136.

Birkeland, P. W., 1984, Soils and geomorphology: Oxford University Press, New York, 372 p.

Bristol, H. M.; and Howard, R. H., 1971, Paleogeologic map of the sub-Pennsylvanian Chesterian (Upper Mississippian) surface in the Illinois basin: Illinois State Geological Survey Circular 458, 14 p.

Collinson, C.; Scott, A. J.; and Rexroad, C. B., 1962, Six charts showing biostratigraphic zones and correlations based on conodonts from the Devonian and Mississippian rocks of the Upper Mississippi Valley: Illinois State Geological Survey Circular 343, 32 p.

Collinson, C.; Rexroad, C. B.; and Thompson, T. L., 1971, Conodont zonation of the North American Mississippian, *in* Sweet, W. C.; and Bergström, S. M. (eds.), Symposium on conodont biostratigraphy: Geological Society of America Memoir 127, p. 353–394.

Cooper, C. L., 1947, Upper Kinkaid (Mississippian) microfauna from Johnson County, Illinois: Journal of Paleontology, v. 21, p. 81–93.

Ethridge, F. G.; and Bailey, L. T., 1976, Sedimentary facies and depositional model, *in* Ethridge, F. G.; and Fraunfelter, G. H. (eds.), Variability within a high constructive lobate delta: the Pounds Sandstone of southeastern Illinois: Sixth Annual Field Conference, Great Lakes Section, Society of Economic Paleontologists and Mineralogists, p. 1–9.

Ethridge, F. G.; Leming, S. L.; and Keck, D. A., 1973, Depositional environments of Lower Pennsylvanian detrital sediments of southern Illinois, *in* Ethridge, F. G.; Fraunfelter, G. H.; and Utgaard, J. (eds.), Depositional environments of selected Lower Pennsylvanian and Upper Mississippian sequences in southern Illinois: 37th Tri-State Field Conference, Carbondale, Illinois, p. 16–34.

Ettensohn, F. R.; Dever, G. R., Jr.; and Grow, J. S., 1988, A paleosol interpretation for profiles exhibiting subaerial exposure "crusts" from the Mississippian of the Appalachian basin, *in* Reinhardt, J.; and Sigleo, W. R. (eds.), Paleosols and weathering through geologic time: principles and applications: Geological Society of America Special Paper 216, p. 49–79.

Follmer, L. R.; McKay, E. D.; Lineback, J. A.; and Gross, D. L., 1979, Wisconsinan, Sangamonian, and Illinoian stratigraphy in central Illinois: Illinois State Geological Survey Guidebook 13, 139 p.

Grayson, R. C., Jr.; Davidson, W. T.; Westergaard, E. H.; Atchley, S. C.; Hightower, J. H.; Monaghan, P. T.; and Pollard, C. D., 1985, Mississippian–"Pennsylvanian" (Mid-Carboniferous) boundary conodonts from the Rhoda Creek Formation: *Homoceras* equivalent in North America, *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg, v. 74, p. 149–180.

Hughes, R. E.; DeMaris, P. J.; White, W. A.; and Cowin, D. K., 1987, Origin of clay minerals in Pennsylvanian strata of the Illinois basin: Proceedings of the International Clay Conference, Clay Minerals Society, Bloomington, Indiana, p. 97–104.

Jennings, J. R.; and Fraunfelter, G. H, 1986, Preliminary report on macro-paleontology of strata above and below the upper boundary of the type Mississippian: Transactions of the Illinois Academy of Science, v. 79, p. 253–261.

Klappa, C. F., 1980, Brecciation textures and tepee structures in Quaternary calcrete (caliche) profiles from eastern Spain; the plant factor in their formation: Geological Journal, v. 15, p. 81–89.

Krumhardt, A. P.; and Harris, A. G., 1990, Conodont biostratigraphy and biofacies of the Wahoo Limestone (Carboniferous), Sadlerochit Mountains, NE Brooks Range, Alaska [abstract]: Geological Society of America Abstracts with Programs, v. 22, no. 1, p. 12.

Lamar, J. E., 1925, Geology and mineral resources of the Carbondale Quadrangle: Illinois State Geological Survey Bulletin 48, 172 p.

Lane, H. R., 1967, Uppermost Mississippian and lower Pennsylvanian conodonts from the type Morrowan region, Arkansas: Journal of Paleontology, v. 41, p. 920–942.

_____1968, Symmetry in conodont element-pairs: Journal of Paleontology, v. 42, p. 1258–1263.

Lane, H. R.; and Straka, J. J., II, 1974, Late Mississippian and early Pennsylvanian conodonts Arkansas and Oklahoma: Geological Society of America Special Paper 152, 144 p.

Merrill, G. K.; and Rexroad, C. B., 1981, Conodonts from the Middle Carboniferous of southern Illinois and the base of the Pennsylvanian System [abstract]: Geological Society of America Abstracts with Programs, v. 13, no. 6, p. 309.

Morrow, J. R.; and Webster, G. D., 1990, Conodont biofacies and related carbonate microfacies, Mid-carboniferous boundary strata, Granite Mountain, west-central Utah [abstract]: Geological Society of America Abstracts with Programs, v. 22, no. 1, p. 29.

Norby, R. D.; and Rexroad, C. B., 1985, *Vogelgnathus*, a new Mississippian conodont genus: Indiana Geological Survey Occasional Paper 50, 21 p.

Ostrom, M. E.; and Potter, P. E., 1961, A clay mineral sequence at the Mississippian–Pennsylvanian unconformity in the Illinois basin: Journal of Geology, v. 69, p. 341–351.

Retallack, G. J., 1988, Field recognition of paleosols, *in* Reinhardt, J.; and Sigleo, W. R. (eds.), Paleosols and weathering through geologic time: principles and applications: Geological Society of America Special Paper 216, p. 1–20.

Retallack, G. J.; Grandstaff, G. E.; and Kimberly, N. M., 1984, The promise and problems of Precambrian paleosols: Episodes, v. 7, p. 8–12.

Rexroad, C. B.; and Burton, R. C., 1961, Conodonts from the Kinkaid Formation (Chester) in Illinois: Journal of Paleontology, v. 6, p. 1143–1158.

Rexroad, C. B.; and Merrill, G. K., 1979, Conodont biostratigraphy and the Mississippian–Pennsylvanian boundary in southern Illinois [abstract]: Ninth International Congress of Carboniferous Stratigraphy and Geology, Urbana, Illinois, Abstracts of Papers, p. 179–180.

_____1985, Conodont biostratigraphy and paleoecology of Middle Carboniferous rocks in southern Illinois, *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg, v. 74, p. 35–64.

Siever, R., 1951, Mississippian–Pennsylvanian unconformity in southern Illinois: American Association of Petroleum Geologists Bulletin, v. 35, p. 542–581.

Sloss, L. L., 1963, Sequences in the cratonic interior of North America: Geological Society of America Bulletin, v. 74, p. 93–114.

Swann, D. H., 1963, Classification of Genevievian and Chesterian (Late Mississippian) rocks of Illinois: Illinois State Geological Survey Report of Investigations 216, 91 p.

Wanless, H. R., 1955, Pennsylvanian rocks of Eastern Interior basin: American Association of Petroleum Geologists Bulletin, v. 39, p. 1753–1820.

Worthen, A. H., 1866, Stratigraphical geology, Tertiary deposits and coal measures, *in* Geology: Geological Survey of Illinois, v. 1, p. 40–76.

APPENDIX: Collecting Localities

1. COGEOMAP L-2 (ISGS County No. 20427, Core No. C13459), NW¼NE¼ sec. 31, T. 11 S., R. 2 E., Johnson County, Illinois, Lick Creek 7.5′ Quadrangle. Total depth is 264 ft.

2. NE¼NW¼ sec. 31, T. 11 S., R. 2 E., Johnson County, Illinois, Lick Creek 7.5′ Quadrangle. In small ravine near top of Cedar Bluff, 45 m west of road leading northward from Cedar Grove Church. This exposure is the type section of the Grove Church Shale (Swann, 1963) and includes the lower part of the Wayside Sandstone Member, Caseyville Formation. This is locality 2 of Rexroad and Burton (1961) and Rexroad and Merrill (1985).

PLATE 1

Grove Church conodonts. Figures are SEM micrographs of specimens that are reposited in the Illinois State Geological Survey Paleontology Collections under the accession prefix ISGS.

Figures 1–16.—*Adetognathus unicornis* (Rexroad and Burton, 1961), pairs of upper and inner lateral views of 8 Pa elements. **1,2**, dextral; core C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-1). **3,4**, sinistral; C13459, 59.2–61.2 ft depth; x60 (ISGS 81P-2). **5,6**, sinistral; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-3). **7,8**, dextral; C13459, 59.2–61.2 ft depth; x60 (ISGS 81P-4). **9,10**, sinistral; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-5). **11,12**, holotype, dextral; near top of limestone bed, ~6.75 ft below basal sandstone bed of Wayside at type Grove Church Shale locality; x60 (ISGS 15P-39). **13,14**, paratype, sinistral; within limestone bed, ~10 ft below basal sandstone bed of Wayside at type Grove Church Shale locality; x60 (ISGS 15P-90). **15,16**, paratype, dextral; within limestone bed, ~10 ft below basal sandstone bed of Wayside at type Grove Church Shale locality; x60 (ISGS 15P-89).

Figures 17,19,20.—*Cavusgnathus naviculus* (Hinde, 1900). **17**, inner lateral view of dextral(?) Pa element; C13459, 61.2–63.7 ft depth; x50 (ISGS 81P-6). **19**, inner lateral view of dextral M element; C13459, 61.2–63.7 ft depth; x45 (ISGS 81P-8). **20**, inner lateral view of sinistral Pb element; C13459, 61.2–63.7 ft depth; x45 (ISGS 81P-9).

Figure 18.—*Cavusgnathus unicornis* Youngquist and Miller, 1949, inner lateral view of dextral Pa element; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-7).

Figure 21.—*Hindeodus minutus* (Ellison, 1941), inner lateral view of sinistral Pa element; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-10).

Figure 22.—*Hindeodus cristula* (Youngquist and Miller, 1949), inner lateral view of dextral(?) Pa element; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-11).

Figure 23.—*Vogelgnathus* n. sp. of Norby and Rexroad, 1985, inner lateral view of sinistral(?) Pa element; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-12).

Figures 24–26.—*Kladognathus* cf. *K. tenuis* (Branson and Mehl, 1941). **24**, inner lateral view of antero-lateral bar of dextral Sb(?) element; C13459, 61.2–63.7 ft depth; x45 (ISGS 81P-13). **25**, inner lateral view of dextral Sc element; C13459, 61.2–63.7 ft depth; x60 (ISGS 81P-14). **26**, inner lateral view of sinistral M element; C13459, 61.2–63.7 ft depth; x45 (ISGS 81P-15).

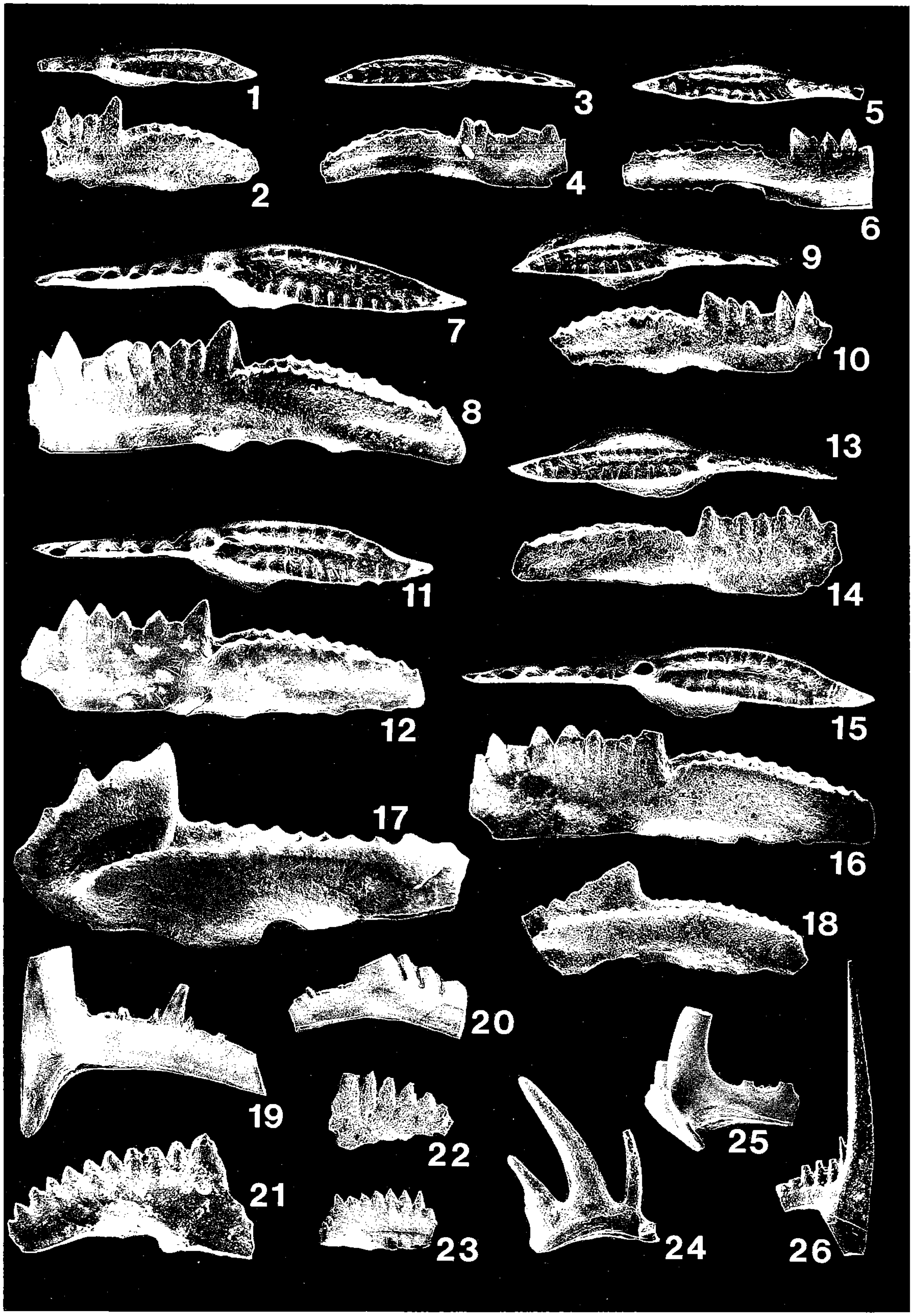

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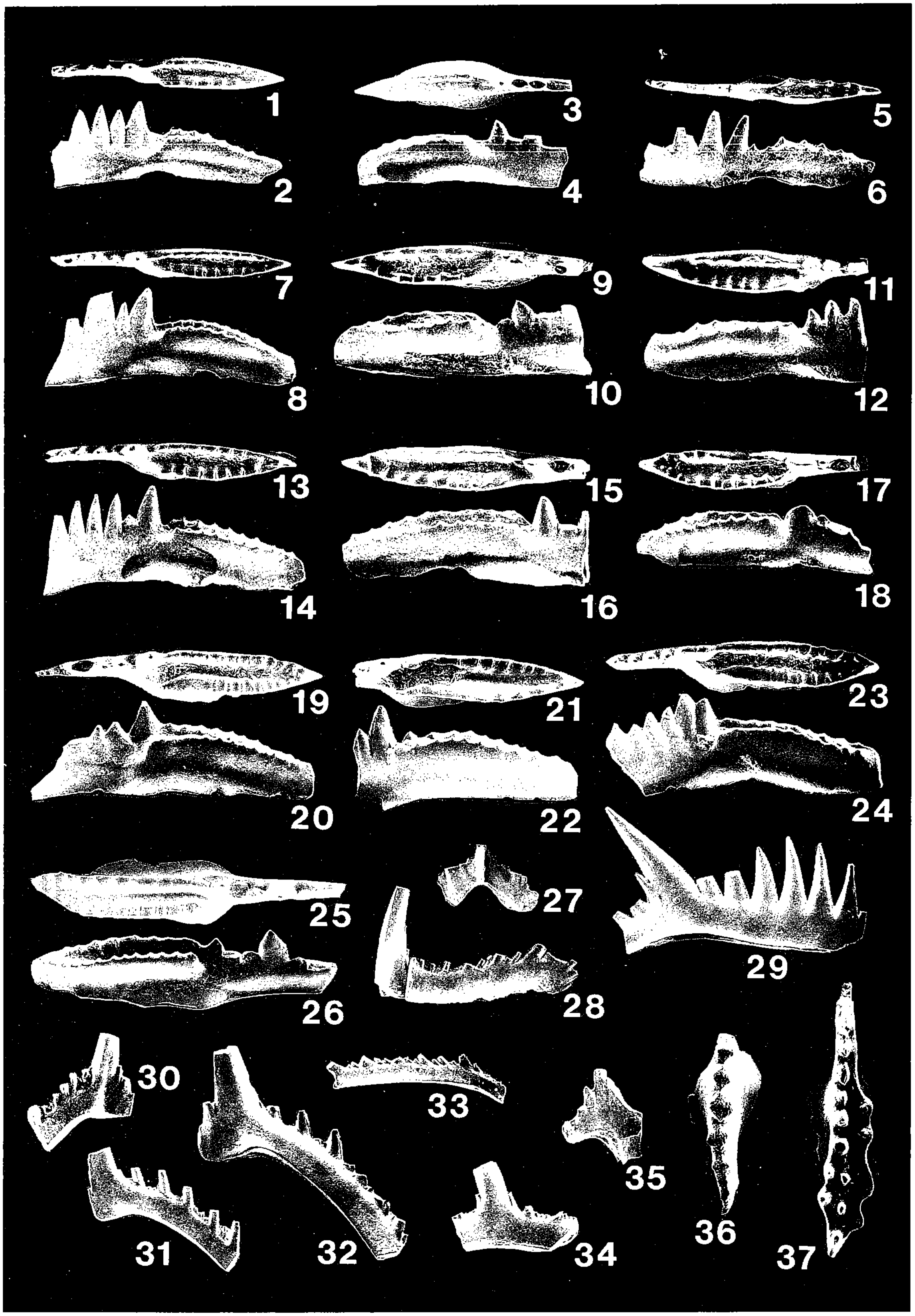
PLATE 2

Wayside conodonts. Figured specimens are from Wayside Sandstone at locality 2 (type section of the Grove Church Shale). Sample intervals are shown on Text-figure 2. Additional figure notes are given on Plate 1.

Figures 1–34.—*Adetognathus lautus* (Gunnell, 1933) morphotype A. **1–26** are pairs of upper and inner lateral views of 13 Pa elements, except Fig. 6 is an outer lateral view. **1,2,** dextral, sample C3, ×60 (ISGS 81P-16). **3,4,** sinistral, sample C1, ×60 (ISGS 81P-17). **5,6,** sinistral, sample A, ×75 (ISGS 81P-18). **7,8,** dextral, sample C3, ×60 (ISGS 81P-19). **9,10,** sinistral, sample B2, ×75 (ISGS 81P-20). **11,12,** sinistral, sample A, ×75 (ISGS 81P-21). **13,14,** dextral, sample C2, ×60 (ISGS 81P-22). **15,16,** dextral, sample C2, ×75 (ISGS 81P-23). **17,18,** sinistral, sample C1, ×60 (ISGS 81P-24). **19,20,** dextral sample C1, ×60 (ISGS 81P-25). **21,22,** dextral, sample A, ×60 (ISGS 81P-26). **23,24,** dextral, sample C1, ×60 (ISGS 81P-27). **25,26,** sinistral, sample A, ×60 (ISGS 81P-28). **27,** posterior view of Sa element, sample C1, ×60 (ISGS 81P-29). **28,** lateral view of Sa element, sample B1, ×60 (ISGS 81P-30). **29,** inner lateral view of sinistral Pb element, sample A, ×75 (ISGS 81P-31). **30,** inner lateral view of sinistral M element, sample A, ×60 (ISGS 81P-32). **31,** inner lateral view of dextral M element, sample C2, ×60 (ISGS 81P-33). **32,** inner lateral view of dextral M element, sample A, ×75 (ISGS 81P-34). **33,** inner lateral view of sinistral Sb element, sample B2, ×60 (ISGS 81P-35). **34,** inner lateral view of sinistral Sc element, sample C1, ×60 (ISGS 81P-36).

Figures 35,36.—*Lochriea commutata* (Branson and Mehl, 1941). **35,** inner lateral view of sinistral M element, sample from basal sandstone, ×60 (ISGS 81P-37). **36,** upper view of sinistral Pa element, sample C3, ×100 (81P-38).

Figure 37.—*Rhachistognathus* cf. *R. websteri* Baesemann and Lane, 1985, upper view of dextral Pa element, sample C3, ×100 (ISGS 81P-39).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Oklahoma Geological Survey Circular 94, 1992

New Stratigraphic, Petrographic, and Biostratigraphic Data on the Proposed Mississippian–Pennsylvanian Boundary Stratotype, Granite Mountain, West-Central Utah

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ABSTRACT.—The mid-Carboniferous boundary at Granite Mountain occurs in the lower portion of the Ely Limestone. A total of nine shallow-water carbonate microfacies, representing a wide range of fabrics and textures, are recognized from four measured sections across the boundary interval. Lateral stratigraphic changes within the Ely Limestone, such as thinning and pinching-out of individual beds, are common on the outcrop. Important facies changes occurring below the systemic boundary include the thinning and loss of limestone and calcareous shale beds, when traced southward to the extent of exposure. Above the systemic boundary, loss of calcareous shale interbeds, accompanied by decrease in chert, is observed from north to south.

Conodont faunas are dominated by species of *Cavusgnathus*, *Adetognathus*, and *Rhachistognathus*. Plots of conodont abundances versus microfacies types show alternating occurrences of the *Adetognathus* and *Rhachistognathus* biofacies, which reflect small-scale depositional cycles. Gnathodid conodont species, including the important zonal index *Declinognathodus noduliferus*, are rare, and the evolutionary sequence of *D. noduliferus* is not present. Low abundances of the gnathodid conodonts are attributed to environmental control. Blastoid specimens recovered near the top of the *Rhipidomella nevadensis* Zone are assignable to *Pentremites cherokeeus* (= *P. crystalensis*) and questionably *P. laminatus*.

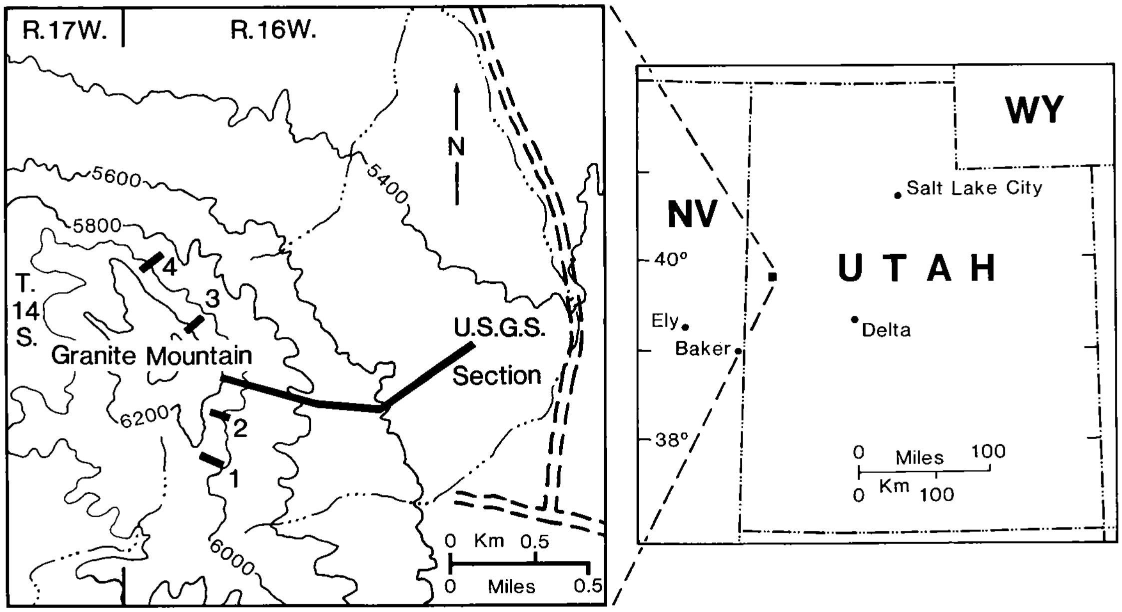
The rare gnathodid conodont species and apparent lack of the evolutionary sequence of *Declinognathodus noduliferus* in the Granite Mountain sections detract from its suitability as a mid-Carboniferous boundary stratotype. *Rhachistognathus primus* appears below *D. noduliferus* and does not mark the mid-Carboniferous boundary at Granite Mountain.

INTRODUCTION

At the 1975 Subcommission on Carboniferous Stratigraphy meeting held in conjunction with the Eighth International Congress on Carboniferous Stratigraphy in Moscow, a proposal for an international classification of the Carboniferous System utilizing the names Mississippian and Pennsylvanian of North America was presented (Lane and Manger, 1985). That proposal focused attention on the need for more precise stratigraphic and biostratigraphic data on the Mississippian–Pennsylvanian boundary in North America, especially if those classifications were to be used on an intercontinental scale.

In the 10 years following the Moscow congress, numerous studies of North American Mississippian–Pennsylvanian boundary sections were made. One location examined for a potential Mississippian–Pennsylvanian boundary stratotype section is the Confusion Range area of west-central Utah (Webster and others, 1984a). Exposures at Granite Mountain, located at the northeastern end of the Confusion Range in southern Juab County, are considered the most important of the possible west-central Utah sections (Text-fig. 1). Gently westward-dipping outcrops on the northeastern and eastern slopes of Granite Mountain provide nearly 100% exposure of the Mississippian–Pennsylvanian boundary, which occurs in a fossiliferous carbonate succession that is considered to contain no significant hiatus or major facies change at the systemic boundary (Gordon and others, 1983,1985). A single Granite Mountain section has been sampled by the U.S. Geological Survey for lithofacies and fossils, including brachiopods (Gordon and others, 1982, 1984,1985), conodonts (Wardlaw, 1984), and calcareous foraminifers and algae (Mamet, 1982,1984; Gordon and others, 1985).

The purpose of the present paper is to give new data on the stratigraphy, carbonate microfacies, and conodont biofacies of the mid-Carboniferous boundary strata laterally adjacent to the USGS section at Granite Mountain. In addition, previously



Text-figure 1. Location map of study area. Measured sections 1 to 4 are referred to in the text as GM-1 to GM-4, respectively. The USGS section is that of Gordon and others (1983,1984,1985). The USGS section extends from the top of the Pilot Shale (Upper Devonian and Lower Mississippian?) upwards across the Chainman Shale (Mississippian) into the lower portion of the Ely Limestone (Upper Mississippian and Lower Pennsylvanian). Contour interval is 200 ft. Base is from USGS Granite Mountain, Utah, 15′ Quadrangle. Detailed directions and mileages to the Granite Mountain exposure are given by Webster and others (1984b).

reported Mississippian–Pennsylvanian boundary conodonts and brachiopods are illustrated, and new occurrences of the blastoid *Pentremites* are described. Data presented here are based on measurement of four 11–16 m sections, designated GM-1 to GM-4, that span the Mississippian–Pennsylvanian boundary. A total of 150 petrographic and conodont samples, with average sample spacing of 35 cm, were taken from the four sections. The 1 to 1.5 kg samples were processed for conodonts, except near conodont zone boundaries, where 2.0 to 3.0 kg were dissolved.

Acknowledgments

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STRATIGRAPHY AND CARBONATE MICROFACIES

At Granite Mountain, the Mississippian–Pennsylvanian boundary occurs in the lower portion of the Ely Limestone, an extensive, shallow, platform carbonate recognized over 50,000 km^2 in eastern Nevada and western Utah (Steele, 1960; Hose, 1974). The Ely Limestone reaches 610 m in nearby sections (Hose, 1974), but only its lower 215 m forms an erosional remnant capping Granite Mountain. Based on the lowest stratigraphic occurrence of the conodont *Declinognathodus noduliferus*, the Mississippian–Pennsylvanian boundary occurs from 45 to 53 m above the base of the Ely Limestone.

The Ely Limestone typically weathers into alternating cliff and slope topography. Resistant beds are composed of medium- to thick-bedded or massive grainstones, packstones, and wackestones that are separated by recessive, thin-bedded to laminated wackestones, lime mudstones, calcareous shales, and rarely grainstones and packstones. Often, recessive units containing intergranular lime mud matrix also contain common quartz silt and very fine quartz sand. Bioturbation is a common feature in all units, giving many beds a mottled appearance. Rarely, discrete burrows are visible. The slope-forming beds may contain abundant silicified brachiopods. Tabular cross-bedding and cross-laminations, enhanced by pressure solution and differential weathering, occur within resistant grainstone units. These cross-strata have dip angles of 20–25° (corrected for bedding plane dip), with a dip direction due east. Sucrosic dolomite has locally replaced both cliff- and slope-forming units. Although rarely present below the systemic boundary, chert nodules and lenses become very common in the Pennsylvanian portion of the Ely Limestone at Granite Mountain.

When traced laterally, many beds adjacent to the systemic boundary thin, pinch-out, and/or change lithology. A calcareous shale bed present from 5.7 to 6.1 m at section GM-4, and 6.2 to 6.5 m at GM-3, pinches-out to the south (Text-fig. 2), where it is probably equivalent to the horizon between dolomitized and nondolomitized strata at 5.0 m at the USGS section and at 4.5 m at GM-2. Another nodular, calcareous shale interval present at GM-4 (9.6–9.9 m) and GM-3 (9.5–9.9 m) is correlated to the south with a thinly bedded dolostone unit from 6.7 to 7.3 m at the USGS section (Text-fig. 2). This interval is in turn correlative with a silty, recessive-weathering unit present from 5.4 to 5.8 m at GM-2, and is probably equivalent to calcareous shales and thinly bedded limestones present from 4.0 to 5.4 m at GM-1. Thinly bedded dolostone present from 7.7 to 7.9 m at the USGS section develops southward into a silty, recessive-weathering unit present from 7.0 to 7.3 m at GM-2, but is covered at GM-1 (6.7–7.4 m).

At section GM-3, a resistant grainstone/packstone bed (11.2–12.2 m) is bounded above and below by recessive-weathering, in part silty and nodular, calcareous shale beds. The grainstone/packstone bed thins to the south and occupies 8.4–8.6 m at the USGS section. By GM-2, this grainstone/packstone bed is present only as limestone nodules within a calcareous shale interval from 7.8 to 8.3 m. Farther to the south, the grainstone/packstone bed is gone, although the bounding calcareous shale beds are still present from 8.0 to 8.2 m at GM-1. To the north of GM-3, the bounding beds thin and disappear, so that the GM-3 grainstone/packstone interval correlates to the 11.0–11.3 m portion of a thicker, resistant grainstone/packstone that spans the 9.9–12.0 m interval at GM-4.

At Granite Mountain, the lowest occurrence of the conodont *Declinognathodus noduliferus* falls just below or within the lowest horizon of persistent chert. This chert lies in a grainstone/packstone unit present from 8.2 to 9.2 m at section GM-1, 8.3 to 8.9 m at GM-2, 8.7 to 9.6 m at the USGS section, 12.2 to 12.8 m at GM-3, but not clearly recognizable at GM-4. Based on the lowest occurrence of *Declinognathodus noduliferus,* the equivalent interval occurs from 11.3 to 12.0 m at GM-4.

Overlying the lower resistant chert interval at all Granite Mountain sections is a zone of chert-rich limestone and recessive-weathering, nodular calcareous shale containing locally abundant productid brachiopods typical of the Gordon and others (1984,1985) "*Rugoclostus* bed" (Text-fig. 2). At GM-2, the USGS section, and GM-3, the "*Rugoclostus* bed" interval is split by a medium- to thick-bedded, chert-rich packstone bed present at 9.2–9.5 m, 9.8–10.3 m, and 13.0–13.3 m, respectively. This bed is not recognized to the north at GM-4 or to the south at GM-1. In addition, the amount of chert present in the upper portion of the "*Rugoclostus* bed" interval decreases significantly from north to south.

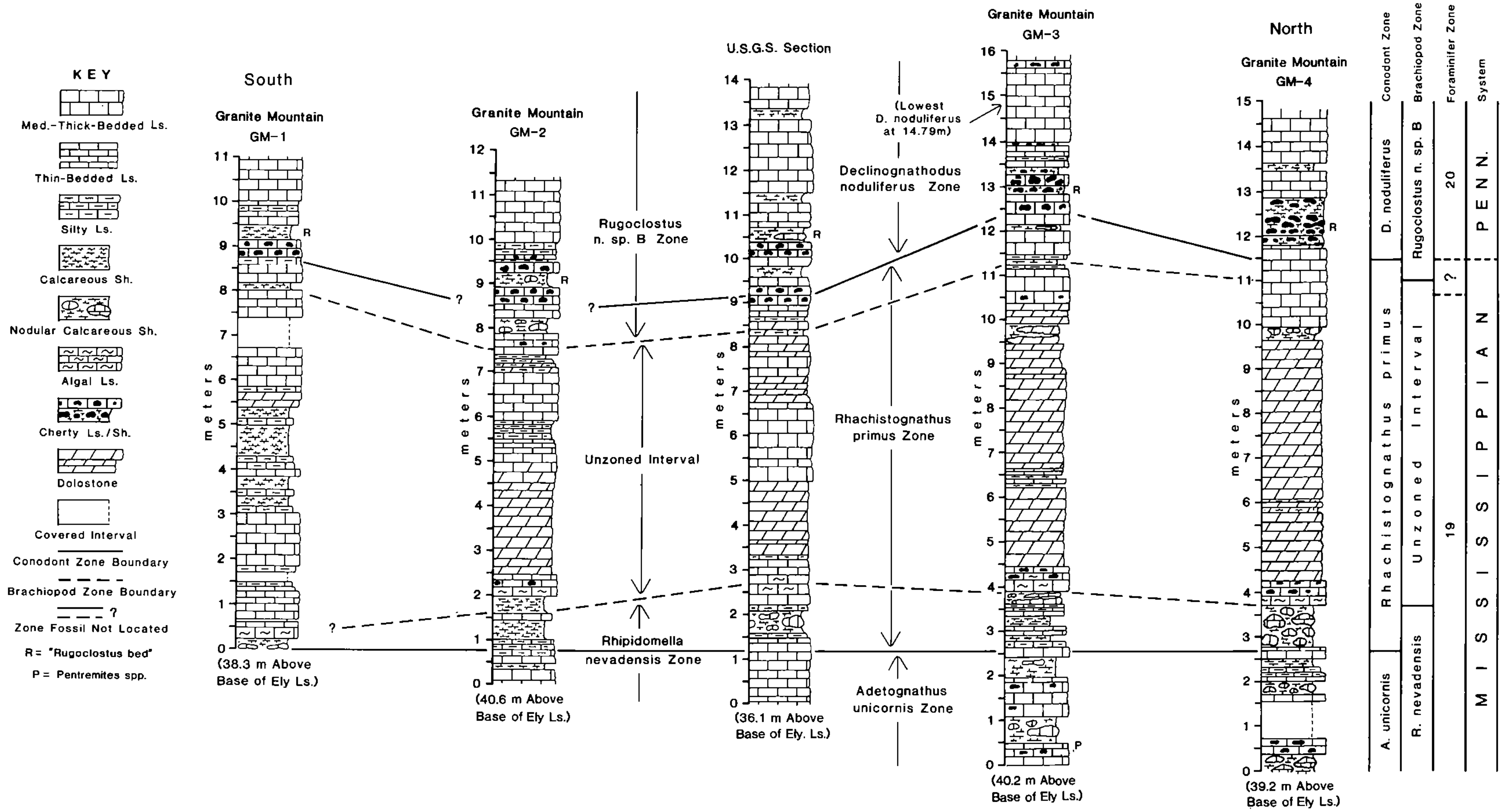
Above the "*Rugoclostus* bed" interval, a thick-bedded to massive, cliff-forming grainstone/packstone interval is broken by discontinuous horizons of calcareous shale or thin-bedded, silty limestone. A calcareous shale present from 13.4 to 13.6 m at GM-4 can be correlated southward to GM-3 (13.9–14.0 m) and the USGS section (11.2–11.5 m). South of the USGS section, however, this bed correlates to a thin-bedded, silty limestone unit that is present at GM-2 (9.6–9.8 m) and GM-1 (9.8–9.9 m).

Although the variations in strata noted above are attributed primarily to lateral facies changes, the presence of sharp lithologic contacts and lime mud-rich intraclasts within higher energy grainstone microfacies gives evidence of at least minor localized scouring or reworking between beds. Good examples are present in sections GM-1 (e.g., 1.7–2.2 m, 3.7–4.0 m, and 5.7–6.1 m), and GM-2 (e.g., 4.4–5.0 m and 5.9–6.9 m). Similarities in style of sedimentation of bounding beds, as well as ranges and occurrences of conodonts and brachiopods, suggest a very short duration for these possible diastems, although the lowest occurrence of the conodont *Rhachistognathus muricatus* is directly above an intraclast grainstone lag horizon at GM-4 (2.0–2.5 m).

Differential weathering of dolomitized verses nondolomitized zones within the same bed reflects significant lateral variations in outcrop character and apparent bed thickness. An example is a medium-gray packstone bed present from 4.9 to 6.7 m at the USGS section, which changes northward into a tan-orange weathered dolostone (i.e., GM-3, 6.6–8.7 m). In another example, the medium-gray grainstone/packstone bed from 1.8 to 3.0 m at GM-1 abruptly grades into a tan-orange, poorly exposed, blocky, dolostone within a lateral distance of ~15 m to the southeast. In general, there is a change from more resistant, thickly bedded or massive dolomitized strata in the north to slope-forming, thinly bedded, nondolomitized strata in the south.

Carbonate Microfacies

A total of nine recurring carbonate microfacies were recognized in the lower Ely Limestone at Granite Mountain. These microfacies are typically thin and reflect a wide range of depositional fabrics



Text-figure 2. Correlation chart, Granite Mountain sections. Foraminifer zones are those of Mamet (1982,1984) and in Gordon and others (1984,1985). The measured sections begin 36–41 m above the base of the Ely Limestone as mapped by Hose (1974). Datum is the base of the *Rhachistognathus primus* Zone.

and textures. For ease in plotting on stratigraphic columns, the microfacies were categorized using the Standard Microfacies Types (SMF Types) of Wilson (1975) as modified by Flügel (1982). The following SMF Types were identified in the study: SMF-9, bioclast wackestone microfacies; SMF-10, coated and worn bioclast packstone/wackestone microfacies; SMF-11, coated bioclast grainstone microfacies; SMF-12, bioclast packstone/grainstone microfacies; SMF-14, intraclast lag microfacies; SMF-15, ooid grainstone microfacies; and SMF-16/17, pellet and lump packstone microfacies. SMF-11, the coated bioclast grainstone microfacies, was further subdivided into a fine to medium calcarenite microfacies (designated SMF-11f) and a medium calcarenite to medium calcirudite microfacies (designated SMF-11c). The shallow-water, spiculitic lime mudstone microfacies (SMF-1-s), originally described by Stevens and Armin (1983) from the Middle Pennsylvanian portion of the Oquirrh Group, central Utah, was also recognized.

SMF-9 is the most common microfacies, comprising 30% total of the measured sections. SMF-12 comprises 29%; SMF-10, 20%; SMF-11c, 10%; and SMF-11f and SMF-16/17, 4% each. The remaining microfacies, SMF-1-s, SMF-14, and SMF-15, each comprise 1% of the total exposure. Detailed discussion of the carbonate microfacies and a proposed depositional model for the Mississippian–Pennsylvanian boundary strata at Granite Mountain are given by Morrow and Webster (1991).

Petrographic observations indicate that dolomite replacement was a post-depositional diagenetic process altering an original skeletal limestone, with some dolomitized samples retaining sufficient relict texture of the precursor limestone to allow tentative assignment to a SMF Type. Precursor textures are preserved by silica replaced skeletal grains and whole fossils that survived dolomitization, and by dolomite replacement that mimics earlier calcium carbonate crystal size (e.g., very finely crystalline dolomite replacing micrite and coarse, single-crystal dolomite replacing echinoderm grains with syntaxial rims). In some cases, high abundances of conodont elements recovered from dolomitized samples were also used to infer precursor microfacies types (see discussion under conodont biofacies). Based on cross-cutting relationships, dolomitization postdates pervasive silica-replacement of bioclasts and whole fossils. The silicified grains are composed primarily of echinoderm and brachiopod debris that indicate a relatively normal marine precursor limestone host for the dolomite.

CONODONT BIOSTRATIGRAPHY AND BIOFACIES

A total of 15 previously described conodont species representing six genera are recognized from the Mississippian–Pennsylvanian boundary strata at Granite Mountain. Average conodont abundance was 28 Pa elements per kg of sample, with values ranging from 0 to 141 elements per kg. Important Pa elements, and ramiform elements associated with Pa elements of *Adetognathus lautus* and *A. spathus*, are illustrated in Plate 1.

Published reports on conodont biostratigraphy and proposed zonation schemes for the mid-Carboniferous utilizing data from the Granite Mountain–Confusion Range area include Dunn (1970a), Sandberg and others (1980), Tynan (1980), Gordon and others (1984), Wardlaw (1984), Webster (1984), and Webster and others (1984b). The mid-Carboniferous conodont zonation scheme used by Gordon and others (1984) and Wardlaw (1984), which is based on material collected from Granite Mountain, is followed in the present study (Text-fig. 2). In ascending order, the conodont zones recognized are the late Chesterian *Adetognathus unicornis* Zone (upper part only), the *Rhachistognathus primus* Zone, and the succeeding *Declinognathodus noduliferus* Zone (lower part only). These zones are interval zones defined by the first occurrence and successive occurrences of the defining species below the first occurrence of the species used to define the base of the overlying zone.

The lowest occurrence of *Rhachistognathus primus* was originally considered to be of earliest Pennsylvanian age by Lane and Baesemann (1982). However, the first occurrence of *R. primus* has been reported to be consistently below *Declinognathodus noduliferus* at several localities in the western U.S., i.e., Granite Mountain by Wardlaw (1984); Willow Wash in the Confusion Range, Jensen Wash in the Burbank Hills, and Soldier Canyon in the Oquirrh Range by Webster and others (1984b); and the Alaska Bench Formation of Montana by Davis and Webster (1985a). Only at Arrow Canyon in southern Nevada are the first occurrences of both species coincident (Lane and Baesemann, 1982). Since the lowest occurrence of *D. noduliferus* is presently designated the base of the Pennsylvanian, we consider the *R. primus* Zone to be latest Chesterian.

Other Pa elements occurring in the *Adetognathus unicornis* Zone include: *A. lautus*, *Cavusgnathus unicornis*, *Gnathodus bilineatus* (reworked), *G. defectus*, *G. girtyi girtyi*, *G. girtyi simplex*, *Hindeodus minutus*, and *Rhachistognathus muricatus*. The fauna at the base of sections GM-1 to GM-4 is dominated by elements of *A. unicornis* and *C. unicornis*, with *A. unicornis*, *A. lautus*, and *R. muricatus* being most common in samples from the uppermost portion of the zone. Elements of *Gnathodus* spp. and *H. minutus* form a rare component of the fauna in all sections.

Overlying the *Adetognathus unicornis* Zone is the *Rhachistognathus primus* Zone. Pa elements associated with *R. primus* in this zone include: *A. lautus*, *A.* n. sp. (of Tynan, 1980), *A. spathus*, *Gnathodus bilineatus* (reworked), *G. defectus*, *G. girtyi girtyi*, *G.*

girtyi simplex, Hindeodus minutus, Rhachistognathus muricatus, R. prolixus, and *R. websteri.* The fauna within this zone is dominated by *Adetognathus* spp. and *Rhachistognathus* spp., with *Gnathodus* spp. and *Hindeodus minutus* comprising rare components.

The *Declinognathodus noduliferus* Zone is the youngest zone found. The upper boundary of the *D. noduliferus* Zone, defined by the first occurrence of *Idiognathoides sinuatus,* is recognized in the overlying portion of the Ely Limestone at both Granite Mountain (Wardlaw, 1984) and in the Confusion Range (Webster and others, 1984b). Other Pa elements occurring with *D. noduliferus* are: *Adetognathus lautus, A. spathus, Gnathodus defectus, Rhachistognathus muricatus, R. primus, R. prolixus,* and *R. websteri.* In all sections, *D. noduliferus* and *G. defectus* form a minor component of the conodont faunas, which are dominated by species of *Adetognathus* and *Rhachistognathus.* Rarity of *D. noduliferus* makes placement of this zone at Granite Mountain equivocal with respect to the Mississippian–Pennsylvanian and mid-Carboniferous boundaries.

Dunn (1970b) and Lane and others (1985) have proposed that *Declinognathodus noduliferus* evolved from *Gnathodus girtyi simplex* in a Y-branch evolutionary sequence. At Granite Mountain, evidence of the evolutionary development of *D. noduliferus* from *G. girtyi simplex,* or from any other gnathodid species, is not present. The overall low abundance of gnathodid species is attributed to biofacies control. Based on combined occurrence data from all sections, it appears that *G. girtyi simplex* extends upward to directly below the lowest occurrence of *D. noduliferus.* An overlap in the ranges of these two species, as noted at the Arrow Canyon, Nevada, section by Lane and others (1985), was not observed at Granite Mountain.

Conodont Biofacies

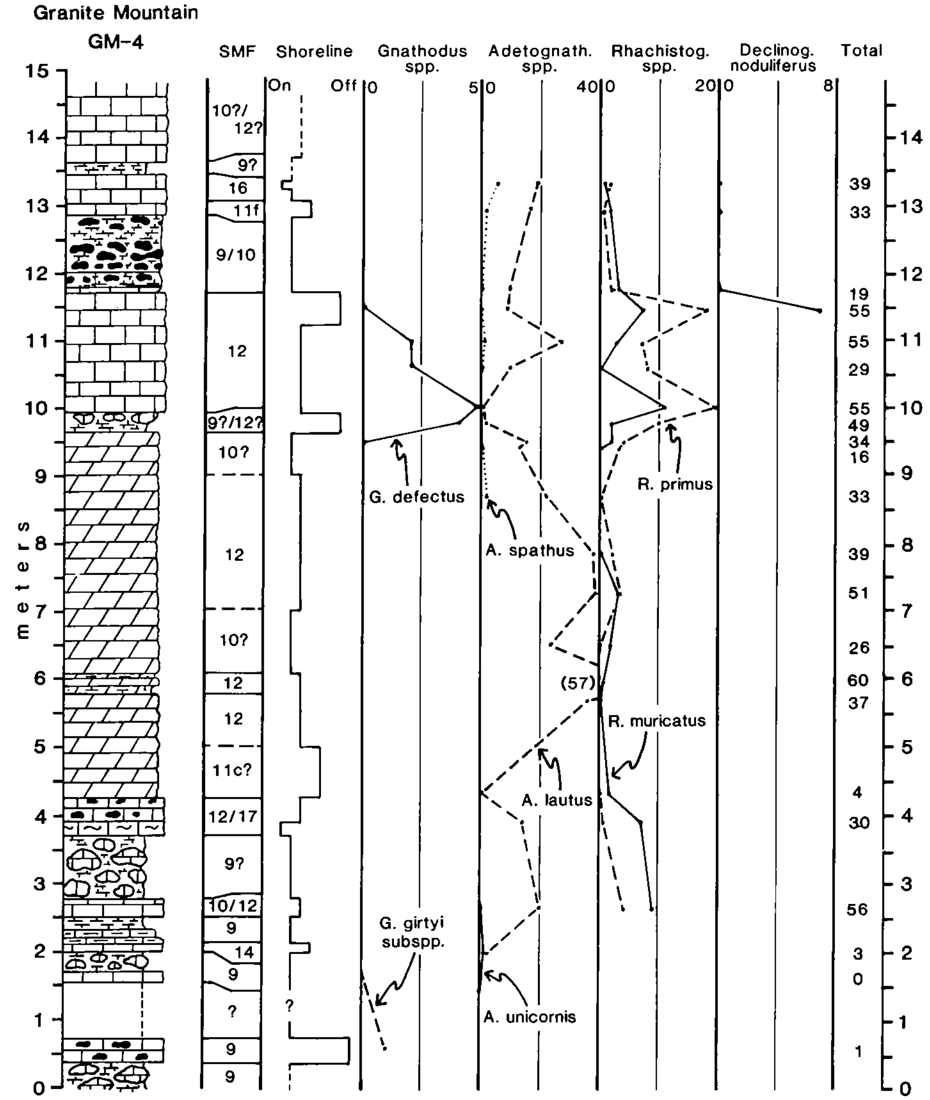
Conodont faunas recovered from the Mississippian–Pennsylvanian boundary strata at Granite Mountain are dominated by *Adetognathus* and *Rhachistognathus.* Plots of species abundances, ranges, and SMF Types show an alternating dominance of these two genera that reflect parallel, small-scale, repetitive lithofacies changes (Text-fig. 3). Overall, species of *Adetognathus* (and the closely allied genus *Cavusgnathus*) comprise 64% of the 4,200 total Pa elements recovered in the study. Species of *Rhachistognathus* comprise 31%, followed by species of *Gnathodus* which comprise 3%, and species of *Declinognathodus* that comprise 2%. The alternating dominance of *Adetognathus* and *Rhachistognathus* can be used to recognize two previously proposed conodont biofacies: the *Adetognathus* biofacies (Davis and Webster, 1985b; considered ecologically equivalent to the *Cavusgnathus* biofacies of Merrill, 1973; and Merrill and von Bitter, 1976) and the *Rhachistognathus* biofacies (Davis and Webster, 1985b). Detailed discussion of conodont species occurrences within individual SMF Types at Granite Mountain is given by Morrow and Webster (1991). Maximum numbers of both *Adetognathus* and *Rhachistognathus* occur in samples of the bioclast packstone/grainstone microfacies (SMF-12), where Pa element abundances are between 1.5 and 20 times greater than in other microfacies. In several samples, these relatively high conodont abundances were used to infer the presence of the bioclast packstone/grainstone microfacies in zones, where dolomitization has obscured or destroyed the original sediment fabric (e.g., Text-fig. 3, GM-4, 5.5–6.1 m interval). This procedure, when combined with other available petrographic data, provides a possible application of conodont biostratigraphy to limestone microfacies studies involving recurring lithologies that have been partially or totally obscured by dolomitization.

BRACHIOPOD BIOSTRATIGRAPHY

Previous reports on Mississippian–Pennsylvanian boundary brachiopod occurrences and zonations utilizing material from the Granite Mountain–Confusion Range area include the work of Bacon (1948), Sadlick (1965), and Gordon and others (1982, 1984,1985). Gordon and others (1982,1984,1985) have reported the occurrence of at least 48 species of brachiopods from the late Chesterian to early Morrowan beds at Granite Mountain. At the specific level, the majority of these taxa appear to be restricted both geographically and stratigraphically, and few were noted to cross the systemic boundary. In addition, many of these species are new, and have not been systematically described. In the present study, 26 species (from 19 genera) of those previously recorded by Gordon and others (1982, 1985) are tentatively recognized. The most abundant of these forms are illustrated in Plate 2.

The large majority of brachiopods recovered at Granite Mountain occur in two previously proposed brachiopod zones: the late Chesterian *Rhipidomella nevadensis* Zone (proposed by Sadlick, 1955) and the earliest Morrowan "*Rugoclostus* Zone" (informally proposed by Gordon and Duncan, 1970). At Granite Mountain, the "*Rugoclostus* Zone" is further refined to the *Rugoclostus* n. sp. B Zone, which is defined by an undescribed *Rugoclostus* species reported by Gordon and others (1984,1985). The lowest occurrence of *Rugoclostus* n. sp. B is between 0.5 and 1.0 m below the first occurrence of the conodont *Declinognathodus noduliferus* (Text-fig. 2), so in the strictest sense the base of the *Rugoclostus* n. sp. B Zone must be considered latest Chesterian in age.

Brachiopods occurring in the *Rhipidomella nevadensis* Zone at Granite Mountain include: *Anthracospirifer* cf. *A. occiduus, Buxtonia*? sp., *Cleiothyridina* sp., *Composita* spp. (2), *Crurithyris*?



Text-figure 3. Ranges and abundances of major conodont species, Granite Mountain section GM-4. All species and total abundance values are Pa elements per kg of sample. Total abundance values are based on all Pa elements recovered, including unidentified specimens and rarer species not shown on figure. "Shoreline" column shows inferred relative shoreline position, where "On" and "Off" correspond to onshore and offshore facies, respectively. "SMF" column shows Standard Microfacies Types discussed in text. Queried SMF Types are from dolostone or unsampled intervals. Lithologic symbols are the same as in Text-figure 2. Similar plots for Granite Mountain sections GM-1, GM-2, and GM-3 are given by Morrow and Webster (1991).

sp., *Flexaria*? sp., *Inflatia*? sp., *Punctospirifer transversus, Reticulariina gonionota, Rugoclostus* cf. *R.* n. sp. A (of Gordon and others, 1984, 1985), and *Spiriferellina campestris.* In the overlying *Rugoclostus* n. sp. B Zone, brachiopods occurring with the namebearer include: *Anthracospirifer* cf. *A. occiduus, A.* cf. *A. opimus, A.* cf. *A. birdspringensis, Antiquatonia* sp., *Buxtonia*? sp., *Cleiothyridina* sp., *Composita* sp., *Crurithyris*? sp., *Echinochonchus*? sp., *Hustedia* sp., *Inflatia*? sp., *Linoproductus*? sp., *Phricodothyris*? sp., *Punctospirifer transversus, Reticulariina gonionota, Rugoclostus* cf. *R.* n. sp. C (of Gordon and others, 1984,1985), *Spirifer* sp., and *Spiriferellina campestris.*

The *Rhipidomella nevadensis* and *Rugoclostus* n. sp. B Zones are separated by a 5.7–7.5-m-thick unzoned interval containing a low diversity fauna of disarticulated, broken, and partially silicified brachiopods dominated by species of *Anthracospirifer.* These species include: *A.* cf. *A. birdspringensis, A.* cf. *A. occiduus,* and *A.* cf. *A. opimus.* Also rarely present are *Cleiothyridina* sp., *Composita*? sp., *Reticulariina gonionota, Schizophoria* sp., and *Spiriferellina campestris. Anthracospirifer* cf. *A. birdspringensis* and *A.* cf. *A. opimus* both appear to have their first occurrence within the unzoned interval. It is possible that a future zonation could be established for the upper part of the unzoned interval, utilizing one of these species as the zone namebearer.

BLASTOIDS

Disarticulated, silicified specimens of the blastoid *Pentremites* were recovered from a conodont sample taken near the base of GM-3 (0.43 m) and from a float block of the same lithology found just below the base of GM-3. In addition, six complete *Pentremites* specimens were recovered from the lower portion of the Ely Limestone exposed on the western side of the Foote Range, located ~20 km southwest of Granite Mountain (SE¼SW¼NE¼ sec. 36, T. 15 S., R. 18 W., USGS Gandy, Utah, 15′ Quadrangle). Both the Granite Mountain and Foote Range samples occur within the uppermost part of the *Adetognathus unicornis* and *Rhipidomella nevadensis* Zones. These pentremitids (Pl. 2, Figs. 24,25) show flaring deltoid plates and shallow sulcate ambulacra that are indicative of Glen Dean or later (Namurian A) rocks in the Illinois basin (A. Horowitz, personal communication, 1988). They are assignable to *Pentremites cherokeeus* (= *P. crystalensis*) which occurs in the late Chesterian as high as the Kinkaid Limestone. In addition, one of the specimens (Pl. 2, Figs. 22,23) is relatively narrow and could be assignable to *P. laminatus*?, described from the late Chesterian Pitkin Limestone in Arkansas.

CONCLUSIONS

In agreement with Gordon and others (1983, 1984,1985), we consider deposition across the Mississippian–Pennsylvanian systemic boundary at Granite Mountain to have been essentially continuous. However, the boundary strata at Granite Mountain show numerous, small-scale changes in bed thickness and character when traced laterally. These variations are attributed primarily to facies changes and diagenetic alteration. Dolomite, which is locally abundant at Granite Mountain, is considered a post-depositional diagenetic feature altering an essentially normal marine precursor limestone. Sharp bedding contacts and coarse-grained, intraclast-rich grainstone microfacies give evidence of possible scouring or reworking between some units. These potential diastems are considered minor, below the limits of current lithostratigraphic and biostratigraphic resolution.

Mid-Carboniferous conodont and brachiopod zonations previously proposed for Granite Mountain are extended laterally to the sections of the present study with little or no modification. The important zone conodont *Declinognathodus noduliferus* is generally rare from the Granite Mountain sections. The low abundance of this form is attributed primarily to biofacies control, and the complete evolutionary sequence of *D. noduliferus* from its proposed ancestor *Gnathodus girtyi simplex* is not recognized at Granite Mountain. *Rhachistognathus primus* appears below *D. noduliferus* and does not mark the mid-Carboniferous boundary at Granite Mountain.

REFERENCES CITED

Bacon, C. S., Jr., 1948, Geology of the Confusion Range, west-central Utah: Geological Society of America Bulletin, v. 59, p. 1027–1052.

Davis, L. E.; and Webster, G. D., 1985a, The age of the Alaska Bench Formation, central Montana: Northwest Geology, v. 13, p. 1–4.

_____1985b, Late Mississippian to Early Pennsylvanian conodont biofacies in central Montana: Lethaia, v. 18, p. 67–72.

Dunn, D. L., 1970a, Conodont zonation near the Mississippian–Pennsylvanian boundary in western United States: Geological Society of America Bulletin, v. 81, p. 2959–2974.

_____1970b, Middle Carboniferous conodonts from western United States and phylogeny of the platform group: Journal of Paleontology, v. 44, p. 312–342.

Flügel, E., 1982, Microfacies analysis of limestones: Springer-Verlag, New York, 633 p.

Gordon, M., Jr.; and Duncan, H. M., 1970, Biostratigraphy and correlation, *in* Tooker, E. W.; and Roberts, R. J. (eds.), Upper Paleozoic rocks in the Oquirrh Mountains and Bingham Mining District, Utah: U.S. Geological Survey Professional Paper 629-A, p. A38–A70.

Gordon, M., Jr.; Henry, T. W. ; and Sutherland, P. K., 1982, Brachiopod zones delimiting the Mississippian–Pennsylvanian boundary in the United States, *in* Ramsbottom, W. H. C.; Saunders, W. B.; and Owens, B. (eds.), Biostratigraphic data for a mid-Carboniferous boundary: International Union of Geological Sciences, Subcommission on Carboniferous Stratigraphy (Symposium), Leeds, p. 69–76.

Gordon, M., Jr.; Henry, T. W., and Mamet, B. L., 1983, A sedimentologically continuous section across the Mississippian–Pennsylvanian boundary, Granite Mountain, Utah, U.S.A. [abstract]: 10^{e} Congres International de Stratigraphie et de Geologie du Carbonifere, Madrid, p. 360.

_____1984, Stop 3.—Granite Mountain section, *in* Lintz, J., Jr. (ed.), Western geological excursions, vol. 1: Geological Society of America and Affiliated Societies, Reno, Field-Trip Guidebook, p. 26–38.

_____1985, Carboniferous succession and Mississippian–Pennsylvanian boundary, Granite Mountain, Utah, U.S.A.: Compte Rendu, 10^{e} Congres International de Stratigraphie et de Geologie du Carbonifere, Madrid, 1983, v. 4, p. 441–450.

Hose, R. K., 1974, Geologic map of the Granite Mountain SW Quadrangle, Juab and Millard Counties, Utah: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-831.

Lane, H. R.; and Baesemann, J. F., 1982, A mid-Carboniferous boundary based on conodonts and revised intercontinental correlations, *in* Ramsbottom, W. H. C.; Saunders, W. B.; and Owens, B. (eds.), Biostratigraphic data for a mid-Carboniferous boundary: International Union of Geological Sciences, Subcommission on Carboniferous Stratigraphy (Symposium), Leeds, p. 6–12.

Lane, H. R.; and Manger, W. L., 1985, Toward a boundary in the middle of the Carboniferous (1975–1985): ten years of progress: Courier Forschungsinstitut Senckenberg, v. 74, p. 15–34.

Lane, H. R.; Baesemann, J. F.; Brenckle, P. L.; and West, R. R., 1985, Arrow Canyon, Nevada—A potential mid-Carboniferous boundary stratotype: Compte Rendu, 10^{e} Congres International de Stratigraphie et de Geologie du Carbonifere, Madrid, 1983, v. 4, p. 429–439.

Mamet, B. L., 1982, A report on algal microfacies at the Mississippian–Pennsylvanian boundary in the United States, *in* Ramsbottom, W. H. C.; Saunders, W. B.; and Owens, B. (eds.), Biostratigraphic data for a mid-Carboniferous boundary: International Union of Geological Sciences, Subcommission on Carboniferous Stratigraphy (Symposium), Leeds, p. 52–60.

______1984, Algal and foraminiferal biotas and microfacies, Mississippian–Pennsylvanian boundary, Granite Mountain section, *in* Lintz, J., Jr., (ed.), Western geological excursions, vol. 1: Geological Society of America and Affiliated Societies, Reno, Field-Trip Guidebook, p. 84–86.

Merrill, G. K., 1973, Pennsylvanian conodont paleoecology, *in* Rhodes, F. H. T. (ed.), Conodont paleozoology: Geological Society of America Special Paper 141, p. 239–274.

Merrill, G. K.; and von Bitter, P. H., 1976, Revision of conodont biofacies nomenclature and interpretations of environmental controls in Pennsylvanian rocks of eastern and central North America: Royal Ontario Museum, Life Sciences Contribution Number 108, 46 p.

Morrow, J. R.; and Webster, G. D., 1991, Carbonate microfacies and related conodont biofacies, Mississippian–Pennsylvanian boundary strata, Granite Mountain, west-central Utah: Brigham Young University Geology Studies, v. 37, p. 99–124.

Sadlick, W., 1955, Carboniferous formations of northeastern Uinta Mountains: Wyoming Geological Association Annual Field Conference Guidebook, p. 49–58.

______1965, Biostratigraphy of the Chainman Formation (Carboniferous), eastern Nevada and western Utah: University of Utah, Salt Lake City, unpublished Ph.D. dissertation, 228 p.

Sandberg, C. A.; Poole, F. G.; and Gutschick, R. C., 1980, Devonian and Mississippian stratigraphy and conodont zonation of Pilot and Chainman Shales, Confusion Range, Utah, *in* Fouch, T. D.; and Magathan, E. R. (eds.), Paleozoic paleogeography of the west-central United States, Rocky Mountain Paleogeography Symposium I: Society of Economic Paleontologists and Mineralogists, Rocky Mountain Section, p. 71–79.

Steele, G., 1960, Pennsylvanian–Permian stratigraphy of east-central Nevada and adjacent Utah, *in* Guidebook to the geology of east-central Nevada: Intermountain Association of Petroleum Geologists and Eastern Nevada Geological Society, p. 91–113.

Stevens, C. H.; and Armin, R. A., 1983, Microfacies of the Middle Pennsylvanian part of the Oquirrh Group, central Utah: Geological Society of America Memoir 157, p. 83–100.

Tynan, M. C., 1980, Conodont biostratigraphy of the Mississippian Chainman Formation, western Millard County, Utah: Journal of Paleontology, v. 54, p. 1282–1309.

Wardlaw, B. R., 1984, Conodont faunas, Mississippian–Pennsylvanian boundary, Granite Mountain section, *in* Lintz, J., Jr. (ed.), Western geological excursions, vol. 1: Geological Society of America and Affiliated Societies, Reno, Field-Trip Guidebook, p. 82–84.

Webster, G. D., 1984, Conodont zonations near the Mississippian–Pennsylvanian boundary in the eastern Great Basin, *in* Lintz, J., Jr. (ed.), Western geological excursions, vol. 1: Geological Society of America and Affiliated Societies, Reno, Field-Trip Guidebook, p. 78–82.

Webster, G. D.; Brenckle, P.; Gordon, M., Jr.; Lane, H. R.; Langenheim, R. L., Jr.; Sanderson, G. A.; and Tidwell, W. D., 1984a, The Mississippian–Pennsylvanian boundary in the eastern Great Basin: Compte Rendu, 9^{e} Congres International de Stratigraphie et de Geologie du Carbonifere, Washington D.C. and Champaign-Urbana, 1979, v. 2, p. 406–418.

Webster, G. D.; Gordon, M., Jr.; Langenheim, R. L.; and Henry, T. W.,1984b, The Mississippian–Pennsylvanian Boundary in the eastern Great Basin (Field Trip 1), *in* Lintz, J., Jr. (ed.), Western geological excursions, vol. 1: Geological Society of America and Affiliated Societies, Reno, Field-Trip Guidebook, p. 1–86.

Wilson, J. L., 1975, Carbonate facies in geologic history: Springer-Verlag, New York, 471 p.

PLATE 1

Locality designations indicate section number and distance above base of section (in meters).

Figure 1.—*Cavusgnathus unicornis* Youngquist and Miller, Pa element, lateral view; GM-3 1.70, USNM 448937, x50.

Figure 2.—*Gnathodus defectus* Dunn, Pa element, oral view; GM-1 7.40, USNM 448939, x70.

Figure 3.—*Gnathodus girtyi girtyi* Hass, Pa element, oral view; GM-2 0.90, USNM 448941, x70.

Figure 4.—*Gnathodus girtyi simplex* Dunn, Pa element, oral view; GM-3 1.70, USNM 448940, x50.

Figure 5.—*Declinognathodus noduliferus* (Ellison and Graves), Pa element, oral view; GM-4 11.49, USNM 448943, x40.

Figure 6.—*Rhachistognathus muricatus* (Dunn); Pa element, oral view; GM-3 3.81, USNM 448946, x60.

Figure 7.—*Rhachistognathus primus* Dunn, Pa element, oral view; GM-2 7.40, USNM 448949, x50.

Figure 8.—*Rhachistognathus websteri* Baesemann and Lane, Pa element, oral view; GM-3 14.79, USNM 448944, x70.

Figure 9.—*Rhachistognathus prolixus* Baesemann and Lane, Pa element, oral view; GM-3 11.53, USNM 448950, x50.

Figures 10,11.—*Adetognathus unicornis* (Rexroad and Burton), Pa element, oral and lateral views, respectively; GM-3 1.06, USNM 448916, x70.

Figure 12.—*Adetognathus lautus* (Gunnell), Pa element, oral view; GM-2 9.30, USNM 448918, x40.

Figure 13.—*Adetognathus spathus* (Dunn), Pa element, oral view; GM-3 14.79, USNM 448921, x40.

Figures 14–22.—Ramiform elements occurring with Pa elements of *Adetognathus lautus* and *Adetognathus spathus*. **14**, Pb element, lateral view; GM-3 15.27, USNM 448925. **15**, Sa element, lateral view; GM-1 9.74, USNM 448927. **16**, Sa element, posterior view; GM-3 15.42, USNM 448928. **17**, Sc element, posterior process; GM-1 9.33, USNM 448932. **18**, Sc element, posterior process; GM-1 9.33, USNM 448933. **19**, Sc element, lateral process and anterior cusp; GM-1 9.33, USNM 448934. **20**, M element, lateral view; GM-1 9.33, USNM 448931. **21**, M element, lateral view; GM-1 9.74, USNM 448930. **22**, M element, lateral view; GM-3 15.42, USNM 448929. All figures are x40.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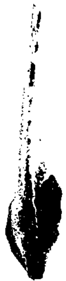
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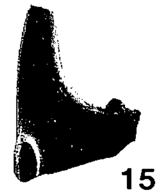
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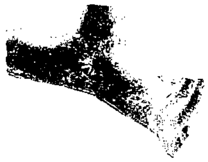
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PLATE 2

Locality designations indicate section number and distance above base of section (in meters).

Figures 1–3.—*Rhipidomella nevadensis* (Meek); pedicle valve exterior, anterior commissure, and lateral views, respectively; GM-3 3.31, USNM 448956, ×0.75.

Figures 4,5.—*Spiriferellina campestris* (White), pedicle valve exterior and interior views, respectively; GM-3 4.23, USNM 448959, x1.5.

Figures 6–8.—*Rugoclostus* cf. *R.* n. sp. B (Gordon and others, 1984,1985); pedicle valve exterior, lateral, and posterior views, respectively; GM-1 9.07, USNM 448957, ×0.75.

Figures 9,10.—*Reticulariina gonionota* (Meek); pedicle valve exterior and interior views, respectively; GM-3 11.37, USNM 448955, x1.5.

Figures 11–13.—*Rugoclostus* cf. *R.* n. sp. B (Gordon and others, 1984,1985); pedicle valve exterior, lateral, and posterior views, respectively; GM-3 12.87, USNM 448958, ×0.75.

Figure 14.—*Punctospirifer transversus* (McChesney); pedicle valve exterior; GM-3 12.11, USNM 448954, x1.5.

Figures 15–17.—*Anthracospirifer* cf. *A. opimus* (Hall); pedicle valve exterior, lateral, and interior views, respectively; GM-3 11.53; USNM 448951; x1.0.

Figure 18.—*Cleiothyridina* sp.; pedicle valve exterior; GM-3 3.31, USNM 448953, x1.0.

Figures 19–21.—*Anthracospirifer* cf. *A. occiduus* (Sadlick); pedicle valve exterior, pedicle and brachial valve lateral, and pedicle and brachial valve posterior views, respectively; GM-2 5.45, USNM 448952, x1.0.

Figures 22,23.—*Pentremites laminatus*? Easton; lateral and oral views, respectively; Foote Range locality, USNM 448961, x1.5.

Figures 24,25.—*Pentremites cherokeeus* Hall; lateral and oral views, respectively; Foote Range locality, USNM 448960, x1.5.


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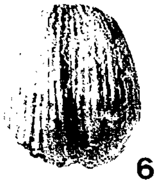
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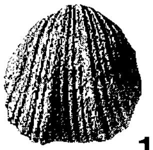
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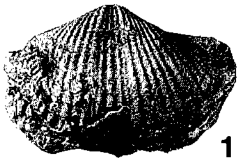
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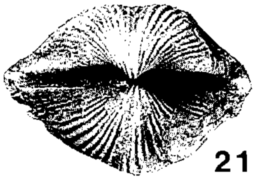
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Oklahoma Geological Survey Circular 94, 1992

Analysis of Sections Presumed to be Complete Across the Mississippian–Pennsylvanian Boundary, Southern Midcontinent

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ABSTRACT.—Cratonic settings now located in the southern Illinois basin, western Arkoma basin, Oklahoma and eastern Llano uplift, central Texas, have been proposed as sites of continuous deposition across the Mississippian–Pennsylvanian boundary even though that boundary is a major regional unconformity elsewhere in the southern Midcontinent. Analysis of the data supporting the presumed continuity identifies problems involving lithostratigraphic procedure (Illinois), biostratigraphic interpretation (Oklahoma), or the two in combination (Texas). Seemingly arbitrary invoking of conodont "biofacies" to explain dissimilar assemblages at the boundary further complicates its analysis. We appeal for use of *Occam's razor* in analyzing the Mississippian–Pennsylvanian boundary. The data better support an interpretation of the boundary as a major, regional truncating unconformity on the shelf areas of the southern Midcontinent.

INTRODUCTION

The Mississippian–Pennsylvanian boundary has been regarded as a significant regional unconformity across the North American Midcontinent for more than 100 years. That relationship illustrates the stratigraphic concept that cratonic sequences typically develop regional unconformities during cyclic transgression and regression by epieric seas. Theoretically, deposition would be expected to approach completeness only in deeper-water areas, either adjacent to the craton or in intracratonic basins where unconformities have diminished duration. Figure 1 illustrates the distribution of Mississippian and Pennsylvanian rocks in the southern Midcontinent and identifies those areas where duration of the unconformity separating the two systems is thought to be minimal or absent.

During the 1970s, lithologic facies and depositional modeling arguments were proposed as evidence for continuous deposition across the Mississippian–Pennsylvanian boundary on the eastern cratonic margin (see Ferm, 1974; Horne and others, 1974). That interpretation viewed the formations at the boundary as a contemporaneous facies succession from marine to continental environments transgressing westward. Therefore, the Pennsylvanian boundary simply marked any change from the marine conditions that had prevailed in the Mississippian, and thus, was both diachronous and meaningless in any stratigraphic sense. Although accepted by some, those proposals could not withstand a strong biostratigraphic challenge (e.g., Ettensohn and Peppers, 1979; Ettensohn, 1980), and by the end of that decade the controversy had mostly subsided.

During the past decade, biostratigraphic evidence of continuous deposition across the Mississippian–Pennsylvanian boundary in cratonic settings on the southern Midcontinent has presented a far more serious challenge to the traditional boundary interpretation. Localities in the southern Illinois basin, the western Arkoma basin (Oklahoma), and the eastern Llano region (central Texas) have yielded sequences of conodont assemblages cited as proving uninterrupted deposition in predominantly shale successions (Fig. 1). The situation is, however, equivocal. Differences in application of lithostratigraphic nomenclature have a profound effect on the precise positioning of the boundary and its depositional interpretation. Different conodont assemblages commonly flank the boundary, but those faunal differences may reflect paleoecologic or provincial controls.

The lithostratigraphic and conodont biostratigraphic data provided do not support the presumed depositional continuity across the Mississippian–Pennsylvanian boundary, which we contend is a pronounced regional unconformity on the southern North American craton. Figures 2 and 3

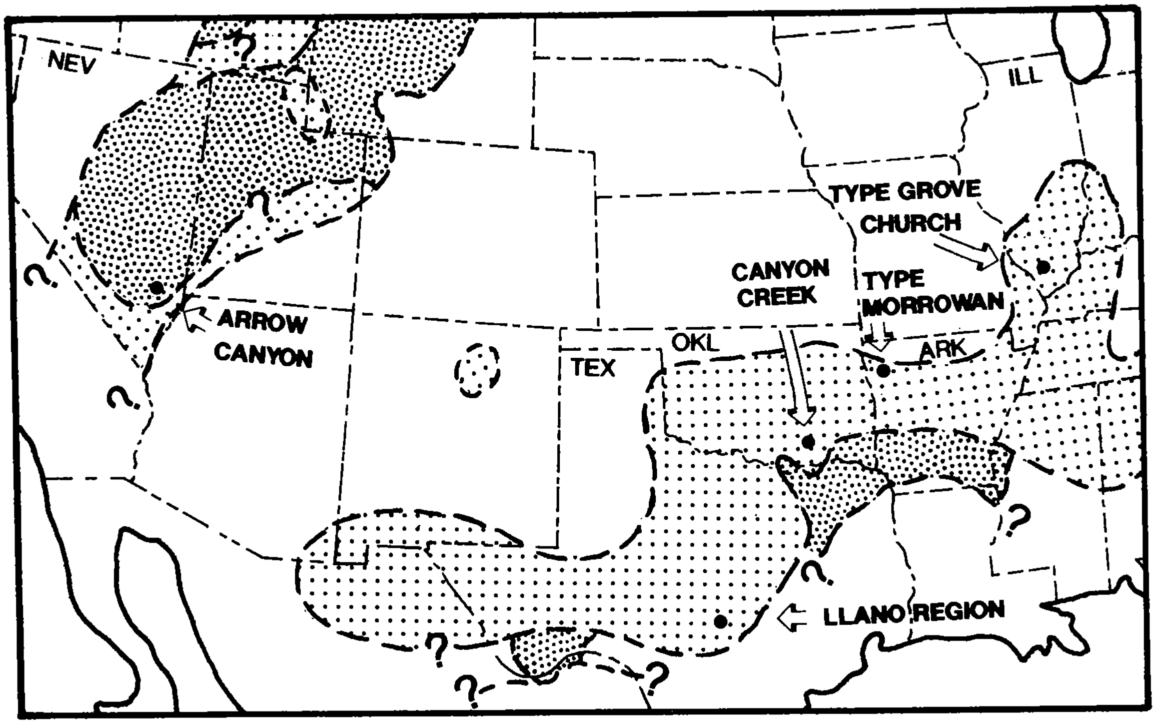

Figure 1. Map of southern Midcontinent showing where Morrowan strata rest on Chesterian strata (light stipple) and where deposition may be continuous across boundary (heavy stipple). Note that the sections presumed to be continuous across boundary (Type Grove Church, Canyon Creek, and Llano region), and discussed herein, occupy a cratonic shelf setting (modified from Sutherland and Manger, 1984, fig. 1).

illustrate varying biostratigraphic data and interpretations at the Mississippian-Pennsylvanian boundary for localities discussed herein.

Acknowledgments

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CONODONT BIOSTRATIGRAPHY ACROSS THE MISSISSIPPIAN–PENNSYLVANIAN BOUNDARY

Much of the controversy addressed in this paper results from lack of agreement on the conodont biostratigraphic succession across the Mississippian–Pennsylvanian boundary. Most sections in North America *are* incomplete at the boundary, and European zonations through presumably complete successions are not easily correlated with the former area (see Lane and Baesemann, 1982, for discussion of problem). Furthermore, paleoecologic controls on conodont distributions undoubtedly exist, and coeval assemblages of different elements, commonly called conodont biofacies (Merrill and Von Bitter, 1976), are well known in the literature. Understanding of both conodont biostratigraphy and paleoecology remains cursory. That situation and differing taxonomic philosophies accommodate the diverse interpretations of conodont sequences at the boundary analyzed herein.

Are there any complete conodont successions across the Mississippian–Pennsylvanian boundary? That question would seem to be manifestly affirmative in view of the recent decisions by the Subcommission on Carboniferous Stratigraphy (Lane and others, 1985a). Yet to our knowledge, there is no section in the world that has been universally embraced for its completeness. Nevertheless, for this paper, the Craven basin (Great Britain) and Arrow Canyon (Nevada, United States) will serve as standards for complete sections across the boundary.

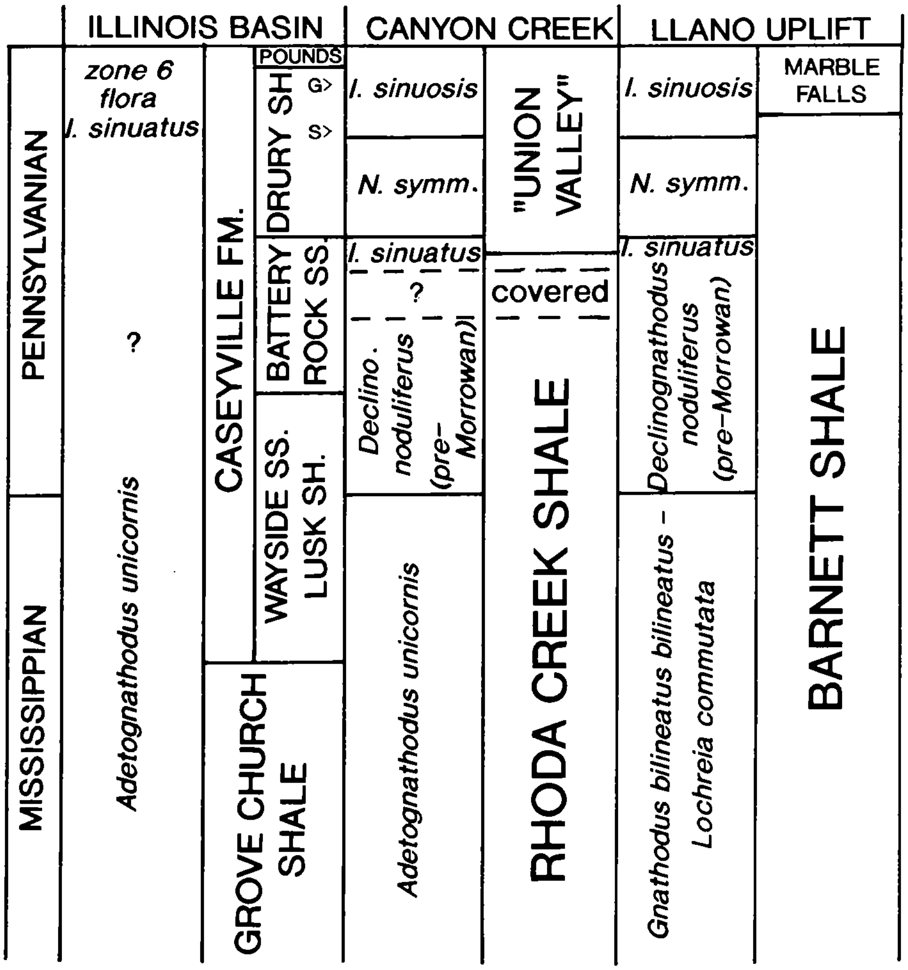

Figure 2. Lithostratigraphic succession and biostratigraphic data for sections presumed complete at the Mississippian–Pennsylvanian boundary in Illinois basin (from Rexroad and Merrill, 1985), Canyon Creek (from Grayson and others, 1985), and the Llano uplift (from Grayson and others, 1987). Conodont zonal taxa may be abbreviated; see text for further discussion. G> is position of Gentry Coal; S> is position of Sellers Limestone. (Our reinterpretation of these successions is shown in Fig. 3.)

The type Morrowan region in Arkansas (United States) will serve as an example of a succession with diminished, but, nevertheless, interrupted boundary sequences.

Craven Basin, Great Britain

The conodont biostratigraphy for the Late Viséan–early Westphalian Series was well documented by Higgins (1975) from the central and southern Pennine region of northern England. This broad anticlinal flexure reflects Hercynian tectonics and lies in the Craven basin (formerly Central Province basin), bounded by the Wales–Brahant Massif to the south and the Craven faults to the north (Ramsbottom, 1969; Ramsbottom and others, 1978). The Craven basin contains as much as 6,000 ft of middle Carboniferous deltaic sequences along its northern and southern margins and central basinal shales (Ramsbottom, 1969). Shelly, benthonic faunas occur in the marginal deltaic intervals, whereas the basinal facies yields only pelagic faunas, particularly ammonoids, conodonts, and pectinid bivalves. The section is cyclic, both in its faunal distribution (reflecting changing salinity and depth) and its lithologic succession (reflecting eustacy) (Ramsbottom, 1969,1977). Basin fill did not approach effective wave base until the Marsdenian Stage (R_2); thus, the Mississippian–Pennsylvanian boundary falls in the basinal Edale Shales, containing few limestones, essentially no sandstones, and a fauna of ammonoids and conodonts (Higgins, 1975). We believe that this area can serve as a model for conodont biostratigraphy in terrigenous-clastic deposi-

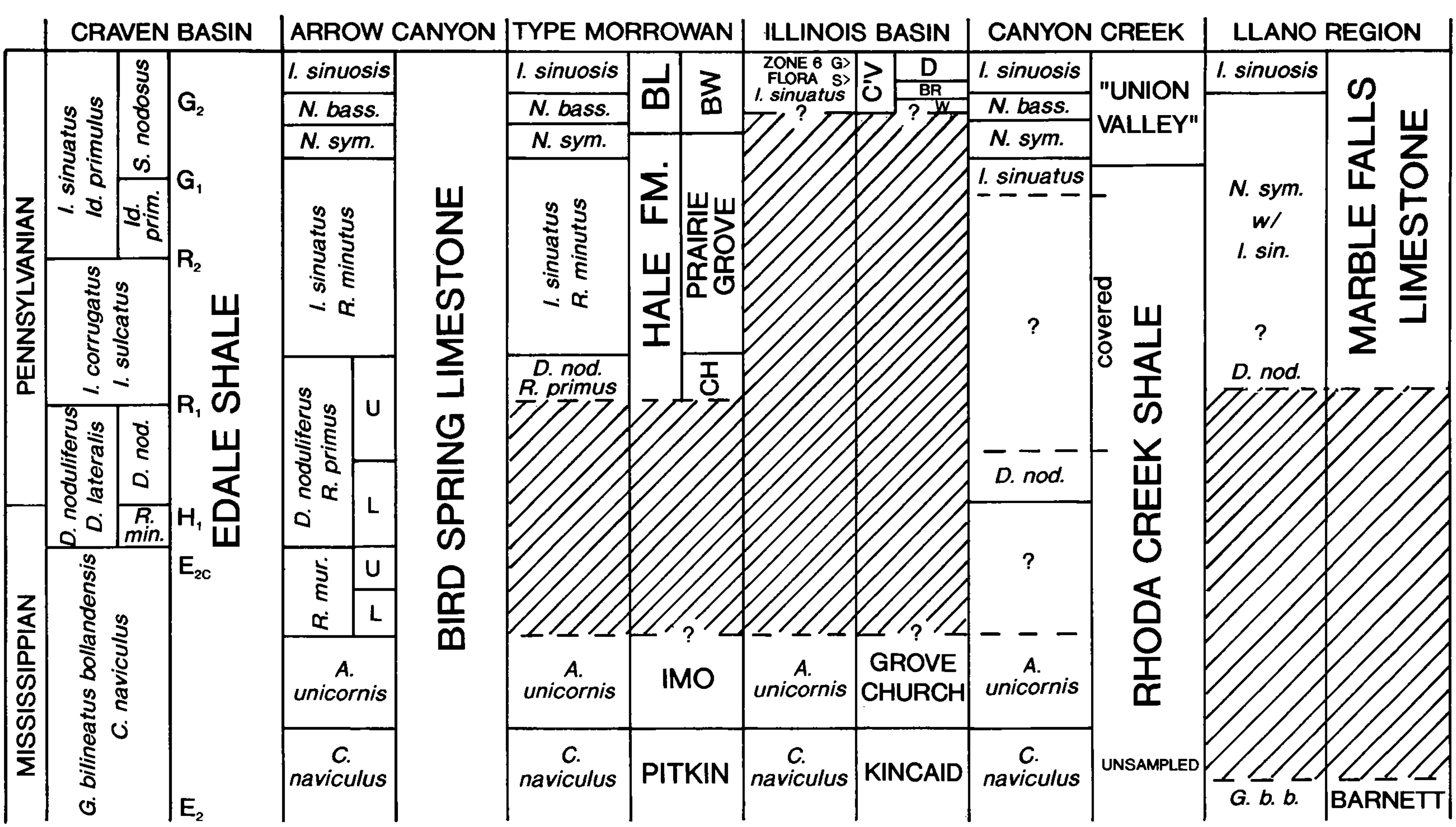

Figure 3. Reinterpretation of the correlation of sections in the Illinois basin, Canyon Creek, and Llano uplift with standard mid-Carboniferous boundary successions in the Craven basin (from Higgins, 1975, incorporating data from Riley and others, 1987), Arrow Canyon (from Baesemann and Lane, 1985), and type Morrowan region (from Lane and Straka, 1974) with modifications. Conodont zonal taxa may be abbreviated; see text for further discussion. G> is position of Gentry Coal; S> is position of Sellers Limestone. Lithostratigraphic abbreviations are: BL = Bloyd Formation, CH = Cane Hill Member, BW = Brentwood Member, C'V = Caseyville Formation, W = Wayside Shale, BR = Battery Rock Sandstone, D = Drury Shale.

tional regimes that are "deep" and/or oxygen depleted.

In the Craven basin (Higgins, 1975; Riley and others, 1987), youngest Mississippian assemblages comprise the *Gnathodus bilineatus bollandensis–Cavusgnathus naviculus* Zone and the overlying *Rachistognathus* (= *Idiognathoides*) *minutus* Subzone of the *Declinognathodus* (= *Idiognathoides*) *noduliferus–D.* (= *Stretognathodus*) *lateralis* Zone (original form-generic assignments are shown in parentheses, after presently accepted name, where important) (Higgins, 1975). Pa-elements comprise a small portion of the total assemblage, but the succession clearly shows that the E_{2a}–E_{2c} (upper Chesterian) interval is dominated in its lower part by *Gnathodus girtyi girtyi, G.g. intermedius, G. bilineatus bilineatus, Lochriea commutata,* and *Cavusgnathus naviculus,* succeeded by an upper assemblage of mostly *G.b. bollandensis. Rhachistognathus minutus* appears abruptly as the only Pa element in an improverished assemblage at the top of E_{2c} and base of H_{1a} (Riley and others, 1987). The Mississippian–Pennsylvanian boundary clearly falls within the *Declinognathodus noduliferus–D. lateralis Zone* (Higgins, 1975, p. 14–16 and fig. 3), at the base of the *D. noduliferus* Subzone indicated by the appearance of *Declinognathodus noduliferus noduliferus, D.n. inaequalis, D. lateralis* and *Neognathodus higginsi* (= *N. bassleri* of Higgins, 1975) slightly above the base of H_1 and beginning the Pennsylvanian Subsystem by international agreement (Lane and others, 1985b).

It has been recognized for nearly a decade that the ranges of most of the biostratigraphically useful taxa in Britain are markedly different when correlated with North America on the basis of ammonoids (Lane and Baesemann, 1982). A major unconformity has been proposed for the Craven basin, at the base of the *Rhachistognathus minutus* Subzone, that would accommodate the presumably missing upper Chesterian and Lower Morrowan conodont zones across the Mississippian–Pennsylvanian boundary (base of *Adetognathus unicornis* to base of *Neognathodus symmetricus*) (Lane and Baesemann, 1982; Lane and others, 1985a). Although that interpretation would resolve the problem with conodont correlations, it creates two other problems. No lithostratigraphic or lithologic evidence supports any regional unconformity at this level in the Craven basin, nor is there any ammonoid assemblage or succession of assemblages known anywhere in the world that could be viewed as representing the unconformable interval. *Homoceras* Zone (H_1–H_2) equivalents are known from the Ural Mountains and Donets basin, USSR, Bechar region, Algeria, most of western Europe, and now also from the United States (Nevada), in sequence with *Eumorphoceras* and *Cravenoceras* (E_1–E_2) below and *Reticuloceras* (R_1–R_2) above. Thus, there is no measurable unconformity in the Craven basin succession, and disparate conodont ranges (when compared with North America) must be the result of environmental regime and provincialism.

Arrow Canyon, Nevada

Conodont biostratigraphy across the Mississippian–Pennsylvanian boundary has been widely studied in Arrow Canyon because of its accessibility, presumed completeness, uniformity of its carbonate lithologic succession, and the sequence of its biostratigraphically sensitive fossil groups (see Langenheim and others, 1984, for a brief summary of early investigations).

The Arrow Canyon section was positioned on the eastern portion of a foreland basin situated on the eastern side of the Antler Orogenic Belt. Here, the Chesterian–Morrowan interval received at least 600 ft of chert-bearing shelf carbonates (typically with a high carbonate-mud content) that are referred to the Indian Springs and Bird Spring Formations (Webster and others, 1984). Terrigenous clastics are absent, except for a thin interval (~5 ft) just above the Mississippian–Pennsylvanian boundary within the Bird Spring Formation (Baesemann and Lane, 1985). The region was tectonically stable during mid-Carboniferous deposition, but eustatic cycles in its carbonate record produced attendant diastems. The boundary sequence is, nevertheless, thought to be complete in the shallow basinal setting represented by the Arrow Canyon section, but it becomes a pronounced unconformity eastward onto the craton (Webster and others, 1984). We believe that this section can serve as a model for carbonate depositional regimes that are moderately "shallow" with good oxygen exchange together with their associated conodont biostratigraphy.

Knowledge of the conodont biostratigraphy of the Mississippian–Pennsylvanian boundary at Arrow Canyon has developed over the past 20 years, principally through publications by Webster, 1969; Lane and others, 1972; and Baesemann and Lane, 1985. In spite of its prominence, however, the conodont biostratigraphy of Arrow Canyon has never been fully documented. The following discussion summarizes occurrence data and analyses presented by Baesemann and Lane (1985, particularly fig. 4). The base of the Indian Spring Formation falls in the *Cavusgnathus naviculus* Zone as that concept has developed from the type Chesterian Series of the Mississippi Valley region. That zone is succeeded by the *Adetognathus unicornis* Zone, which caps the type Chesterian Series, but there its top was not defined (Collinson and others, 1971). Lane and Straka (1974) restricted the zone to the range of its name-bearer below the appearance of *Rachistognathus muricatus.* At Arrow Canyon, the *A. unicornis* Zone assemblage also includes Pa elements such as *Gnathodus girtyi simplex, G. bilineatus bilineatus, G.b. bollandensis,* and *Cavusgnathus* spp., which are typical of the zone elsewhere in the United States. The lower *Rhachistognathus muricatus*

Zone of Baesemann and Lane (1985) begins with the appearance of the name-bearer and extends to the appearance of *A. lautus*. That zonal assemblage includes all earlier Pa elements from the *A. unicornis* Zone, but *G.b. bilineatus* and *G.b. bollandensis* do not occur in the highest portion of the lower *R. muricatus* Zone as it is developed at Arrow Canyon (note particularly Baesemann and Lane's samples 37–45). The upper *Rhachistognathus muricatus* Zone of Baesemann and Lane (1985) includes the overlapping ranges of the name-bearer and *A. lautus* below the appearance of *R. primus*. At Arrow Canyon, Pa elements of that zone include only *G.g. simplex* in addition to *A. lautus* and *R. muricatus*. *Rhachistognathus primus* and *Declinognathodus noduliferus* usher in the Pennsylvanian Subsystem along with *R. prolixus* and *R. websteri*, while *R. muricatus*, *G.g. simplex*, and *A. lautus* continue across the boundary from the upper *R. muricatus* Zone (sample 62). *Idiognathoides sinuatus* is the next zonal name-bearer to appear at Arrow Canyon (sample 89), but it is joined in the next sample (90) by zonal Pa elements *Idiognathoides convexus*, *Rachistognathus minutus* and *Neognathodus symmetricus* in contrast to expectation of a *I. sinuatus* Zone succeeded by an *N. symmetricus* Zone, as found in the type Morrowan region.

Type Morrow Region, Arkansas

Conodont biostratigraphy across the Mississippian–Pennsylvanian boundary in the type Morrowan region is well known through published studies, principally Lane and Straka (1974) and Mapes and Rexroad (1986). Many other contributions have utilized those data with attendant minor alterations in zonal names and element ranges, and the region has been a standard reference for nearly two decades. No one has ever questioned the presence of an unconformity at the Mississippian–Pennsylvanian boundary in this region, where Morrowan strata rest variously (ascending order) on the Fayetteville, Pitkin, or Imo Formations of Chesterian age.

The most reduced duration of the Mississippian–Pennsylvanian hiatus in the type Morrowan region would have the Imo Formation (Chesterian) overlain by the Cane Hill Member of the Hale Formation (Morrowan). The Imo Formation is a "shallow," terrigenous clastic interval dominated by shale with minor sandstones and carbonates. It contains abundant ammonoids indicating an Arnsbergian (E_{2b-c}) age, and an *Adetognathus unicornis* Zone conodont assemblage with *A. unicornis*, *Gnathodus bilineatus bilineatus*, *G.b. bollandensis*, *Lochriea commutata*, and *Kladognathus* (Mapes and Rexroad, 1986). Older formations (Fayetteville and Pitkin) belong to the *Cavusgnathus naviculus* Zone. The succeeding Cane Hill Member of the Hale Formation is a "shallow," tidally influenced succession of siltstone, sandstone, and shale with storm-deposited calcareous sandstone/sandy calcarenite lenses that are highly fossiliferous. The lower Cane Hill ammonoid assemblage may be correlated with the Kinderscoutian Series (R_1), and it is associated with a conodont assemblage including the Pa elements *Rhachistognathus primus*, *Declinognathodus noduliferus*, *Adetognathus lautus*, and *Idiognathoides sulcatus* (Lane and Straka, 1974), now referred to the *R. primus–D. noduliferus* Zone and also found at Arrow Canyon (Baesemann and Lane, 1985). Above that horizon, lower type Morrowan strata exhibit a conodont succession designated by the *I. sinuatus–R. minutus* and overlying *Neognathodus symmetricus* Zones (Lane and Straka, 1974; Baesemann and Lane, 1985). Correlation of the type Morrowan region with Arrow Canyon confirms the biostratigraphic break across the unconformable Mississippian–Pennsylvanian boundary, but seems to minimize the effects of gross lithologic control on conodont distributions.

ANALYSIS OF SOUTHERN MIDCONTINENT SECTIONS

Three localities (Fig. 1) recently have been reported as complete across the Mississippian–Pennsylvanian boundary in the southern Midcontinent. These are in the southern Illinois basin (Rexroad and Merrill, 1985); Canyon Creek in the western Arkoma basin, Oklahoma (Grayson and others, 1985; Grayson, 1990); and the Llano region, central Texas (Merrill, 1980; Grayson and others, 1987). Although depositional continuity is claimed for each case, these localities cannot be treated as a group since each represents a different set of lithostratigraphic and biostratigraphic problems, assumptions, and conclusions that have clouded the Mississippian–Pennsylvanian boundary interpretation.

Southern Illinois Basin

Rexroad and Merrill (1985) reported continuous deposition across the Mississippian–Pennsylvanian boundary in the southern Illinois basin, a "shallow," intracratonic basin (Fig. 1). There, middle Carboniferous deltaic, fluvial, and shallow-marine cycles accumulated to a thickness of ~1,900 ft. Conodonts representing the *Adetognathus unicornis* Zone were recovered from shales exposed at the closely approximate type areas for both the Grove Church Shale (highest type Chesterian) and Wayside Member, Caseyville Formation (Morrowan). Higher in the section (but still in the Caseyville), the Sellers Limestone bed of the Drury Shale Member, yielded a conodont assemblage with *Idiognathoides sinuatus* that is unquestionably Pennsylvanian. Since the Grove Church/Caseyville contact is the traditional Mississippian–Pennsylvanian boundary in the southern Illinois basin, depositional continuity across that boundary was claimed.

Assessment of these conodont assemblages is difficult since the number of samples, location, and abundances were not reported, and conodont preservation and potential reworking were not discussed. We concede the identification of the conodonts (although the identification of the Wayside *A. unicornis* is not accepted by Weibel and Norby, this volume) and acknowledge the evolving *Adetognathus* plexus. However, those points are irrelevant to the real problem. Our doubt concerning the proposed conformable relationship rests in its lithostratigraphic interpretation, which Rexroad and Merrill (1985) based on two lines of evidence: (1) the Grove Church/Caseyville lithologic succession is complete because it apparently fits both constructive delta and barrier-beach models; and (2) the beds in question above the Grove Church are unequivocally assignable to the Wayside Member of the Caseyville Formation, so the boundary must be continuous because the shales contain the same conodonts.

In terms of lithologic succession, we have not examined the sections in detail and reserve comment on the depositional succession, except to observe that repetition of the same environmental couplet as an explanation by Rexroad and Merrill for gross lithologic changes in the Grove Church/Caseyville is hardly an application of a depositional model and in no way proves continuity. In fact, it could be construed to favor interruptions, since environments appear to change abruptly. Analysis of the sandstone beds should identify which depositional model is applicable here and thus provide a better understanding of the Grove Church/Caseyville succession. Little description of the boundary interval is provided by Rexroad and Merrill (1985), especially the contact. Weibel and Norby (1990, and this volume) contend that obvious evidence exists for an unconformable Grove Church/Caseyville relationship in a core drilled near the type Grove Church. They cite a paleosol, abrupt contact, distinct mineralogic differences, and disparate environmental regimes on either side of the boundary as indicating a break before onset of Caseyville deltaic sedimentation. They could not replicate *Adetognathus unicornis* in the Wayside, but conodont recoveries were low. Finally, Weibel and Norby (this volume) also reassign the Wayside *Adetognathus* of Rexroad and Merrill to *A. lautus* morphotype A. However, that morphotype may be either Mississippian or Pennsylvanian.

Lithostratigraphic assignment of the shales to the Wayside at the sampled exposures is the critical element in evaluating this implied continuity across the boundary, yet it is questionable. The only evidence available at the type Grove Church section is that the first sandstone invariably has been assumed to be Pennsylvanian. In fact, the conodont evidence presented by Rexroad and Merrill (1985) might argue that the Grove Church encompasses beds above that point. The only evidence available in the type Wayside area is that this shale locality, located 1 mi away, is the "only good shale exposure in the immediate type area of the Wayside Sandstone Member of the Caseyville Formation" (Rexroad and Merrill, 1985, p. 51). The possibility that this shale exposure is either an erosional remnant or pinnacle on an irregular Mississippian surface or an erosional window through the Wayside is not discussed.

An unconformity in southern Illinois might still be indicated by the biostratigraphic data of Rexroad and Merrill (1985) and other sources. The Drury Shale Member of the Caseyville Formation contains the Sellers Limestone bed and overlying Gentry Coal. The Sellers Limestone contains conodonts identified by Rexroad and Merrill (1985) as *Idiognathoides sinuatus,* which can be no older than middle Morrowan (upper Prairie Grove to Dye Shale). The Gentry Coal, 66 ft above the probable Sellers Limestone at the type Caseyville section (Kosanke and others, 1960) contains a microflora that represents Read and Mamay (1964) Zone 6, precisely equivalent to that of the middle type Morrowan Baldwin Coal. Thus, undoubted middle Morrowan strata lie within <200 ft of the *Adetognathus unicornis* Zone reported in the supposed Wayside Member of the Caseyville Formation, suggesting a thin and possibly incomplete Morrowan section (see also Jennings and Fraunfelter, 1986, for concurring macropaleontological analyses of the Grove Church and Wayside assemblages).

The Mississippian–Pennsylvanian boundary is a pronounced unconformity throughout most of the Illinois basin (Rexroad and Merrill, 1985, p. 36), but we admit that continuous deposition across the boundary is possible in intracratonic basins. However, the case for that occurring at the Grove Church/Wayside contact in southern Illinois is an artifact of an unquestioned and unproven lithostratigraphic assignment. In the absence of any hard evidence to the contrary, we stand by our earlier analysis of the Mississippian–Pennsylvanian boundary as a pronounced unconformity in the type Chesterian region (Manger and Sutherland, 1984).

Canyon Creek, Oklahoma

Grayson and others (1985), Orth and others (1986), and Grayson (1990) reported an essentially conformable Mississippian–Pennsylvanian contact, both lithologically and biostratigraphically, along Canyon Creek, western Arkoma basin, Oklahoma (Fig. 1). During the Chesterian and Morrowan, the Canyon Creek locality occupied a "shallow" shelf setting, accumulating <800 ft of mostly terrigenous clastic sediment, not far from the margin of the "deeper" Ardmore basin where coeval strata exceed 4,500 ft in thickness. These rocks subsequently became tilted as part of the Arkoma structural basin during Atokan and later diastrophism (Sutherland, 1988). The boundary interval falls within the Rhoda

Creek Formation, which comprises variable shales that may be either black or gray, sandy to silty, phosphatic, concretionary, laminated, and intermittently covered (Grayson, 1990, fig. 3). The Mississippian–Pennsylvanian contact is placed at a lithologic change from silty, fissile to laminated, black shale which coincides with an elemental abundance anomaly, including the platinum group (Orth and others, 1986). Biostratigraphically, that horizon is marked by the appearance of *Declinognathodus noduliferus* above an assemblage with *Adetognathus unicornis*. *Rhachistognathus* is not reported, although a specimen figured as *D. noduliferus* (Grayson, 1990, pl. 1, fig. 9) may be *R. muricatus*.

The contention is made that the Canyon Creek appearance of *Declinognathodus noduliferus* is older than other North American appearances of that taxon, and that this zone is a biofacies of the upper *Rachistognathus muricatus* Zone, which is illustrated as spanning the Mississippian–Pennsylvanian boundary (Grayson and others, 1987, p. 155; fig. 8). Evidence for this biostratigraphic assignment of the Canyon Creek *D. noduliferus* is based presumably on its association with *Neognathodus higginsi*, a characteristically European taxon of H_1–H_2 age, as well as the typically Chesterian elements. Consequently, Grayson (1990) reported the basal Pennsylvanian section at Canyon Creek as pre-Morrowan, equivalent to the *D. noduliferus–D. lateralis* Zone of Higgins (1975), developed in the Craven basin, and not recognized previously in North America. The implication is left that the Canyon Creek section is complete across the boundary.

Canyon Creek is located in an area in which there are numerous well-documented Carboniferous unconformities including the truncation of the Wapanucka and Atoka Formations 1.5 and 0.5 mi, respectively, west of Canyon Creek (Morgan, 1924). The evaluation of boundary continuity at the Mississippian–Pennsylvanian boundary at Canyon Creek, however, must rest on the conodont biostratigraphy. Analysis is hampered by uncertainties related to conodont taxonomy and environmental controls or presumptions of control. The *Adetognathus unicornis* assemblage identified by Grayson and others (1985) and Grayson (1990) is no different from that of the Grove Church or Imo. It is also the same as that recovered from Arrow Canyon, except for the absence of *Rachistognathus*. The Rhoda Creek and Imo ammonoid assemblages are nearly identical as well (Saunders, 1973). Consequently, there is no doubt or confusion regarding the presence of late Chesterian strata at the boundary at Canyon Creek.

The succeeding beds contain unquestioned *Declinognathodus noduliferus* and are, therefore, Pennsylvanian by international agreement (Lane and others, 1985a). While the assemblage does not represent a typical *D. noduliferus–Rhachistognathus primus* Zone, it does not correspond exactly to the European situation either. The first appearance of *D. noduliferus* at Canyon Creek is in association only with *Lochriea commutata* through an interval of ~25 ft (Grayson and others, 1985, p. 156, samples 11–13). That occurrence is followed by horizons variously containing *D. noduliferus, A. lautus, G.* aff. *girtyi, L. commutata*, and *Neognathodus higginsi* through another 60 ft before a covered interval separates those beds from ones with typical North American Morrowan assemblages bearing *N. symmetricus* and *Idiognathoides sinuatus*. In Britain, *D. noduliferus* and *N. higginsi* join *R. minutus* simultaneously, with the absence of *A. lautus, G. girtyi*, and *L. commutata*, in the *D. noduliferus–D. lateralis* Zone.

We see no reason why Canyon Creek has to be interpreted in European biostratigraphic terms. The only anomaly from the North American view is *N. higginsi*, which may have been overlooked previously since Pa elements called *N. higginsi* include some morphotypes that could have been accommodated previously within the *Gnathodus girtyi* s.l. and *D. noduliferus–D. lateralis* s.l. groups, well known for their plasticity (Austin, 1972). Appeal to conodont "biofacies" to explain these assemblages and successions lacks rigor and becomes circular. If this is a European provincial assemblage or "biofacies," why is *A. unicornis* present and *R. minutus* absent? If it is a North American provincial assemblage, as we believe, then the succession here is probably discontinuous; *Rhachistognathus* equivalents (such as the appearance of *A. lautus* before *D. noduliferus*) should be developed. Splitting of the *D. noduliferus–D. lateralis* morphologic transition must be attempted with caution (see Austin, 1972) and tested rigorously before it is utilized for regional biostratigraphy, particularly in support of continuity at the Mississippian–Pennsylvanian boundary.

Llano Uplift, Central Texas

Assertions of Mississippian–Pennsylvanian boundary continuity at the Barnett–Marble Falls contact in the Llano region have been made sporadically for nearly 75 years. The history of this controversy and summaries of the current contrasting views may be found in Manger and others (1985) and Grayson and others (1987). This problem arises from a combination of lithostratigraphic practice and biostratigraphic interpretations.

The Barnett Shale is a slope-forming, black to light-brown, petroliferous, phosphatic shale with isolated concretions and beds of phosphatic packstones in its upper part. The Barnett is a condensed sequence representing anaerobic conditions in its lower part, but becoming more open marine toward its top. It was deposited in an extension of the Fort Worth basin, below effective wave base, but still in "shallow" conditions. Thicknesses range from a maximum near 45 ft in the eastern uplift, thinning westward to ~20 ft, until it is lost on the western margin of the Llano. There, it is considered coeval with a light-colored, crinoidal grainstone called informally the "White's Crossing Coquina."

The Marble Falls Limestone overlies the Barnett or older strata across the uplift. It develops a variety of carbonate facies and subordinate shales typical of open-shelf to "shallow" basinal environments and averages ~270 ft thick. The Marble Falls may be divided into informal lower and upper members by the presence of an intraformational unconformity. Westward across the uplift, that unconformity increases in duration, removing the lower member, so that the upper member rests on the Barnett, its equivalents, or older strata (Manger and others, 1985). There is no dispute regarding that unconformable relationship in the western Llano, but eastern Barnett–Marble Falls contacts are still regarded as continuous by some (Merrill, 1980; Grayson and others, 1987).

Lithostratigraphic arguments for a conformable Barnett–Lower Marble Falls contact generally center on the gradational, interbedded nature of two presumed lithosomes across the boundary and lack of any major physical evidence of erosion (i.e., conglomerates, truncation) at the contact, which may be a bedding plane. These conclusions generally ignore the phosphatic and glauconitic character of the Barnett shales and carbonates contrasted with those of the Marble Falls. No one has ever looked at the mineralogy of the shales in any detail, although there is an elemental anomaly in the lower Pennsylvanian section similar to that at Canyon Creek (Orth and others, 1986). Lithostratigraphic practice in the Llano region has arbitrarily placed the Barnett–Marble Falls contact at either the highest shale or lowest limestone in that part of the section. This practice places beds of markedly different age in the same formation and obscures the Mississippian–Pennsylvanian unconformity in a manner similar to that in the Illinois basin as discussed earlier.

Ignoring the lithostratigraphic problem for the moment, the conodont succession is characterized by a *Gnathodus bilineatus bilineatus–Lochreia commutata* assemblage succeeded by assemblages of unquestioned Pennsylvanian age. These include either *Declinognathodus noduliferus* (J. R. Walker Ranch Section; Grayson and others, 1987) or *Neognathodus symmetricus* elsewhere in the northeastern Llano, *Idiognathodus delicatus* in the southeastern Llano (type Marble Falls), or progressively younger assemblages (as young as lower Atokan) in the western uplift. Biostratigraphic continuity, however, is only claimed for the northeastern sections, particularly J. R. Walker Ranch.

The *Gnathodus bilineatus bilineatus–Lochreia commutata* assemblage is regarded as indicating a highest Chesterian horizon and a presumed "biofacies" comparable to the *G.b. bollandensis–Cavusgnathus naviculus* Zone of the Craven basin and the *Adetognathus unicornis* and *Rhachistognathus muricatus* Zones of Arrow Canyon (Grayson and others, 1987, fig. 8). Yet none of those zonal name-bearers have been recovered with the Llano assemblage, which is dominated by unequivocal *G.b. bilineatus*. Age assignment based on the actual elements recovered, particularly *G.b. bollandensis*, could be no younger than the lower portions of the *R. muricatus* and *G.b. bollandensis–C. naviculus* Zones based on ranges reported from Britain and southern Nevada, respectively. The Llano conodont assemblage occurs with ammonoids such as *Eumorphoceras bisulcatum* and *Richardsonites*, which can be no younger than the lower Arnsbergian Stage (E_{2a}) (Miller and Youngquist, 1948; Saunders and others, 1977). In the absence of any taxa suggesting a younger age, we regard the *G.b. bilineatus* assemblage as middle Chesterian. That assemblage is succeeded by *Declinognathodus noduliferus* associated with *Neognathodus higginsi* and several species of *Rhachistognathus*. Analysis of that assemblage evokes the same questions raised for Canyon Creek, particularly the contention that it is pre-Morrowan (Grayson and others, 1987). Nevertheless, we see no evidence to suggest anything other than lower Pennsylvanian resting on middle Chesterian, and thus an unconformable relationship.

The Llano situation is further complicated by lithostratigraphic practice. Shales at the boundary are generally called Barnett, but may contain either the Chesterian *Gnathodus bilineatus bilineatus–Lochreia commutata* assemblage or the Morrowan *Declinognathodus noduliferus* or a younger assemblage. In all cases, there is a break in the conodont succession as we have described. Arbitrary placement of the Marble Falls contact at the highest shale is most commonly followed and seemingly supports continuous deposition across the Mississippian–Pennsylvanian boundary just as it does in the southern Illinois basin. That practice, however, forces a major intraformational unconformity into the Barnett. It seems far more logical to us to view the basal Pennsylvanian transgression as redepositing terrigenous clastic sediments derived from the Barnett during pre-*D. noduliferus* or younger erosion. Initiation of carbonate deposition followed clearing of the water column as the transgression progressed and was highly diachronous across the uplift.

CONCLUSIONS

The *reductio ad absurdum* of this boundary controversy in the southern Midcontinent would be that the southern Ozark succession is complete (*Adetognathus unicornis* spans the Mississippian–Pennsylvanian boundary and is succeeded by *Declinognathodus noduliferus*) or that there is a major gap in the Arrow Canyon section (pre-Morrowan *Declinognathodus noduliferus* assemblages are not present). We hold both situations as untenable. Close examination of the three sections claimed to be complete across the boundary suggests that those claims are based on a lithostratigraphic (and perhaps taxonomic) procedural problem in the Illinois basin, a biostratigraphic interpretation problem at Canyon Creek, and the two problems together in the Llano region.

The depositional, structural, and provincial setting of each of these southern Midcontinent localities is basically the same. The Mississippian–Pennsylvanian boundary is a major truncating unconformity in the type Morrowan region which also occupies the same setting. It is inconceivable to us that such an event would not be recorded in these other closely adjacent cratonic settings in the southern Midcontinent. Our reading of the evidence presented suggests that it is.

REFERENCES CITED

Austin, R. L., 1972, Problems of conodont taxonomy with special reference to Upper Carboniferous forms: Geologica et Palaeontologica JB1, p. 115–126.

Baesemann, J. F.; and Lane, H. R., 1985, Taxonomy and evolution of the genus *Rachistognathus* Dunn (Conodonta: Late Mississippian to Early Middle Pennsylvanian) *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg 74, p. 93–136.

Collinson, C.; Rexroad, C. B.; and Thompson, T. L., 1971, Conodont zonation of the North American Mississippian *in* Sweet, W. C.; and Bergström, S. M. (eds.), Symposium on conodont biostratigraphy: Geological Society of America Memoir 127, p. 353–394.

Ettensohn, F. R., 1980, An alternative to the barrier-shoreline model for deposition of Mississippian and Pennsylvanian rocks in northeastern Kentucky: Geological Society of America Bulletin, v. 91, pt. 1, p. 130–135; pt. 2, p. 934–1056.

Ettensohn, F. R.; and Peppers, R. A., 1979, Palynology and biostratigraphy of Pennington shales and coals (Chesterian) at selected sites in northeastern Kentucky: Journal of Paleontology, v. 53, p. 453–474.

Ferm, J. C., 1974, Carboniferous environmental models in eastern United States and their significance, *in* Briggs, G. (ed.), Carboniferous of southeastern United States: Geological Society of America Special Paper 148, p. 79–95.

Grayson, R. C., Jr., 1990, Canyon Creek: a significant exposure of a predominantly mudrock succession recording essentially continuous deposition from the Late Devonian through the Middle Pennsylvanian, *in* Ritter, S. M. (ed.), Early to middle Paleozoic conodont biostratigraphy of the Arbuckle Mountains, southern Oklahoma: Oklahoma Geological Survey Guidebook 27, p. 85–105.

Grayson, R. C., Jr.; Davidson, W. T.; Westergaard, E. H.; Atchley, S. C.; Hightower, J. H.; Monaghan, P. T.; and Pollard, C., 1985, Mississippian–"Pennsylvanian" (mid-Carboniferous) boundary conodonts from the Rhoda Creek Formation: *Homoceras* equivalent in North America, *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg 74, p. 149–180.

Grayson, R. C., Jr.; Merrill, G. K.; and Miller, J. F., 1987, Early and Late Paleozoic conodont faunas of the Llano uplift region, central Texas—biostratigraphy, systemic boundary relationships, and stratigraphic importance: Geological Society of America, South-Central Section, Guidebook for Field Trip 1, Waco, Texas, 152 p.

Higgins, A. C., 1975, Conodont zonation of the late Viséan–early Westphalian strata of the south and central Pennines of northern England: Geological Survey of Great Britain Bulletin 53, 90 p.

Horne, J. C.; Ferm, J. C.; and Swinchatt, J. P., 1974, Depositional model for the Mississippian–Pennsylvanian boundary in northeastern Kentucky, *in* Briggs, G. (ed.), Carboniferous of southeastern United States: Geological Society of America Special Paper 148, p. 97–114.

Jennings, J. R.; and Fraunfelter, G. H., 1986, Preliminary report on macropaleontology of strata above and below the upper boundary of the type Mississippian: Transactions of the Illinois State Academy of Sciences, v. 78, p. 253–261.

Kosanke, R. M.; Simon, J. A.; Wanless, H. R.; and Willman, H. B., 1960, Classification of Pennsylvanian strata of Illinois: Illinois Geological Survey Report of Investigations 214, 84 p.

Lane, H. R.; and Baesemann, J. F., 1982, A mid-Carboniferous boundary based on conodonts and revised intercontinental correlation, *in* Ramsbottom, W. H. C.; Saunders, W. B.; and Owens, B. (eds.), Biostratigraphic data for a mid-Carboniferous boundary, Subcommission on Carboniferous Stratigraphy, Leeds, p. 6–12.

Lane, H. R.; and Straka, J. J., II, 1974, Late Mississippian and Early Pennsylvanian conodonts, Arkansas and Oklahoma: Geological Society of America Special Paper 152, 144 p.

Lane, H. R.; Sanderson, G. A.; and Verville, G. J., 1972, Uppermost Mississippian–basal Middle Pennsylvanian conodonts and fusulinids from several exposures in the south-central and southwestern United States: Proceedings of the 24th International Geological Congress, Montreal, Section 7, Paleontology, p. 549–555.

Lane, H. R.; Bouckaert, J.; Brenckle, P. L.; Einor, O. L.; Havlena, V.; Higgins, A. C.; Yang, J. Z.; Manger, W. L.; Nassichuk, W. W.; Nemirovskaya, T.; Owens, B.; Ramsbottom, W. H. C.; Reitlinger, E. A.; and Weyant, M., 1985a, Proposal for an international mid-Carboniferous boundary: Compte Rendu, Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, Madrid, 1983, v. 4, p. 323–339.

Lane, H. R.; Baesemann, J. F.; and Groves, J. R., 1985b, Is the base of the *Reticuloceras* zone a reliably recognizable biostratigraphic level?, *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg 74, p. 137, 148.

Langenheim, R. L., Jr.; Webster, G. D.; and Weibel, C. P., 1984, Atokan rocks of the Bird Spring Group, Arrow

Canyon, Clark County, Nevada, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 133–156.

Manger, W. L.; and Sutherland, P. K., 1984, The Mississippian–Pennsylvanian boundary in the southern Midcontinent, United States: Compte Rendu, Neuvième Congrès International de Stratigraphie et de Géologie du Carbonifère, Champaign-Urbana, 1979, v. 2, p. 369–376.

Manger, W. L.; Sutherland, P. K.; Johnson, D. W.; and Brown, C. M., 1985, Biostratigraphy of the Mississippian–Pennsylvanian boundary, Llano region, central Texas: Compte Rendu, Dixième Congrès International de Stratigraphie et de Géologie du Carbonifère, Madrid, 1983, v. 2, p. 363–369.

Mapes, R. H.; and Rexroad, C. B., 1986, Conodonts from the Imo Formation (upper Chesterian), north-central Arkansas: Geologica et Palaeontologica, v. 20, p. 113–124.

Merrill, G. K., 1980, Geology of the Llano region, central Texas: West Texas Geological Society Guidebook 80-73, 246 p.

Merrill, G. K.; and Von Bitter, P. H., 1976, Revision of conodont biofacies nomenclature and interpretations of environmental controls in Pennsylvanian rocks of eastern and central North America: Life Sciences Contributions, Royal Ontario Museum, No. 108, 46 p.

Miller, A. K.; and Younquist, W., 1948, The cephalopod fauna of the Mississippian Barnett Formation of central Texas: Journal of Paleontology, v. 22, p. 649–671.

Morgan, G. D., 1924, Geology of the Stonewall Quadrangle, Oklahoma: Oklahoma Bureau of Geology Bulletin 2, 248 p.

Orth, C. J.; Quintana, L. R.; Gilmore, J. S.; Grayson, R. C., Jr.; and Westergaard, E. H., 1986, Trace-element anomalies at the Mississippian/Pennsylvanian boundary in Oklahoma and Texas: Geology, v. 14, p. 986–990.

Ramsbottom, W. H. C., 1969, The Namurian of Britain: Sextieme Congrés International de Stratigraphie et de Géologie du Carbonifère, Sheffield, 1967, v. 1, p. 219–232.

______1977, Major cycles of transgression and regression (Mesotherms) in the Namurian: Proceedings, Yorkshire Geological Society, v. 41, p. 261–291.

Ramsbottom, W. H. C.; Calver, M. A.; Eagar, R. M. C.; Hodson, F.; Holliday, D. W.; Stubblefield, C. J.; and Wilson, R. B., 1978, A correlation of Silesian rocks in the British Isles: Geological Society of London Special Report 10, 81 p.

Read, C. B.; and Mamay, S. H., 1964, Upper Paleozoic floral zones and floral provinces of the United States: U.S. Geological Survey Professional Paper 454-K, p. 1–35.

Rexroad, C. B.; and Merrill, G. K., 1985, Conodont biostratigraphy and paleoecology of middle Carboniferous rocks in southern Illinois, *in* Lane, H. R.; and Ziegler, W. (eds.), Toward a boundary in the middle of the Carboniferous: stratigraphy and paleontology: Courier Forschungsinstitut Senckenberg 74, p. 35–64.

Riley, N. J.; Varker, W. J.; Owens, B.; Higgins, A. C.; and Ramsbottom, W. H. C., 1987, Stonehead Beck, Cowling, North Yorkshire, England: a British proposal for the mid-Carboniferous boundary stratotype, *in* Brenckle, P. L.; Lane, H. R.; and Manger, W. L. (eds.), Selected studies in Carboniferous paleontology and biostratigraphy: Courier Forschungsinstitut Senckenberg 98, p. 159–178.

Saunders, W. B., 1973, Upper Mississippian ammonoids from Arkansas and Oklahoma: Geological Society of America Special Paper 145, 110 p.

Saunders, W. B.; Manger, W. L.; and Gordon, M., Jr., 1977, Upper Mississippian and lower and middle Pennsylvanian ammonoid biostratigraphy of northern Arkansas, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, p. 117–137.

Sutherland, P. K., 1988, Late Mississippian and Pennsylvanian depositional history in the Arkoma basin area, Oklahoma and Arkansas: Geological Society of America Bulletin, v. 100, p. 1787–1802.

Sutherland, P. K.; and Manger, W. L., 1984, The Mississippian–Pennsylvanian boundary in North America, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Ninth International Congress on Carboniferous Stratigraphy and Geology, v. 2, p. 319–329.

Webster, G. D., 1969, Chester through Derry conodonts and stratigraphy of northern Clark and southern Lincoln Counties, Nevada: University of California Publications in the Geological Sciences, v. 79, 105 p.

Webster, G. D.; Brenckle, P. L.; Gordon, M., Jr.; Lane, H. R.; Langenheim, R. L., Jr.; Sanderson, G. A.; and Tidwell, W. D., 1984, The Mississippian–Pennsylvanian boundary in the eastern Great basin: Compte Rendu, Neuvième Congrès International de Stratigraphie et de Géologie du Carbonifère, Champaign-Urbana, 1979, v. 2, p. 406–418.

Weibel, C. P.; and Norby, R. D., 1990, Sedimentologic and micropaleontologic study of a core through the Mississippian–Pennsylvanian boundary in southern Illinois: Geological Society of America Abstracts with Programs, v. 22, p. 35.

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Morrowan and Atokan (Pennsylvanian) Biostratigraphy in the Ardmore Basin, Oklahoma

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ABSTRACT.—The Pennsylvanian stratigraphic sequence in the Ardmore basin reflects a complex depositional history. Deposition occurred contemporaneously with deformation resulting in rapid lateral and vertical changes in facies and in direction of sediment source. Scattered units are fossiliferous, but faunal as well as lithologic correlations are generally difficult.

Occurrences of fusulinids, conodonts, goniatites, and brachiopods are synthesized to evaluate the positions of the Chesterian–Morrowan, Morrowan–Atokan, and Atokan–Desmoinesian boundaries. Goniatites are useful only in the lower portion and brachiopod occurrences are scattered. The fusulinid sequence is incomplete, lacking *Eoschubertella*, *Profusulinella*, and primitive species of *Fusulinella*. Included are advanced forms of *Fusulinella* and species of *Fusulina*. Ten tentative conodont associations have been recognized in the Morrowan and Atokan interval, and some of these have potential for both intra- and interbasinal correlation.

The Chesterian–Morrowan boundary, based on rare occurrences of the conodont *Declinognathodus*, falls in the upper part of the unnamed shale between the Overbrook and Lake Ardmore Members of the Springer Formation.

The Morrowan–Atokan boundary, based on fusulinid occurrences, falls at a probable unconformity either at the base of the Bostwick or in the covered part of Unnamed Unit 2.

The Atokan–Desmoinesian boundary, based on a combination of fusulinid, conodont, and brachiopod data, most likely occurs in the upper part of Unnamed Unit 3, below the Lester Limestone Member (Lake Murray Formation).

INTRODUCTION

Scope of Investigation

The main faunal elements in the Morrowan and Atokan intervals in the Ardmore basin are fusulinids, conodonts, brachiopods, and ammonoids; they form a significant basis for understanding the depositional history of the area and for the correlation of various stratigraphic intervals with other basins. This paper describes the biostratigraphic framework for the Ardmore basin and provides preliminary documentation of the marked vertical and horizontal changes in facies, depositional regimes, and faunal distributions that characterize the Morrowan–Atokan record in the region.

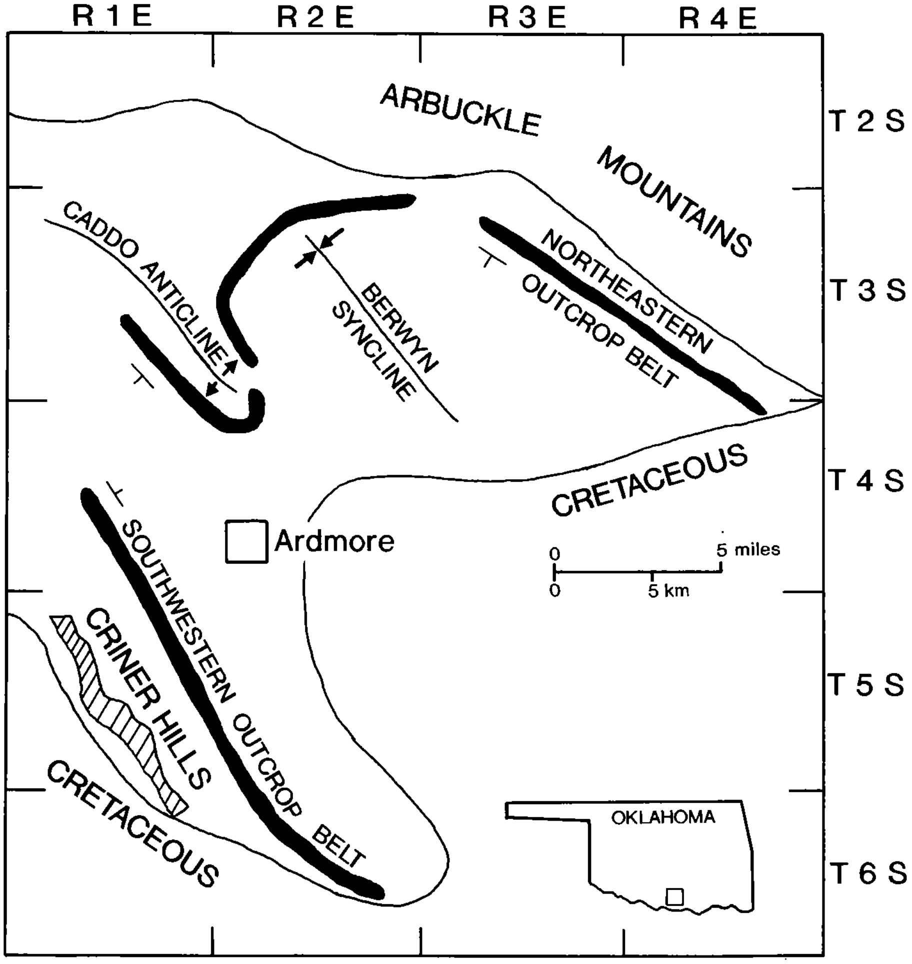
Geological Setting

The Ardmore basin is an elongate, northwest-trending feature that lies between the Arbuckle uplift to the northeast and the Wichita–Criner Hills uplift to the southwest (Text-fig. 1). The Ardmore basin was originally a part of the southern Oklahoma aulacogen, a much larger northwest-trending trough extending through southwest Oklahoma (Ham and others, 1964). The present-day structural belt that represents the aulacogen consists of several northwest-trending uplifts and basins. The Anadarko, Ardmore, and Marietta basins and the Arbuckle, Criner, and Wichita uplifts are the result of a series of Pennsylvanian orogenic events that occurred during the final deformational stage in the history of the aulacogen.

Tectonic history of the Ardmore basin is complex. Deposition occurred contemporaneously with deformation resulting in some important changes in direction of sediment source that are reflected in the lithostratigraphy of the region (Text-fig. 2). These can be outlined as follows:

1) Sandstones of the Springer Formation, beginning in Chesterian time, were derived primarily from the northwest and transported southeastward along the axis of the aulacogen (Meek and others, 1988).

2) The Springer depositional pattern was partly disrupted by uplift along the Wichita–Criner Hills axis, which differentiated the Ardmore and Marietta basins (Ham and Wilson, 1967). It is generally stated that this tectonic pulse began in early Penn-



Text-figure 1. Index map showing distribution of strata of Morrowan and Atokan age in the Ardmore basin.

sylvanian time (Ham and Wilson, 1967), but, in fact, it began in late Mississippian (Chesterian) time, at least as early as the deposition of the lower or middle part of the Springer Formation. Deformation and the subsequent erosion of the Criner uplift must have been rapid since Ordovician limestone pebbles occur in two thin Springer conglomerates that contain Chesterian conodonts in their matrix.

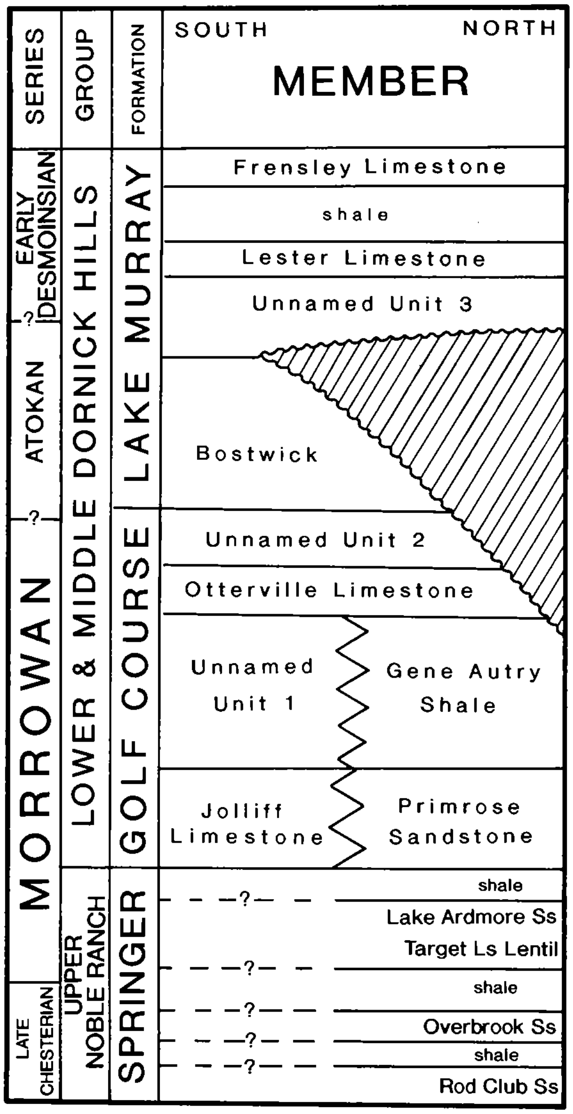
3) More pronounced uplift took place in Morrowan time, which led to the deposition in the southwestern part of the basin of extensive limestone-cobble conglomerates, composed mostly of Ordovician clasts (Jolliff Member), followed by even more profound uplift during the Atokan (Bostwick Member). Facies changes within Morrowan and Atokan strata demonstrate that the Caddo anticline and Berwyn syncline were being slowly uplifted and downwarped during these same events.

4) By Middle Desmoinesian time, chert-pebble conglomerates of the Devils Kitchen Member of the Deese Group (not included in this paper) may reflect the initial effects of uplift of the Ouachita tectogene to the southeast.

Previous Investigations

Tomlinson and McBee (1959) provide the only comprehensive treatment of Ardmore basin Carboniferous stratigraphy and depositional history. Faunal studies are also limited. Waddell (1966) established a Middle and Late Pennsylvanian fusulinid zonation that has been the principal basis for chronostratigraphic subdivision of this interval in the Ardmore basin. Straka (1972) and Lane and Straka (1974) investigated Late Mississippian and Early Pennsylvanian conodonts from the Springer Formation in the northern portion of the basin, and they applied a conodont zonation to the interval.

Much new Carboniferous stratigraphic and biostratigraphic data have been generated by graduate students at the University of Oklahoma working under the direction of P. K. Sutherland. Fourteen master's theses have been completed in as many



Text-figure 2. Table of formations and members.

years. The present authors are currently attempting to integrate all available information for a planned synthesis of the Carboniferous depositional history in the basin. As part of this larger investigation, R. C. Grayson, Jr., is studying the character and distribution of Pennsylvanian conodonts through the sequence.

Acknowledgments

We thank Russel A. Peppers, Illinois Geological Survey, for having processed and studied samples that we sent to him from a thin coal in Measured Section 363. We are grateful to Thomas W. Henry and William W. Clopine for helpful critical reviews of an earlier version of this paper.

LITHOSTRATIGRAPHY

Lower and Middle Pennsylvanian strata are exposed in two generalized areas, north and south of Ardmore, shown on Text-figure 1. Rocks in these areas are composed predominantly of shale, but include much thinner, mostly lenticular layers of synorogenic conglomerate, limestone, and sandstone. Most intervals show marked facies changes within each outcrop area, but the changes are particularly pronounced between the southwestern and northeastern outcrop belts (Text-fig. 1). Text-figures 2 and 3 summarize lithostratigraphic and biostratigraphic correlations between the northern and southern portions of the basin.

Late Chesterian–Early Morrowan Series

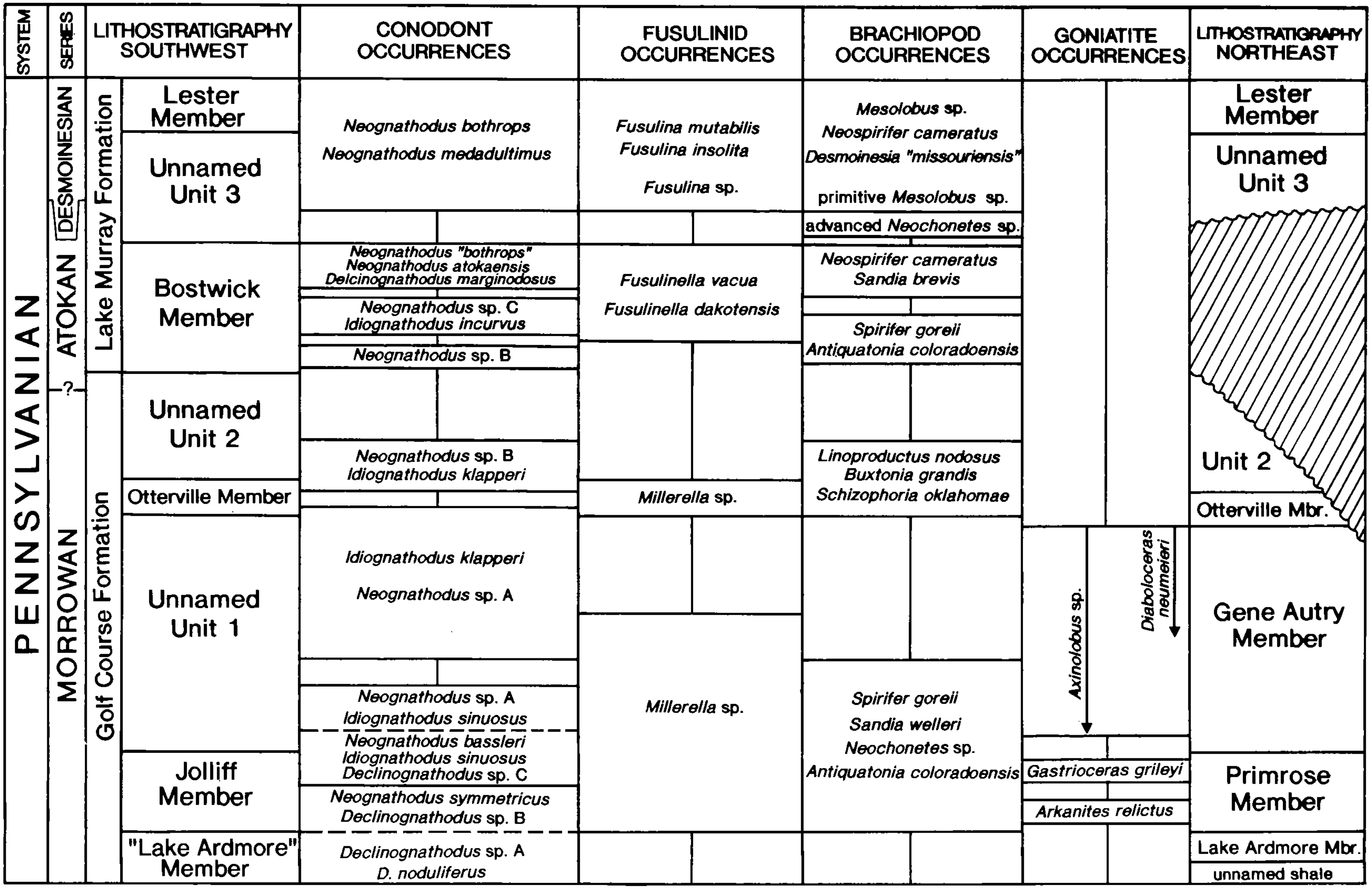
Northern Outcrop Areas

Subdivisions of the Springer Formation were named by Tomlinson (1929). Numerous variations developed historically in the way the Springer and its subdivisions have been used; these are summarized by Tomlinson and McBee (1959), Meek (1983), and Sutherland (1988). The Springer Formation, as currently used (Text-fig. 2), is restricted to the Rod Club, Overbrook, and Lake Ardmore Members and adjacent shales. Meek (1983) recorded a maximum thickness of 2,400 ft for the Springer Formation in the northern part of the basin. The Springer plus the overlying Primrose Member of the Golf Course Formation in that area form an apparently continuous sequence of laterally lensing and discontinuous units. The sandstones of the Springer Formation are fine grained, well sorted, and coarsen upward. They represent mostly delta-front deposits prograding from northwest to southeast (Meek and others, 1988).

The biostratigraphic nature of this sequence cannot be evaluated closely due to the sparseness of diagnostic fossils and the lack of outcrops for many of the shales. Sparse conodont recoveries in the northern outcrop belt indicate a late Chesterian age for the Rod Club and Overbrook Members, plus the lower two-thirds of the unnamed shale between the Overbrook and Lake Ardmore Members. Straka (1972) placed the base of the Morrowan at the base of the Lake Ardmore Member (Text-fig. 2), but more recent information places the base of the Morrowan somewhat lower, in the upper part of the underlying unnamed shale (Kleehammer, 1991).

Southern Outcrop Area

Correlation of the Springer Formation and of the overlying Primrose Member of the Golf Course Formation into the southern outcrop area is a major problem. In the latter area, the sequence is comparatively thin and only two or three mostly thin sandstone layers occur below the Jolliff. None are closely similar lithologically to specific Springer members north of Ardmore. Furthermore, Tomlin-



Text-figure 3. Comparison of distributions of conodont, fusulinid, brachiopod, and goniatite occurrences in strata of Morrowan and Atokan age in the Ardmore basin. Conodont data by R. C. Grayson, Jr., fusulinid data from Waddell (1966), brachiopod data by P. K. Sutherland, and goniatite data by W. L. Manger.

son and McBee (1959) postulated that an unconformity occurs at the base of the Jolliff conglomerate. Because of these uncertainties, the thin sandstone immediately below the Jolliff in the southwest is tentatively referred to the "Lake Ardmore" Member.

Morrowan Series

The Morrowan Series in the Ardmore basin extends from the upper part of the unnamed shale below the Lake Ardmore Member of the Springer Formation through the Golf Course Formation (Text-fig. 2). An evaluation of this series is difficult because of marked lateral changes in facies both within and between the northern and southern outcrop areas. Study has been uneven because shales are poorly exposed and interest has typically focused on fossiliferous intervals, particularly limestones.

Primrose and Jolliff Members

The Primrose Member of the Golf Course Formation, as much as 300 ft thick, is recognized only in the northern part of the Ardmore basin where shales are interbedded with thin to thick layers of fine-grained, well-sorted, glauconitic sandstone. The sequence was deposited in a shallow marine shelf setting (Meek and others, 1988). This unit shows an increase in sand to the north, which suggests a northern source (Maley, 1986).

Pennsylvanian conodonts indicate that the Primrose in the north and the Jolliff Member of the Golf Course Formation found only in the southern Ardmore basin, were deposited more or less simultaneously. Jolliff lithologies include limestone- and chert-pebble conglomerate and pebbly limestones interbedded with shale. The conglomerates are best developed in two areas: in the southern Lake Murray area (T. 6 S., R. 2 E.; see Text-fig. 1) and adjacent to U.S. 70, west of Ardmore (T. 4–5 S., R. 1 E.). Maximum thicknesses in both areas are ~80 ft. The conglomerates are distinctly lensing and appear to "shale out" in the central part of the southern outcrop area. Whether this is the result of poor outcrop quality there or of local change in facies cannot be concluded with certainty.

All authors have agreed that the source of the limestone cobbles, derived mostly from the Ordovician, and the chert pebbles, derived mostly from the Woodford (Devonian), are from the uplift and erosion of the nearby Criner Hills axis to the southwest. Some disagreement exists concerning the depositional system that produced the conglomerates. Maley (1986) interpreted the Jolliff as representing submarine-fan deposits. Alternatively, a modified fan-delta model may be more applicable considering the diverse nature of the biota and the obvious shallow shelf nature of the associated limestones.

Gene Autry Member and Unnamed Unit 1

In the northern part of the Ardmore basin, the Gene Autry Shale occurs in the interval between the Primrose and Otterville Members and ranges in thickness from 1,200 to 1,500 ft. The Gene Autry is locally much better exposed than is its general equivalent Unnamed Unit 1 in the south. The Gene Autry is generally characterized by the common occurrence of siderite concretions, which locally occur in great numbers on the exposed shale slopes. The sequence is believed to represent somewhat deeper water with low-oxygen conditions (Maley, 1986). The only biostratigraphically significant megafossils occurring throughout the Gene Autry are ammonoids and nautiloids. Microfauna of the lower 900 ft of the unit is confined to foraminifers, but above that level rare occurrences of ostracodes and conodonts possibly represent slight shallowing. In the top 100–200 ft, continued shallowing is indicated by the scattered occurrence of a few thin limestone units and a calcareous benthic macrofauna that includes solitary rugose corals, bivalves, gastropods, and brachiopods, in addition to the goniatites and nautiloids. The contact with the overlying Otterville is sharp, but apparently conformable (Maley, 1986).

In the southern outcrop area, Unnamed Unit 1, consisting mostly of shale, occurs between the Jolliff and Otterville Members (where exposed). It attains thicknesses of ~900 ft. About 4 mi southeast of the Carter/Love county line, on the Forbes Ranch (MS 144; T. 6 S., R. 2 E.; see Appendix for precise location of all sections discussed herein) discontinuous beds of limestone-cobble conglomerate and pebbly limestone, locally as much as 30 ft thick, occur within the interval. Limestones in the lower part of Unnamed Unit 1 occur within 60 ft of the Jolliff in the area west of Ardmore. They consist of thin-bedded fossiliferous, oolitic grainstones ~40 ft thick (Cromwell, 1974). Above those limestones is a thick, covered interval presumed to be shale. The Otterville is not exposed in this immediate area, so Unnamed Units 1 and 2 cannot be separated there.

Otterville Member

The Otterville Limestone is the only Morrowan unit that has been recognized in both the northern and southern parts of the Ardmore basin and is thus an important marker horizon in the sequence (Text-figs. 2,3). In the northern Ardmore basin, the Otterville is primarily an oolitic or coated skeletal grainstone developed on the flanks of the Caddo anticline. There, the Otterville typically forms two prominent ridges separated by cover presumed to be shale. In the northeastern part of the basin, the Otterville consists of one to three thin beds of finer grained bioclastic packstone separated by shales. Thicknesses of the unit across the area vary from 30 to 50 ft. The lithologic and thickness trends within the Otterville Member suggest that the Caddo anticline was actively being uplifted during Otterville deposition with oolitic sequences being deposited in shallow high-energy environments on the flanks of the anticline. To the northeast, the thinner deposits of mud-dominated limestone with thicker marine shales represent more basinal, lower energy deposition (Maley, 1986).

The Otterville in the south-central part of the southwestern outcrop belt ranges in thickness from 20 to 40 ft. It consists in that area of thin- to medium-bedded, sandy, bioclastic limestone with some interbedded shale (Maley, 1986). The lithologic integrity of Unnamed Unit 1, the Otterville Limestone Member, and Unnamed Unit 2 are mostly lost southeast of the Carter/Love county line (south margin of T. 5 S.). Transitional units, recognized with uncertainty, may be present at Forbes Ranch (MS 144), but at Paschall Ranch, 2.2 mi farther to the southeast (MS 143), all have changed facies to a sequence composed predominantly of interbedded calcareous sandstones, pebbly limestones, and limestone-cobble conglomerates. In that area the sequence cannot be differentiated into named lithologic units or separated lithologically from the overlying Bostwick.

Unnamed Unit 2

In the northern Ardmore basin, Unnamed Unit 2 consists mostly of gray shale and a few thin limestone layers. The thickness is highly variable due to an unconformity cutting irregularly into this interval (see description following that of the Bostwick Member) and possibly in part due to thinning onto rising structural elements. Thicknesses of the unit in this area range from a feather's edge to 600 ft. The thickest preserved occurrences are found in the Berwyn syncline. Thin limestones nearest the Caddo anticline are typically fossiliferous, oolitic grainstones, and those farther to the northeast are skeletal packstones. The unit appears to be continuously deposited with the underlying Otterville. The basal shale of the unit is similar in character to that below the highest Otterville limestone.

Unnamed Unit 2 in the southwestern outcrop belt consists of gray shale and covered intervals considered to be mostly shale. West of Ardmore, no outcrops are found, and the interval forms part of a continuous covered interval from the lower part of Unnamed Unit 1 to the base of the Bostwick. The thickness of Unnamed Unit 2 decreases southeastward from about 370 ft at MS 148 (T. 5 S., R. 1 E.) to 130 ft at MS 146 (T. 5 S., R. 2 E.). Based on this southeastward thinning, Cromwell (1974, p. 23) postulated post-depositional truncation and lithologic discontinuity with the superjacent lenticular conglomerates of the Bostwick. Farther south, however, the stratigraphic relations between Unnamed Unit 2 and the Bostwick become increasingly uncertain because of the introduction into Unnamed Unit 2 of limestone-cobble conglomerates and pebbly limestone (see discussion of Otterville Member).

Atokan Series

Bostwick Member

The Bostwick Member of the Lake Murray Formation (Text-fig. 2) is primarily preserved in the southwestern outcrop belt, but it also occurs immediately north of Ardmore on the nose of the Caddo anticline (Text-fig. 1). Its absence in the northern and northeastern part of the Ardmore basin may result from (1) those areas lying farther from the Criner Hills source, (2) no deposition occurring on positive structural elements, or (3) the unit having been removed by post-depositional erosion.

The Bostwick Member in the southwestern outcrop belt is the most prominent ridge-forming unit in the Ardmore basin. The lithologies within the Bostwick range from coarse limestone- and chert-cobble conglomerates to pebbly limestones, sparse carbonate mudstones, calcareous sandstones, and shales. The conglomerates, derived from the Criner Hills uplift to the southwest, are the most striking lithologies, but sandy, tan to gray shales make up a major portion of the member (Clopine, 1986). The thickness is 400 ft at the Carter/Love county line (north margin T. 6 S., R. 2 E.). This complex lithologic succession has been interpreted as representing switching and reworking of either deltaic (Clopine, 1986) or fan-delta depositional systems (Landis, 1982).

Late Atokan–Early Desmoinesian Series

Unnamed Unit 3

The Morrowan–Atokan boundary in the northern part of the Ardmore basin is marked by an angular unconformity (Clopine, 1986, fig. 11). Only the highest part of the Atokan Series is preserved in that area (Text-fig. 2). This unconformity was first observed on a seismic profile by Shell geologists who brought it to Sutherland's attention in 1984. Subsequent field studies by University of Oklahoma graduate students, including Clopine (1986) and Maley (1986), made it possible to differentiate Unnamed Units 2 and 3 in the northern part of the Ardmore basin (Text-fig. 2) with the aid of conodont interpretations by Grayson.

Cross-sections constructed by Clopine (1986, figs. 11,12) suggest that the Berwyn syncline (Text-fig. 1) was slightly downwarped in Atokan time as the northern margin of the Ardmore basin was uplifted and eroded. Folding was followed by transgression of Unnamed Unit 3, which rests directly on truncated parts of Unnamed Unit 2 across most of the Berwyn syncline. In the most easterly exposures in the northeastern outcrop belt (Text-fig. 1), Unnamed Unit 3 rests locally on the Otterville and laterally to the east directly on the Gene Autry Shale (Clopine, 1986, fig. 11).

Unnamed Unit 3 in the north extends from the basal unconformable contact with Unnamed Unit 2 upward to the base of the Lester Limestone Member (Text-fig. 2). It is composed predominantly of marine shale with a few interbedded, thin, sandy limestones of variable character, and it includes also minor nonmarine shales and a thin coal. It is 180 ft thick in MS 363 (T. 3 S., R. 3 E.), where the base of the unit is marked by a thin conglomerate that contains limestone, clay-ironstone, and calcareous siltstone pebbles (Tennant, 1981; Tennant and others, 1982, fig. 51). Clopine (1986, p. 41) argued that these pebbles were most likely derived from the erosion of underlying Unnamed Unit 2, because of the petrographic similarity of the limestone pebbles to thin limestones in that sequence. There is a lack of any chert in the basal conglomerate.

In the southwestern part of the Ardmore basin, Unnamed Unit 3 is the interval that lies between the Bostwick and Lester Members of the Lake Murray Formation (Text-fig. 2). The contact between the Bostwick and Unnamed Unit 3 cannot be evaluated since the top of the Bostwick cannot be consistently determined laterally because of poor exposures. Unnamed Unit 3 is predominantly gray, sandy shale, but it locally includes significant sandstones as well as rare, thin, sandy limestones. In the extreme southeast, pebbly limestones and limestone-pebble conglomerates also occur. The thickness of the unit in the southwestern outcrop belt has not been satisfactorily determined due to poor exposures. It is ~450 ft thick in the Forbes Ranch section (MS 321; T. 6 S., R. 2 E.).

Lester Member

The Lester consists of oolitic and skeletal limestones interbedded with thin, mature quartz sandstones or sandy limestones, and thicker intervals of marine shale. Individual limestone and sandstone units are lenticular laterally, but the interval as a whole can be correlated throughout the Ardmore basin (Clopine, 1986). Waddell (1966) assigned a Desmoinesian age to the Lester based on fusulinids.

BIOSTRATIGRAPHY

Foraminiferal Occurrences

Morrowan Series

Maley (1986), in a preliminary search, reported the fairly common occurrence of small foraminifers in many of the Morrowan limestones and calcareous shales in the Ardmore basin. *Millerella* sp. was recorded from the Jolliff (Pl. 1, Fig. 7), lower half of Unnamed Unit 1 (Pl. 1, Fig. 8), the Otterville, and the lower beds of Unnamed Unit 2.

The Morrowan–Atokan boundary has generally been placed at the first occurrence of either *Eoschubertella* or *Profusulinella*. Neither of those Early Atokan genera has been found in the Ardmore basin. The search for such forms is inhibited by the poor exposure of the shales in the middle and upper parts of Unnamed Unit 2. That interval has thus far produced no significant fossils of any type.

Waddell (1966) recorded the occurrence of the late Atokan genus *Fusulinella* from throughout most of the Bostwick Member. He stated (p. 21) that the oldest specimens of *Fusulinella dakotensis* occur 50 ft above the lowest conglomerate in the Bostwick Member along the Carter/Love county line (his MS 5, unit 2; T. 5 S., R. 2 E.). He further stated that *F. dakotensis* (Pl. 1, Figs. 9,10) is more advanced than *Fusulinella prolifica* described by Thompson (1935) from 200 ft above the base of the Atoka Formation in Coal County, Oklahoma. He therefore considered the Bostwick to be of younger Atokan age than rocks of the Atoka Formation that contain *F. prolifica*. In addition, John R. Groves (personal communication, 1991) states that both *F. dakotensis* and *F. prolifica* are more advanced evolutionarily than primitive species such as *F. primaeva*, from the Marble Falls Limestone of Texas. *Fusulinella vacua* (Pl. 1, Figs. 11,12) occurs in the upper part of the Bostwick Member.

The base of the Desmoinesian Series is defined by the first occurrence of the fusulinid genus *Fusulina*. Waddell (1966, p. 70) recorded *Fusulina insolita* (Pl. 1, Figs. 1,2) from 40 ft below the top of Unnamed Unit 3 in the southwestern outcrop belt (in his MS 1, unit 7; T. 6 S., R. 2 E.). In the same area but much lower in the section, we found rare, poorly preserved specimens of a probable primitive *Fusulina*? (Pl. 1, Figs. 3,4; identified by John Groves and William Clopine, 1990). They occur in the lower part of Unnamed Unit 3 in MS 321 (T. 6 S., R. 2 E.), an estimated 150 ft above the top of the Bostwick, but are too poorly preserved to form the basis for an interpretation. In the northeastern outcrop belt, we have found a primitive *Fusulina* (Pl. 1, Figs. 5,6; identified by John Groves and William Clopine, 1990) 20–30 ft above the base of unit 11 at MS 363 (T. 3 S., R. 3 E.), ~60 ft below the base of the Lester.

In summary, the base of the Desmoinesian in the Ardmore basin can be placed at least as low as the upper part of Unnamed Unit 3. The Lester Member is definitely early Desmoinesian in age based on Waddell's (1966) description of the genus *Fusulina* from this unit.

Conodont Occurrences

In the Early and Middle Pennsylvanian of the Ardmore basin the conodont genera *Neognathodus*, *Idiognathodus*, and *Declinognathodus* exhibit rapid changes of presumed evolutionary significance. *Declinognathodus* is tentatively subdivided into five Early and Middle Pennsylvanian species-level groups (Grayson and others, 1990), four of which are illustrated in Plate 2. Each declinognathodontid group consists of a broad series of intergradational morphotypes that are distinct because of marked shifts in the range of variation and the appearance of new morphotypes. The lowest two of these, however, need to be accepted with caution since the number of elements recovered at most horizons is small.

Following its mid-Morrowan introduction, *Idiognathodus* dominates Midcontinent Pennsylvanian conodont faunas. Using the general criteria outlined by Grayson and others (1989), three species-level groups marked by a transitional overlap are evident (Pl. 2).

Neognathodus (when present) provides the most sensitive biostratigraphy. Occurrences of that genus represent at least 10 Early and Early–Middle Pennsylvanian species-level groups, nine of which are illustrated in Plate 2.

Ranges and concurrent ranges of the various conodont taxa provide a preliminary basis for biostratigraphic subdivision of the early Morrowan to early Desmoinesian Series in the Ardmore basin (Text-fig. 3). Some of these potential zones are considered to be more significant than others, and all need to be tested in other areas.

The Mississippian–Pennsylvanian boundary can be located in the northern part of the Ardmore basin based on sparse conodont recoveries from some of the shales of the Springer Formation (Text-fig. 2). Straka (1972) recovered the Chesterian conodont *"Adetognathus" unicornis* from the shale above the Rod Club, and Kleehammer (1991) recovered *Gnathodus bilineatus* and *Lochreia commutata* from the lower and middle part of the shale (775 ft thick) between the Rod Club and Lake Ardmore Members in the northeastern Ardmore basin, near Meek's (1983; MS 401, T. 3 S., R. 4 E.), in an area where the Overbrook is not developed. Straka (1972) placed the base of the Morrowan at the base of the Lake Ardmore Member (Text-fig. 2) because of the occurrence of the Early Pennsylvanian conodont *Rhachistognathus primus* in the Target Limestone lentil. More recently, Kleehammer (1991) recovered rare specimens of the Morrowan conodonts *Declinognathodus noduliferus, Gnathodus higginsi,* and *Cavusgnathus lautus* from an interval 137–228 ft below the base of the Lake Ardmore, in the same sequence (near MS 401). The base of the Morrowan falls in the upper part of that shale below the Lake Ardmore Member, but at least 200 ft of barren interval lies above the highest Chesterian forms.

Declinognathodus noduliferus has not been found (in situ) in the southwestern outcrop belt. In the County Line section (MS 409), the Chesterian–Morrowan boundary falls somewhere within a 100-ft interval separating a thin limestone-pebble conglomerate layer with the Chesterian conodont *"Adetognathus" unicornis*, and the lower Jolliff Member with the Pennsylvanian conodonts *Declinognathodus* sp. B and *Neognathodus symmetricus*. Farther to the northwest, on U.S. 70 (north margin sec. 33, T. 4 N., R. 1 E.), the early Morrowan conodont *Declinognathodus* sp. A has been recovered from the "Lake Ardmore" Member 10–15 ft below the base of the Jolliff Member.

The correlation of the Primrose sandstone and shale sequence in the north with the Jolliff limestone and shale interval in the south is based on the occurrence in each of a transition from *Neognathodus symmetricus* to *N. bassleri* joined more or less simultaneously by the appearance of *Idiognathodus sinuosus*.

Conodonts do not provide an unequivocal basis for the recognition of the Morrowan–Atokan boundary in the Ardmore basin. The boundary has historically been based on fusulinid biostratigraphy, and the marker fusulinids *Eoschubertella* and *Profusulinella* have not as yet been found in the Ardmore basin. In addition, fusulinid and conodont biostratigraphies have not been fully integrated. Groves and Grayson (1984) described a mutual occurrence of *Eoschubertella* with conodonts identified herein as *Neognathodus* sp. B and *Idiognathodus klapperi* from the frontal Ouachita Mountains. Those conodonts occur in the southwestern Ardmore basin in rocks that have been variously identified as Otterville and/or lower Bostwick (Text-fig. 3).

Conodont data are not as ambiguous in the vicinity of the Atokan–Desmoinesian boundary. Grayson and Lambert (1987), in a study of species-level changes in the conodont genus *Neognathodus*, placed the Atokan–Desmoinesian boundary at the first appearance of the fusulinid genus *Fusulina* (Text-fig. 3). Samples proximal to the boundary that contained *Neognathodus "bothrops"* (an undescribed new species) were regarded as late Atokan, while those that contained *N. bothrops* and *N. medadultimus* were considered to indicate an early Desmoinesian age (Text-fig. 3).

On the basis of conodonts, the Atokan–Desmoinesian boundary in the northern outcrop belt can be drawn tentatively in MS 363 (T. 3 S., R. 3 E.) at the top of unit 10 (burrowed-sandstone marker) from which *Neognathodus "bothrops"* has been recovered. As noted earlier, an early Desmoinesian *Fusulina* has been recovered from 20 to 30 ft higher stratigraphically, in the lower part of unit 11.

The Lester Limestone in the Ardmore basin is characterized by the occurrence of the Desmoinesian conodonts *Neognathodus bothrops* and *N. medadultimus* (Text-fig. 3).

Brachiopod Occurrences

The biostratigraphic distribution of Morrowan brachiopods is well known in the southern Midcontinent, because of the studies in northern New Mexico by Sutherland and Harlow (1973) and those in the southwestern Ozarks by Henry and Sutherland (1977). A variety of Morrowan brachiopods described by those authors has been collected from the Ardmore basin, in the Jolliff, lower part of Unnamed Unit 1, Otterville, and lower part of Unnamed Unit 2. Included are *Spirifer goreii*, *Antiquatonia coloradoensis*, *Sandia welleri*, *Schizophoria oklahomae* (Pl. 3, Fig. 16), *Eolissochonetes* sp. (Pl. 3, Figs. 8,9), *Linoproductus nodosus* (Pl. 3, Fig. 13), and *Buxtonia grandis*. Those recoveries are, however, from widely scattered localities and do not form a basis for the recognition of Henry and Sutherland's (1977) Morrowan brachiopod zones.

The Morrowan–Atokan boundary in the Ardmore basin cannot be drawn precisely based on brachiopods. The Otterville and lower part of Unnamed Unit 2 contain an array of characteristic Morrowan brachiopods, but diagnostic Atokan species have not been found in the lower part of the Bostwick Member. The most distinctive brachiopods found in the lower Bostwick are *Spirifer goreii* (Pl. 3, Figs. 10–12) and *Antiquatonia coloradoensis* (Pl. 3, Figs. 10–12). These species occur also in the Morrowan, but have been described from rocks of both Morrowan and Atokan ages in northern New Mexico (Sutherland and Harlow, 1973).

From the upper Bostwick (late Atokan) have come *Neospirifer cameratus* (Pl. 3, Fig. 6), *Kozlowskia montgomeryi* (Pl. 3, Fig. 7), and *Sandia brevis* (Pl. 3, Fig. 15).

Unnamed Unit 3 is poorly exposed in the southwestern outcrop belt, where it has produced few brachiopods. In the northeastern outcrop belt, at MS 363-9 (T. 3 S., R. 3 E.), calcareous shales in the middle part of the unit have produced sequences of large numbers of chonetid brachiopods. Preliminary studies indicate a possible transition from *Neochonetes* to *Mesolobus* (Pl. 3, Figs. 1–5). A closer study of those faunas may aid in a more precise definition of the Atokan–Desmoinesian boundary.

The Lester is characterized by the occurrence of typical Desmoinesian brachiopods including *Mesolobus* sp., *Neospirifer cameratus*, and *Desmoinesia "missouriensis."*

The Jolliff has produced a brachiopod of exceptional general interest. The genus *Teguliferina* (Pl. 3, Fig. 18) had not been found previously below the Middle Pennsylvanian. The abundant specimens in the Jolliff are silicified and occur in small "patch reefs" formed by dense clusters of specimens attached by spines (Pl. 3, Fig. 18; Sutherland, 1987; Sutherland, 1989).

Goniatite Occurrences

A goniatite assemblage including *Arkanites relictus*, *Bilinguites eliasi*, and *Cancelloceras huntsvillense* occurs in the lower part of the Primrose Member, Golf Course Formation, 1 mi west of Caddo Village (T. 3 S., R. 1 E.). That assemblage can be correlated with the upper *Arkanites relictus* Zone, upper portion of the Prairie Grove Member, Hale Formation, in the type Morrowan succession of northwestern Arkansas (Manger and others, 1974; Manger and Saunders, 1980).

Elias (1956) reported the occurrence of *Gastrioceras grileyi* and *Eoasianites oblatus* in the upper part of the Primrose Member in the northern part of the Ardmore basin (T. 3 S., R. 2 E.). Those goniatites can be correlated with the Brentwood Member, Bloyd Formation, in the type Morrowan of northwestern Arkansas and with the Union Valley Formation on the northeast flank of the Arbuckle Mountains.

Goniatites occur commonly in some local areas in the Morrowan Jolliff Member in the southwestern outcrop belt, but the rock in which they occur is so well indurated and so highly recrystallized that it invariably breaks across the specimens. No identifiable specimens have been recovered.

Great numbers of identifiable goniatites have been recovered from the Gene Autry Member in the northeastern outcrop belt, particularly *Axinolobus*. An unusual association of the upper Morrowan zonal name-bearer *Axinolobus modulus* with *Diaboloceras neumeieri* occurs in the upper half of the Gene Autry (Text-fig. 3). In the well-known type Morrowan sequence in northwestern Arkansas, *Axinolobus* is known only from the Dye Shale and Kessler Limestone members of the Bloyd Formation (McCaleb, 1968). The lowest occurrence of *Diaboloceras neumeieri* in the Ozarks area is in the Trace Creek Shale Member of the Atoka Formation above the Morrowan–Atokan unconformity at the top of the Kessler (McCaleb, 1968). The two genera have not been found to occur together in northwest Arkansas or elsewhere.

The interpretation by Manger and others (1987, and this volume) is that the upper half of the Gene Autry, the overlying Otterville Member, and the preserved remnant of Unnamed Unit 2 (Text-fig. 3) represent strata that in part fills the gap represented by the Morrowan–Atokan unconformity at the top of the Kessler Limestone in the type Morrowan section. The implication of this interpretation is that the lowest occurrence of *Diaboloceras* in the southern Midcontinent is in strata of late Morrowan and not early Atokan age.

Palynomorph Occurrence

In the northeastern outcrop belt of the Ardmore basin, a thin coal (MS 363-8; T. 3 S., R. 3 E.) occurs 30 ft above the thin conglomerate at the base of Unnamed Unit 3 (Tennant and others, 1982). Samples from this coal have been processed and studied by Russel A. Peppers, who suggests a probable earliest Desmoinesian age for this unit (personal communication, 1989). He states that palynologically the coal is younger than the Rock Island Coal of Illinois, which underlies the Seville Limestone. Most workers place the top of the Atokan in that area at the top of that limestone. The next higher limestone in

Illinois is Desmoinesian and several coals lie between the two limestones. Therefore, the Atokan–Desmoinesian boundary is possibly a little higher than the Seville Limestone. Pepper states that if the Oklahoma coal "is Atokan, it would have to be uppermost Atokan," when the flora is compared with the Illinois basin sequence.

Coral Occurrences

Moderate numbers of solitary rugose corals and tabulate corals have been collected from Morrowan and Atokan strata in the Ardmore basin, but none of these have been described. One colonial rugose coral of considerable regional interest, the genus *Petalaxis*, has been recovered from several localities in the Morrowan Jolliff Member. Colonial rugose corals are extremely rare in Pennsylvanian strata in North America, but *Petalaxis*, described originally from the Middle Carboniferous in the Moscow basin, has been found in four southern Midcontinent Morrowan localities, all associated with the *Idiognathodus sinuosus* conodont zone. The other occurrences are in (1) northeastern Oklahoma, southwestern Ozarks, in the Brewer Bend Member of the Sausbee Formation; (2) in central Texas, Llano uplift, in the Lower Marble Falls Limestone; and (3) in southern New Mexico, Bishop Cap, Organ Mountains, in the La Tuna Formation. Possibly the four occurrences of *Petalaxis* represent a single migration into the southern Midcontinent region (Sutherland, 1985).

CONCLUSIONS

1. The Mississippian (Chesterian)–Pennsylvanian (Morrowan) boundary, based on the scattered and rare occurrences of conodonts, falls in the northern Ardmore basin in the upper part of the unnamed shale above the Overbrook and below the Lake Ardmore Sandstone Members of the Springer Formation. The base of the Morrowan in that area is marked by the lowest occurrence of *Declinognathodus noduliferus*.
2. The sandstone members of the Springer Formation in the northern part of the Ardmore basin cannot be correlated lithostratigraphically into the southwestern outcrop belt. In that area, in the County Line Road section, the Chesterian–Morrowan boundary must occur in a mostly covered interval (100 ft thick) that lies between a thin limestone-pebble conglomerate layer (position unknown within Springer Formation) that contains Chesterian conodonts and the Jolliff Member from which early Morrowan conodonts have been recovered.
3. The Primrose Sandstone Member in the north was deposited more or less simultaneously with the Jolliff Limestone Member in the southwest based on the occurrence in each of the transition of the conodont *Neognathodus symmetricus* to *N. bassleri* and the approximate simultaneous appearance in each of *Idiognathodus sinuosus*.
4. The Otterville Member is the only Morrowan unit that can be recognized across the Ardmore basin. It changes facies laterally and carries a generalized Morrowan fauna.
5. The Morrowan–Atokan boundary cannot be precisely defined in the Ardmore basin. The marker early Atokan fusulinid genera *Eoschubertella* and *Profusulinella* have not been found. This fact, coupled with the occurrence of the late Atokan fusulinid *Fusulinella dakotensis*, a somewhat more advanced species of that genus beginning 50 ft above the base of the Bostwick, indicates the probability of an unconformity at the base of the Bostwick Member or in the mostly covered upper part of Unnamed Unit 2, from which no diagnostic fossils have been recovered. Neither the conodont nor the brachiopod faunas change significantly between the Otterville and Lower Bostwick.
6. The Bostwick Member does not occur in the northern part of the Ardmore basin except for a poorly exposed, thin exposure on the nose of the Caddo anticline immediately north of Ardmore. Farther north, this unit was removed by erosion or never deposited.
7. An unconformity in the north, apparently resulting from the uplift of the northeastern margin of the Ardmore basin, records the absence of most if not all of the Atokan Series in that area. Across most of the area the middle or upper part of Unnamed Unit 3 rests unconformably on a gently folded and truncated Unnamed Unit 2.
8. The location of the Atokan–Desmoinesian boundary in the northern outcrop belt can be drawn tentatively only in MS 363 (T. 3 S., R. 3 E.) approximately at the top of unit 10 (burrowed-sandstone marker) based primarily on conodonts. Atokan conodonts occur in unit 10 and what are possibly Atokan brachiopods occur in unit 9. Early Desmoinesian fusulinids occur in the lower part of unit 11. The top of unit 10 is located 100 ft above the basal conglomerate of Unnamed Unit 3 and 80 ft below the base of the Lester Limestone. A conflicting interpretation of age relations in MS 363 is suggested by palynomorphs. A flora recovered from a thin coal (363-8), located 30 ft above the basal conglomerate, suggests a most likely earliest Desmoinesian age for this unit or, at the earliest, a latest Atokan age. This interpretation could imply a Desmoinesian age for the whole of Unnamed Unit 3 in the north. Further studies of all of the faunal elements are needed in order to resolve this inconsistency.
9. The Desmoinesian age of the Lester Limestone is well established. It is based on the occurrence of characteristic Desmoinesian species of fusulinids, conodonts, and brachiopods.

REFERENCES CITED

Clopine, W. W., 1986, The lithostratigraphy, biostratigraphy, and depositional history of the Atokan Series (Middle Pennsylvanian) in the Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 161 p.

Cromwell, D. W., 1974, The stratigraphy and environment of deposition of the lower Dornick Hills Group (Lower Pennsylvanian), Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 138 p.

Elias, M. K., 1956, Upper Mississippian and Lower Pennsylvanian Formation of south-central Oklahoma, *in* Petroleum geology of southern Oklahoma, vol. 1: American Association of Petroleum Geologists, p. 56–134.

Grayson, R. C., Jr.; and Lambert, L. L., 1987, Evolutionary distinction between Atokan and Desmoinesian (Pennsylvanian) *Neognathodus* [abstract]: Geological Society of America Abstracts with Programs, v. 19, p. 152.

Grayson, R. C., Jr.; Merrill, G. K.; and Lambert, L. L., 1989, Phylogenetic basis for species recognition within the conodont genus *Idiognathodus*, *in* Boardman, D. R.; and others (eds.), Middle and Late Pennsylvanian chronostratigraphic boundaries in north-central Texas: Texas Tech University Studies in Geology 2, p. 75–94.

______ 1990, Carboniferous gnathodontid conodont apparatuses, evidence of a dual origin for Pennsylvanian taxa: Courier Forschungsinstitut Senckenberg, v. 118, p. 353–396.

Groves, J. R.; and Grayson, R. C., Jr., 1984, Calcareous foraminifers and conodonts from the Wapanucka Formation (Lower–Middle Pennsylvanian), frontal Ouachita Mountains, southeastern Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 81–89.

Ham, W. E.; and Wilson, J. L., 1967, Paleozoic epeirogeny and orogeny in the central United States: American Journal of Science, v. 265, p. 332–407.

Ham, W. E.; Denison, R. E.; and Merritt, C. A., 1964, Basement rocks and structural evolution of southern Oklahoma: Oklahoma Geological Survey Bulletin 95, 302 p.

Henry, T. W.; and Sutherland, P. K., 1977, Brachiopod biostratigraphy of Morrowan Series (Pennsylvanian) in northwestern Arkansas and northeastern Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, p. 107–116.

Kleehammer, R. S., 1991, Conodont biostratigraphy of Late Mississippian shale sequences, south-central Oklahoma: University of Oklahoma unpublished M.S. thesis, 100 p.

Landis, Y., 1982, Application of a fan delta model to the Bostwick Conglomerates (Atokan), Ardmore basin, south-central Oklahoma: Baylor University unpublished senior thesis, 110 p.

Lane, H. R.; and Straka, J. J., II, 1974, Late Mississippian and Early Pennsylvanian conodonts, Arkansas and Oklahoma: Geological Society of America Special Paper 152, 144 p.

Maley, M. P., 1986, Depositional history of the Upper Morrowan (Pennsylvanian) strata of the Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 206 p.

Manger, W. L.; and Saunders, W. B., 1980, Lower Pennsylvanian (Morrowan) ammonoids from the North American Midcontinent: Journal of Paleontology Memoir 10, 56 p.

Manger, W. L.; Saunders, W. B.; and Quinn, J. H., 1974, Lower Pennsylvanian (Morrowan) ammonoids from the Primrose Member, Golf Course Formation, south-central Oklahoma [abstract]: Geological Society of America Abstracts with Programs, v. 5, p. 115–116.

Manger, W. L.; Miller, M. S.; and Mapes, R. H., 1987, Highest Morrowan (Pennsylvanian) ammonoid assemblages, Gene Autry Shale, Ardmore basin, Oklahoma, United States [abstract]: 11th International Congress of Carboniferous Stratigraphy and Geology, Beijing, Abstracts of Papers, v. 2, p. 482.

McCaleb, J. A., 1968, Lower Pennsylvanian ammonoids from the Bloyd Formation of Arkansas and Oklahoma: Geological Society of America Special Paper 96, 123 p.

Meek, F. B., III, 1983, The lithostratigraphy and depositional environments of the Springer and lower Golf Course Formations (Mississippian–Pennsylvanian) in the Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 212 p.

Meek, F. B., III; Elmore, R. D.; and Sutherland, P. K., 1988, Progradational sequence in the Springer Formation, Ardmore basin, Oklahoma: American Association of Petroleum Geologists Bulletin, v. 67, p. 512.

Straka, J. J., II, 1972, Conodont evidence of age of Goddard and Springer Formations, Ardmore basin, Oklahoma: American Association of Petroleum Geologists Bulletin, v. 56, p. 1087–1099.

Sutherland, P. K., 1985, The colonial rugose coral genus *Petalaxis* in the Middle Carboniferous of North America and its stratigraphic significance [abstract]: Geological Society of America Abstracts with Programs, v. 17, p. 730.

______1987, A primitive teguliferiniid brachiopod from the Morrowan (Early Pennsylvanian) of Oklahoma, U.S.A. [abstract]: 11th International Congress of Carboniferous Stratigraphy and Geology, Beijing, Abstracts of Papers, v. 1, p. 87.

______1988, Ardmore and Marietta basins *in* Southern Midcontinent region, *chapter 12 of* Sloss, L. L. (ed.), Sedimentary cover—North American craton; U.S.: Geological Society of America, The Geology of North America, v. D-2, , p. 326-331.

______1989, Possible ancestor to the Late Carboniferous–Early Permian teguliferiniid brachiopods in the Middle Carboniferous of Oklahoma, U.S.A.: Compte Rendu, Congrès International de Stratigraphie et de Gèologie du Carbonifère, Beijing, v. 2, p. 355–360.

Sutherland, P. K.; and Harlow, F. H., 1973, Pennsylvanian brachiopods and biostratigraphy in southern Sangre de Cristo Mountains, New Mexico: New Mexico Bureau of Mines and Mineral Resources Memoir 27, 173 p.

Tennant, S. H., 1981, Lithostratigraphy and depositional environments of the upper Dornick Hills Group (Lower Pennsylvanian) in the northern part of the Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 291 p.

Tennant, S. H.; Sutherland, P. K.; and Grayson, R. C., Jr., 1982, Stop descriptions—northern Ardmore basin, *in* Sutherland, P. K. (ed.), Lower and Middle Pennsylvanian stratigraphy in south-central Oklahoma: Oklahoma Geological Survey Guidebook 20, p. 38–42.

Thompson, M. L., 1935, Fusulinids from the Lower Pennsylvanian Atoka and Boggy Formations of Oklahoma: Journal of Paleontology, v. 9, p. 291–306.

Tomlinson, C. W., 1929, The Pennsylvanian system in the Ardmore basin: Oklahoma Geological Survey Bulletin 46, 79 p.

Tomlinson, C. W.; and McBee, W., Jr., 1959, Pennsylvanian sediments and orogenies of Ardmore district, Oklahoma, *in* Petroleum geology of southern Oklahoma: American Association of Petroleum Geologists, v. 2, p. 3–52.

Waddell, D. E., 1966, Pennsylvanian fusulinids in the Ardmore basin, Love and Carter Counties, Oklahoma: Oklahoma Geological Survey Bulletin 113, 128 p.

APPENDIX: Measured Sections Cited

MS 143: Paschall Ranch; SE¼SW¼ sec. 24, T. 6 S., R. 2 E.; Love County.

MS 144: Forbes Ranch (West); NW¼ sec. 22, T. 6 S., R. 2 E.; Love County.

MS 146: County Line; north margin, NE¼ sec. 6, T. 6 S., R. 2 E.; Carter/Love county line.

MS 148: Bussell Ranch; SW¼ sec. 13, T. 5 S., R. 1 E.; Carter County.

MS 321: Forbes Ranch (East); NE¼ sec. 22, T. 6 S., R. 2 E.; Love County.

MS 363: Dutton Ranch; NE¼SE¼ sec. 16, T. 3 S., R. 3 E.; Carter County.

MS 401: Goddard Ranch; NE¼ sec. 19, T. 3 S., R. 4 E.; Johnston County.

MS 409: County Line; north side NE¼NW¼ sec. 6, T. 6 S., R. 2 E.; Carter/Love county line.

Plates

PLATE 1

(All figures ×20 except as noted)

Figures 1,2.—*Fusulina insolita* Thompson. **1,2,** axial and sagittal sections, Unnamed Unit 3, Waddell (1966) MS 1, unit 7, OU 9281, 9282.

Figures 3,4.—*Fusulina*? sp. **3,4,** exterior view and axial section, Unnamed Unit 3, 321-3 (37–39 ft below top), OU 10000.

Figures 5,6.—*Fusulina* sp. **5,6,** axial section and exterior view, Unnamed Unit 3, 363-11 (20–30 ft above base), OU 10126.

Figures 7,8.—*Millerella* sp. axial sections. **7,** Jolliff Member, 338-9, OU 10127, ×80. **8,** Unnamed Unit 1, 145-10, OU 10128, ×80.

Figures 9,10.—*Fusulinella dakotensis* Thompson. **9,10,** sagittal and axial sections, Waddell (1966) MS5, unit 4. **9,** OU 9278. **10,** OU 9275.

Figures 11,12.—*Fusulinella vacua* Waddell. **11,12,** sagittal and axial sections, upper Bostwick Member, Waddell (1966), MS 5, unit 17, OU 9274, 9273.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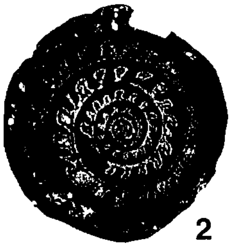
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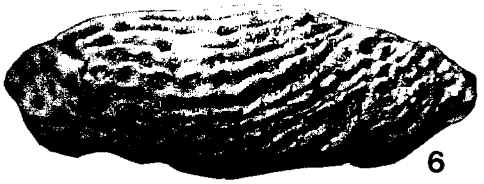
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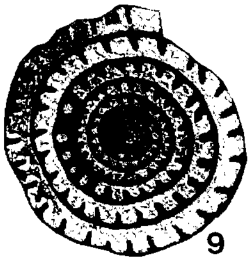
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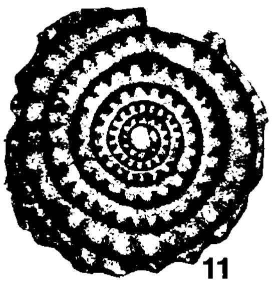
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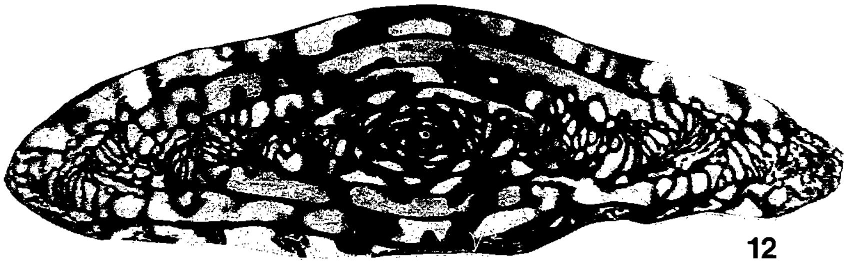
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PLATE 2

Figure 1.—*Hindeodus minutus* (Ellison), lateral view OU 10685, Bostwick Member, 145-33, ×42.

Figure 2.—*Cavusgnathus lautus* Gunnell, lateral view of dextral element OU 10686, Otterville Member, 366-21, ×42.

Figures 3,19,20,33.—*Idiognathoides sinuatus* (Harris and Hollingsworth). **3**, upper view of dextral element OU 10687, Unnamed Unit 3, 321-14, ×42. **19**, upper view of sinistral element OU 10688, 150-2, ×25. **20**, upper view of OU 10689, a troughed morphotype often assigned by authors to *I. ouachitensis* or *I. fossatus*, 150-2, ×25. **33**, upper view of sinistral element OU 10690, Jolliff Member, 153-25 (6.0 ft above base of unit), ×25.

Figures 4,12.—*Idiognathodus incurvus* Dunn. **4**, upper view OU 10691, Lester Member, 408-43, ×25. **12**, upper view OU 10692, Bostwick Member, 144-54, ×25.

Figures 5,6.—*Neognathodus medadultimus* Merrill. **5**, upper view OU 10693, Unnamed Unit 3, 321-14, ×42. **6**, upper view OU 10694, Lester Member, 408-43, ×42.

Figures 7,8.—*Neognathodus bothrops* Merrill. **7**, upper view OU 10695, Lester Member, 408-43, ×42. **8**, upper view OU 10696, Lester Member(?), 319-13, ×42.

Figure 9.—*Diplognathodus* sp. A, lateral view OU 10697, Bostwick Member, 144-54, ×42.

Figure 10.—*Diplognathodus* sp. B, lateral view OU 10698, Bostwick Member, 158-4, ×42.

Figure 11.—*Declinognathodus marginodosus* (Grayson), upper view OU 10699, Unnamed Unit 3, 321-14, ×25.

Figures 13,14.—*Neognathodus "bothrops."* **13**, upper view OU 10700, Bostwick Member, 144-54, ×25. **14**, upper view OU 10701, Bostwick Member, 144-54, ×25.

Figure 15.—*Neognathodus atokaensis* Grayson, upper view OU 10702, Bostwick Member, 144-54, ×25.

Figures 16,17.—*Neognathodus* sp. B. **16**, upper view OU 10703, Bostwick Member, 144-46B, ×25. **17**, upper view OU 10704, Bostwick Member, 150-1, ×25.

Figures 18,21,24,25.—*Idiognathodus klapperi* Lane. **18**, upper view OU 10705, "advanced" specimen showing streptognathodontid character, Bostwick Member, 148-6, ×25. **21**, upper view OU 10706, "advanced" specimen, Bostwick Member, 144-23, ×42. **24**, upper view OU 10707, Unnamed Unit 2, 144-31 (70 ft above base of unit), ×25. **25**, upper view OU 10708, specimen shows incomplete transverse ornamentation, Unnamed Unit 2, 144-31 (70 ft above base of unit), ×25.

Figure 22.—*Neognathodus* sp. C, upper view OU 10709, juvenile specimen, Bostwick Member, 150-8, ×42.

Figures 23,26.—*Neognathodus* sp. A. **23**, upper view OU 10710, Unnamed Unit 1, 144-31 (70 ft above base of unit). **26**, upper view OU 10711, (type Jolliff) Unnamed Unit 1, 147-6, ×25.

Figure 27.—*Declinognathodus* sp. C, upper view OU 10712, Jolliff Member, 153-16, ×42.

Figures 28,29.—*Idiognathodus sinuosus* Ellison and Graves. **28**, upper view OU 10713, Jolliff Member, 339-25 (6 ft above base of unit), ×25. **29**, upper view OU 10714, Jolliff Member, 339-25 (6 ft above base of unit), ×25.

Figure 30.—*Declinognathodus* sp. A, upper view OU 10715, Springer Group, 153-1A, ×42.

Figure 31.—*Rhachistognathus primus* Dunn, upper view OU 10716, Jolliff Member, 153-1A, ×42.

Figure 32.—*Declinognathodus* sp. B, upper view OU 10717, Jolliff Member, 153-11, ×42.

Figures 34,35.—*Neognathodus symmetricus* Lane. **34**, upper view OU 10718, Jolliff Member, 144-10 (top 1 ft of unit), ×25. **35**, upper view OU 10719, Jolliff Member, 144-10 (top 1 ft of unit), ×25.

Figures 36,37.—*Neognathodus bassleri* (Harris and Hollingsworth). **36**, upper view OU 10720, Jolliff Member, 339-23, ×25. **37**, upper view OU 10721, Jolliff Member, 339-23, ×25.

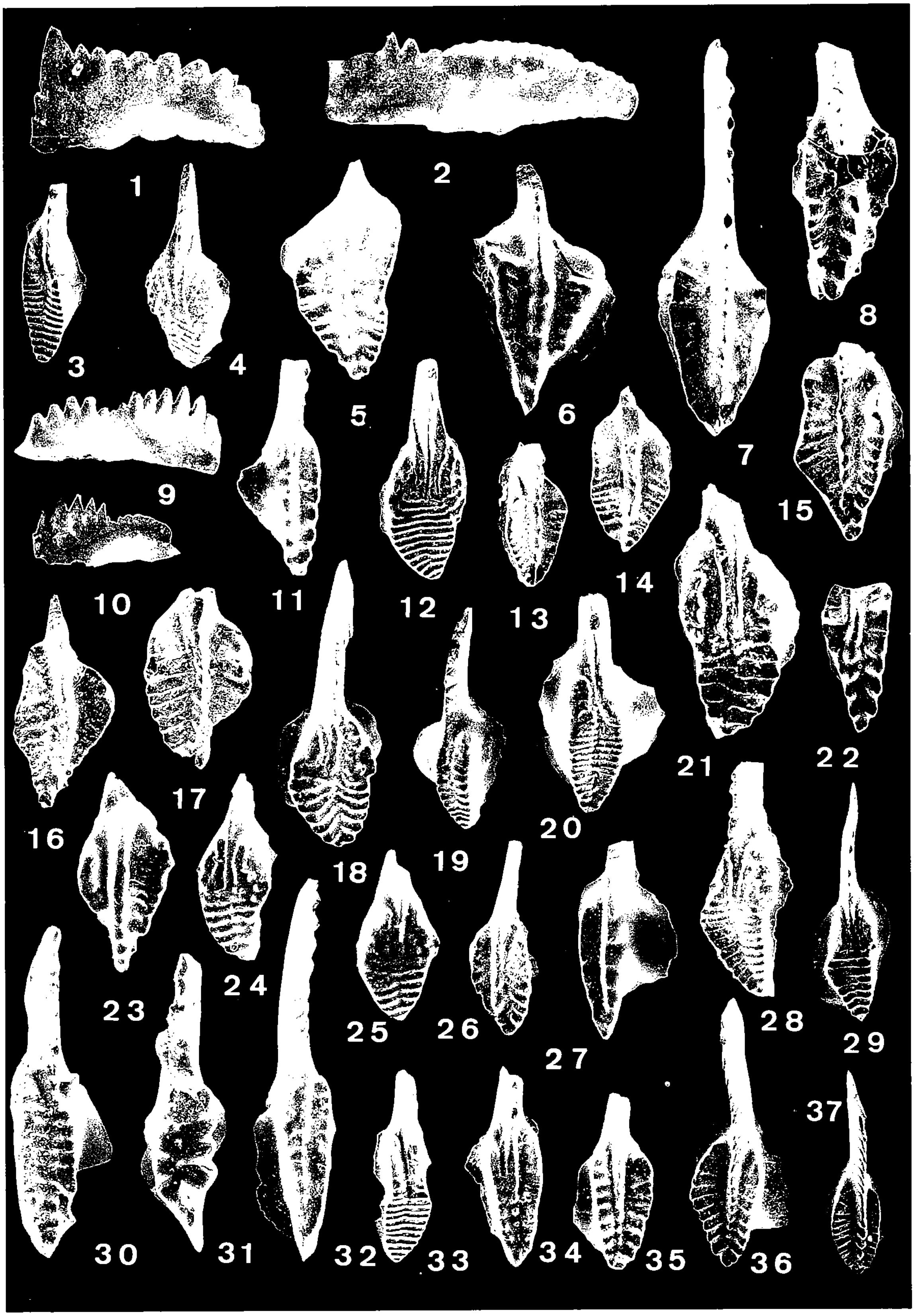
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PLATE 3

(All figures ×1 except as noted)

Figures 1–5.—*Neochonetes–Mesolobus* transition, all ventral views, Unnamed Unit 3, 363-11 (0–10 ft above base), OU 10129-10131, 10455, 10456, ×2.

Figure 6.—*Neospirifer cameratus* (Morton), ventral view, top of Bostwick Member, Cromwell (1974), fossil loc. F, OU 11126.

Figure 7.—*Kozlowskia montgomeryi* Sutherland and Harlow, ventral view, top of Bostwick Member, MS 146-33, OU 11127.

Figures 8,9.—*Eolissochonetes* sp., ventral views, Jolliff Member, 335-6a, OU 11128, 11129, ×2.

Figures 10–12.—*Antiquatonia coloradoensis* (Girty), ventral views, lower Bostwick Member, MS 148-7, OU 11130, 11131, 11120.

Figure 13.—*Linoproductus nodosus* (Newberry), ventral view, lateral equivalent of Otterville Member?, MS 143-1, OU 11121.

Figure 14.—*Spirifer goreii* Mather, ventral view, lower Bostwick Member, MS 148-7, OU 11122.

Figure 15.—*Sandia brevis* Sutherland and Harlow, ventral view, top of Bostwick Member, MS 146-33, OU 11123.

Figure 16.—*Schizophoria oklahomae* Dunbar and Condra, ventral view, Otterville Member, Cromwell (1974), fossil loc. E, OU 11124.

Figures 17,18.—*Petalaxis* n. sp., transverse and longitudinal sections, Jolliff Member, MS 149-4, OU 11125, ×3.5.

Figure 19.—*Tequilferina* n. sp., showing growth pattern, Jolliff Member, 3.2 mi west of Ardmore, Oklahoma, OU 10489, ×2.


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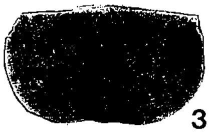
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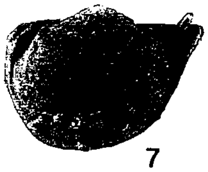
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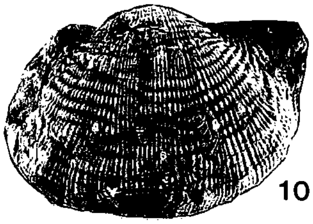
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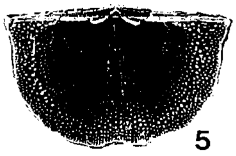
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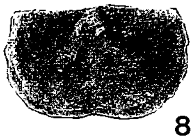
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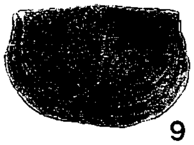
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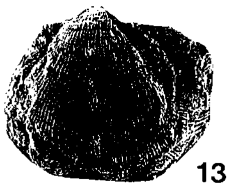
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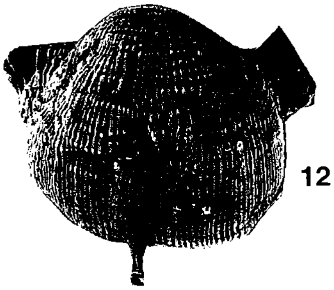
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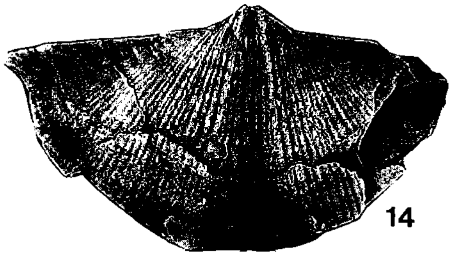
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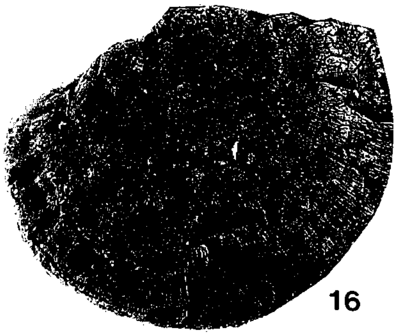
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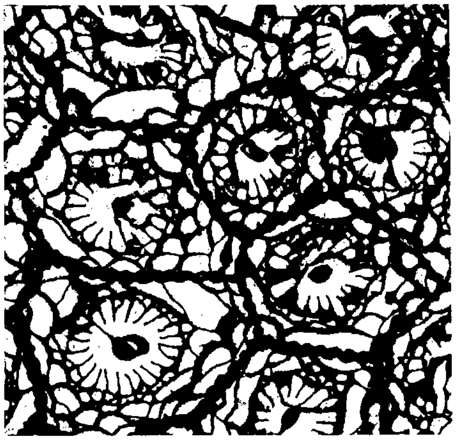

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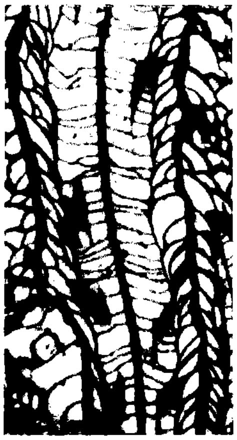

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Oklahoma Geological Survey Circular 94, 1992

Age and Correlation of the Gene Autry Shale, Ardmore Basin, Southern Oklahoma

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ABSTRACT.—The ammonoid assemblage from the Gene Autry Shale Member, Golf Course Formation, northern Ardmore basin, Oklahoma, is dominated by the genus *Axinolobus*. Traditionally, correlations based primarily on that taxon have equated the Gene Autry with the Dye and Kessler Members, Bloyd Formation, of the type Morrowan region, northwestern Arkansas. The appearance of *Diaboloceras* in the upper half of the Gene Autry, in association with *Axinolobus*, suggests that the interval of their concurrence is younger than the Kessler Member, and, while still Morrowan, it fills the gap represented by the Morrowan–Atokan unconformity in the type Morrowan region.

INTRODUCTION

The Ardmore basin, south-central Oklahoma, lies adjacent to the southern flank of the Arbuckle Mountains. Sedimentation during Morrowan time was influenced by the southern Oklahoma aulacogen resulting in a maximum accumulation in the northern part of the basin of >2,000 ft of section that exhibits pronounced facies changes between the northern and southern portions of the basin (Maley, 1986). An assemblage of nearly 2,000 complete and fragmentary ammonoids has been collected, primarily by Maley (1986), from the Gene Autry Member, Golf Course Formation, at 16 localities in Carter and Johnston Counties, southern Oklahoma (Text-fig. 1). That assemblage is dominated by the genus *Axinolobus*, the first 10-lobed goniatite to appear in the Carboniferous, and a well-known upper Morrowan taxon (McCaleb, 1968). Associated with the former genus in the upper part of the Gene Autry succession, *Diaboloceras*, a typically Atokan form, suggests that the Gene Autry assemblage is transitional between the Morrowan and Atokan Series and fills the gap represented by the Morrowan–Atokan unconformity in the type Morrowan region. That conclusion suggests a slightly younger age than those given in previous correlations of the Gene Autry, and gives more significance to the assemblage than has been recognized previously. A complete description of the ammonoid assemblage is in preparation.

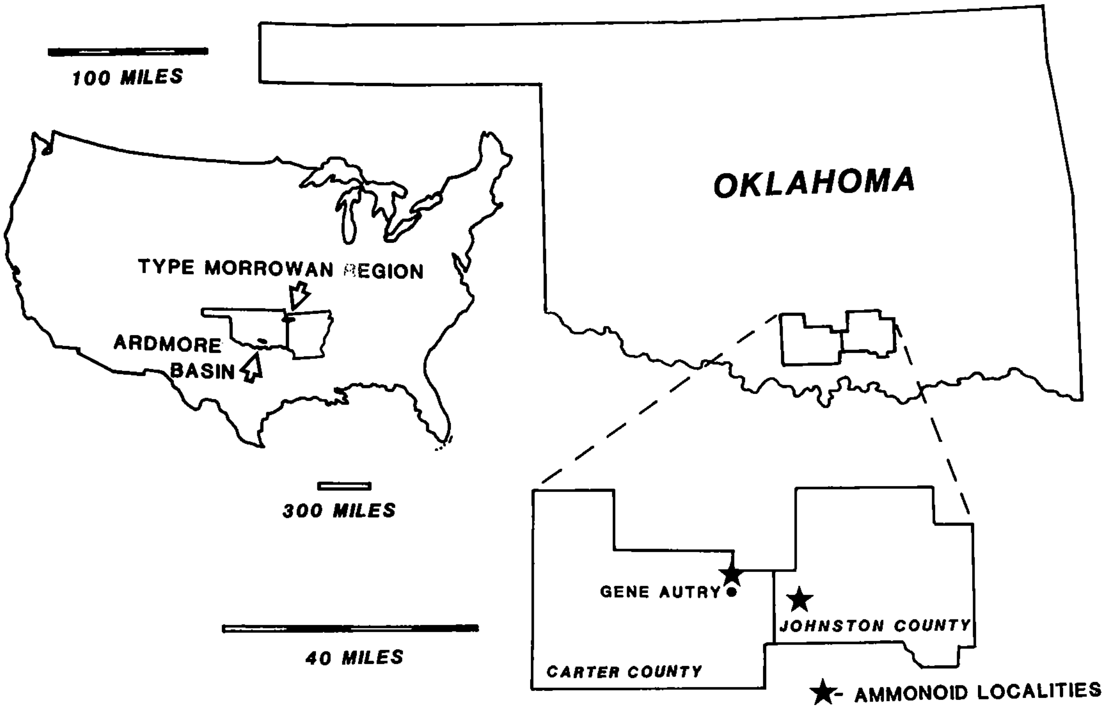
Acknowledgments

Study of the Gene Autry ammonoid assemblage was undertaken with National Science Foundation support (grant EAR 8520684 to Manger), which is gratefully acknowledged. This paper is derived from a thesis by Miller (1985) and was presented by the authors at the Eleventh International Congress on Carboniferous Geology and Stratigraphy held at Beijing, China, in September 1987. Alan Titus, University of Arkansas, helped with the ammonoid photography, and Clifford Cuffey, University of Oklahoma, helped with the figure drafting.

LITHOSTRATIGRAPHY

Lithostratigraphic nomenclature applied to rocks of upper Mississippian and Pennsylvanian age in the Ardmore basin is complex because of common vertical lithologic changes as well as pronounced facies changes between the northern and southern portions of the basin, particularly in the Morrowan section. Terrigenous clastics dominate the section, although carbonates, that may be conglomeratic, are developed as discrete beds at several horizons. Text-figure 2 summarizes Morrowan lithostratigraphy for the Ardmore basin.

The Golf Course Formation was proposed by Harlton (1956) in a redefinition of the Dornick Hills Group, and attains a maximum thickness of ~2,000 ft (Tomlinson and McBee, 1959). The Golf Course



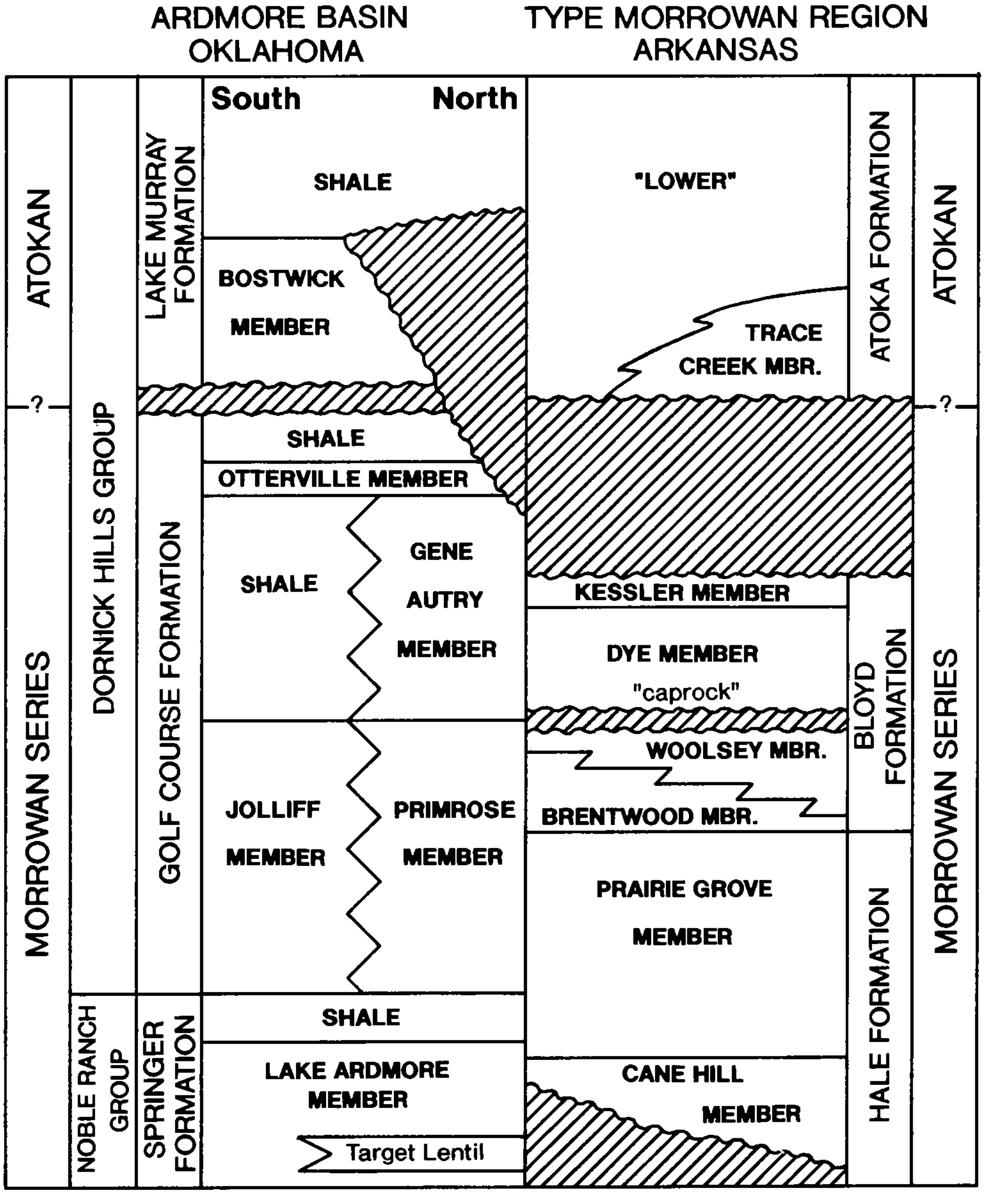
Text-figure 1. Index map for ammonoid localities, southern Oklahoma.

Formation succeeds the Springer Formation as either the Primrose Member in the northern portion of the Ardmore basin, or the Jolliff Member in the south (Text-fig. 2). The Primrose Member varies from 130 to 220 ft, with lower and upper ridge-forming sandstones, separated by gray shales and more thinly bedded, locally calcareous sandstones. The Jolliff Member, as much as 80 ft thick, is a carbonate unit of variable lithology that typically contains limestone-cobble conglomerates derived from the Criner Hills axis to the southwest. Correlation of the Lower Primrose Member and the Upper Prairie Grove Member, Hale Formation in northern Arkansas, is well established (Manger and Saunders, 1980). The Jolliff Member, based on conodonts, is thought to be the facies equivalent of the Primrose Member in the southern Ardmore basin (Text-fig. 2). The correlation of the Jolliff and Primrose with the Prairie Grove and Brentwood Members in northern Arkansas is also supported by conodonts (see Sutherland and Grayson, this volume, for overall discussion).

Both the Primrose and Jolliff Members of the Golf Course Formation are succeeded by shale intervals (Text-fig. 2). In the northern Ardmore basin, the Gene Autry Shale was named by Elias (1956) for ~700 ft of fossiliferous, soft, clay shale with abundant ironstone concretions near the town of Gene Autry (Text-fig. 1). On fresh exposures, the shale is light-gray, but it weathers to a distinctive reddish-brown or maroon color that forms slopes covered with concretions. The Gene Autry is confined to the northern Ardmore basin, but it is thought to be equivalent to unnamed shales in the southern Ardmore basin that lack its distinctive lithologic character, and, hence, application of the name (Text-fig. 2) (Sutherland and Grayson, this volume).

The Otterville Member persists through both the northern and southern Ardmore basin. It contains discrete carbonate beds separated by shales. The carbonates are of highly variable lithology, including oolitic intervals, crinoidal wackestones, and mudstones. It varies in thickness across the area from 30 to 50 ft (Sutherland, 1982).

The top of the Golf Course Formation, above the Otterville, is an unnamed olive-gray shale with ironstone concretions reaching thicknesses of several hundred feet. It is separated by a presumed regional unconformity from sandstones, carbonates, and chert-cobble conglomerates of the Bostwick Member, Lake Murray Formation in the southern Ardmore basin (Text-fig. 2). The interpreted presence of this unconformity rests primarily on the description of advanced *Fusulinella* from the lower portion of the Bostwick in the southern Ardmore basin (Waddell, 1966). Lack of older Lower Atokan faunal elements is thought to represent the unconformity. In the northern Ardmore basin, a more obvious regionally truncating unconformity exists within the Atokan Series, where the

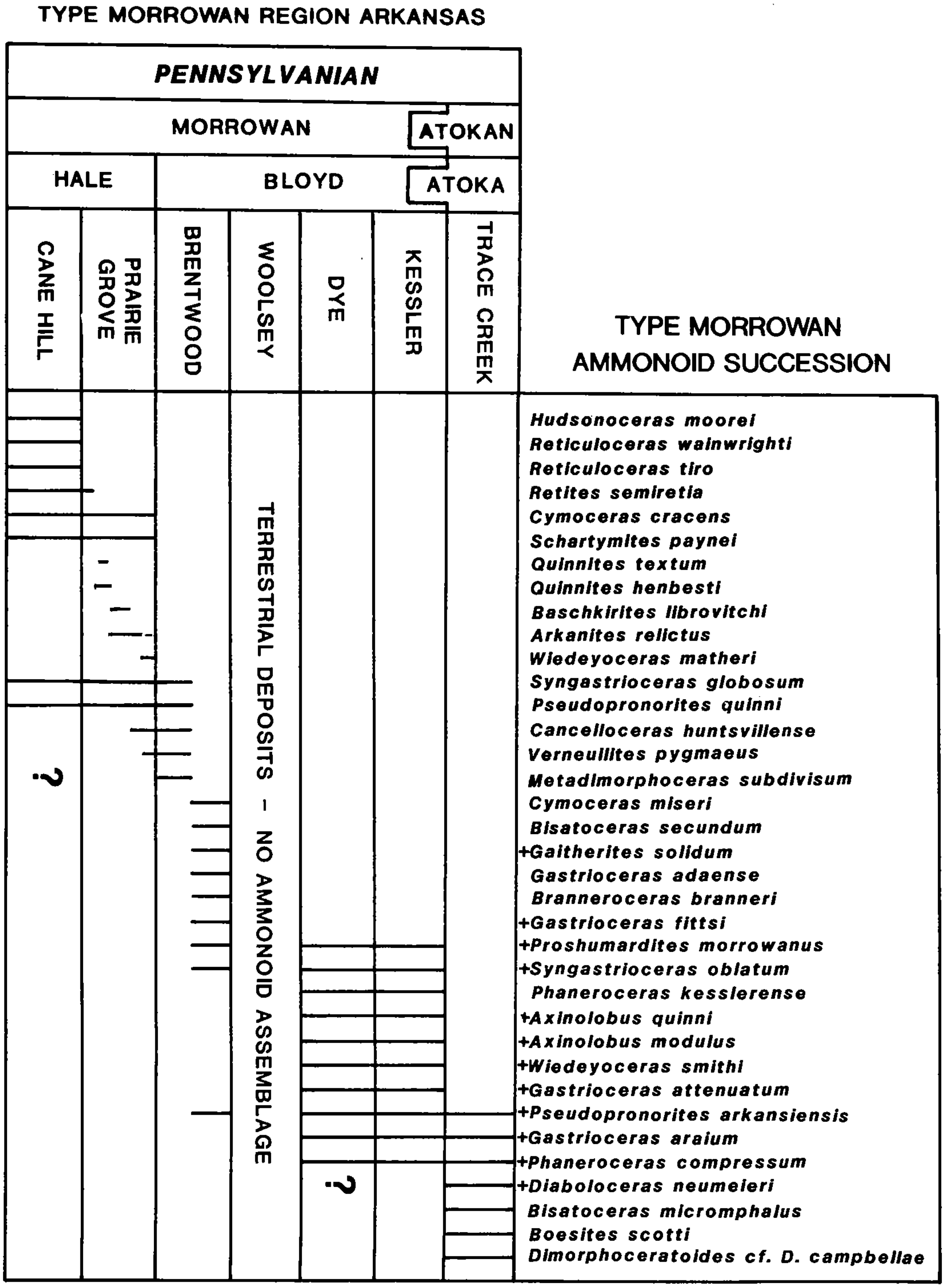


Text-figure 2. Correlation of Morrowan sections between Ardmore basin and type Morrowan region (based in part on Sutherland and Grayson, this volume, text-fig. 2).

Bostwick Member has been removed by erosion and higher Atokan shales succeed horizons in the Golf Course Formation as low as the Gene Autry Shale (Text-fig. 2) (Sutherland, 1988; Sutherland and Grayson, this volume).

GENE AUTRY AMMONOID ASSEMBLAGE

The age of the Gene Autry ammonoid assemblage is based on an understanding of the Morrowan ammonoid succession and its correlation that has been refined over the past two decades, primarily based on studies in the type region, northwestern Arkansas, by Gordon (1965), McCaleb (1968), Saunders and others (1977), and Manger and Saunders (1980). Six range zones based on distinctive and well-documented species of the genera *Retites, Quinnites, Arkanites, Verneuilites, Branneroceras,* and *Axinolobus* are recognized through the Hale and Bloyd Formations, which constitute the type Morrowan section. There are obvious discontinuities in the succession at the base of the Prairie Grove, within the Brentwood, and at the



Text-figure 3. Type Morrowan ammonoid succession. Gene Autry ammonoids indicated by +.

base of the Woolsey Members (Text-fig. 3). The first three zonal genera are also endemic. Nevertheless, the stage of sutural and ornamentational development, combined with the association of other cosmopolitan taxa provide a firm basis for correlations throughout North America, as well as intercontinentally (Manger and Saunders, 1982). Text-figure 3 summarizes the type Morrowan ammonoid assemblage.

Approximately half of the Gene Autry ammonoid assemblage consists of *Axinolobus*, as *A. modulus*, and less commonly *A. quinni*. The remainder of the assemblage, in order of decreasing abundance, includes *Gastrioceras araium*, *G. attenuatum*, *G. fittsi*, *Wiedeyoceras smithi*, *W.* n. sp., *Gaitherites solidum*, *G.* n. sp., *Diaboloceras neumeieri*, *Proshumardites morrowanus*, *Syngastrioceras oblatum*, *Anthracoceratoides serratoides*, *Phaneroceras compressum*, *Pseudopronorites arkansiensis*, and several new taxa. All of those taxa are typical Morrowan forms except *Anthracoceratoides*, which is rare in the Gene Autry, and *Diaboloceras neumeieri*, which is the zonal name-bearer of the succeeding lower Atokan assemblage (Text-fig. 3). Plate 1 illustrates some typical representatives of the Gene Autry ammonoid assemblage.

Axinolobus modulus is the zonal name-bearer for the highest ammonoid range zone recognized in the type Morrowan succession (Saunders and others, 1977). *Axinolobus* appears as a cryptogene in the Bloyd Formation above an unconformity at the base of the "caprock," which marks a return to marine conditions at the base of the Dye Member. It does not cross the Morrowan–Atokan boundary at the top of the Kessler Member in northern Arkansas (Text-fig. 3). *Axinolobus* is the most abundant form in the Gene Autry assemblage and its presence alone would seem to confirm correlation of the Gene Autry and the upper Bloyd Formation (McCaleb and Furnish, 1964; McCaleb, 1968). In addition, the upper Bloyd and Gene Autry Shales have *Proshumardites morrowanus*, *Syngastrioceras oblatum*, *Wiedeyoceras smithi*, *Gastrioceras attenuatum*, *Pseudopronorites arkansiensis*, *G. araium*, and *Phaneroceras compressum* in common. *Gaitherites solidum* and *Gastrioceras fittsi* are older, upper Brentwood Member elements found in the Gene Autry (Text-fig. 3).

An alternative conclusion for the age of the Gene Autry assemblage is possible, if more significance is placed on the appearance of *Diaboloceras neumeieri*. That taxon was originally recovered from the Trace Creek Shale (Quinn and Carr, 1963). The Trace Creek Shale was traditionally included as part of the Bloyd Formation, type Morrowan, but has been removed as part of the Atoka Formation because of the recognition of a regional unconformity at its base (Sutherland and others, 1978). Thus, the Kessler–Trace Creek contact is now regarded as the Morrowan–Atokan boundary in northern Arkansas. Although *Diaboloceras* has been reported from Morrowan strata below the Trace Creek Member in north-central Arkansas (Greers Ferry Dam and Mill Creek sections of Quinn and Carr, 1963; McCaleb, 1968), we have been unable to confirm those identifications. Lithostratigraphic assignment of the Greers Ferry section is problematic as well. Consequently, in northern Arkansas, we currently regard *Diaboloceras* as restricted to the Atokan Series (Text-fig. 3).

Detailed measured sections of the Gene Autry in Johnston County (Text-fig. 1) (Maley, 1986) demonstrate that *Diaboloceras* joins *Axinolobus* and other typically Morrowan ammonoids in the upper half of the unit. Yet, no other unequivocal Atokan elements occur in the Gene Autry. The succeeding Otterville Member contains a meager, but nevertheless Morrowan, conodont and brachiopod assemblage (see Sutherland and Grayson, this volume). We conclude that the portion of the Gene Autry Shale yielding *Axinolobus* without *Diaboloceras* is equivalent to the Dye and Kessler Members, Bloyd Formation, type Morrowan section, as has been proposed previously (McCaleb, 1968). However, the association of *Axinolobus* with *Diaboloceras* found in the upper Gene Autry Shale does not occur in the type Morrowan succession, where it is represented by the gap at the Morrowan–Atokan unconformity (Text-fig. 2). *Diaboloceras*, as *D. neumeieri*, spans the Morrowan–Atokan boundary, while the genus ranges into Desmoinesian strata in the Canadian Arctic (Nassichuk, 1975).

REFERENCES CITED

Elias, M. K., 1956, Upper Mississippian and Lower Pennsylvanian formations of south-central Oklahoma, *in* Petroleum geology of southern Oklahoma, vol. 1: American Association of Petroleum Geologists, p. 56–134.

Gordon, M., Jr., 1965, Carboniferous cephalopods of Arkansas: U.S. Geological Survey Professional Paper 460, 322 p.

Harlton, B. H., 1956, The Harrisburg trough, Stephens and Carter Counties, Oklahoma, *in* Petroleum geology of southern Oklahoma, vol. 1: American Association of Petroleum Geologists, p. 137–139.

Maley, M. P., 1986, Depositional history of the Upper Morrowan (Pennsylvanian) strata of the Ardmore basin, Oklahoma: University of Oklahoma unpublished M.S. thesis, 206 p.

Manger, W. L.; and Saunders, W. B., 1980, Lower Pennsylvanian (Morrowan) ammonoids from the North American Midcontinent: Journal of Paleontology Memoir 10, 56 p.

______1982, An ammonoid-based Middle Carboniferous boundary, *in* Biostratigraphic data for a mid-Carboniferous boundary: International Union of Geological Sciences Subcommission on Carboniferous Stratigraphy Publication 11, Leeds, England, p. 95–100.

McCaleb, J. A., 1968, Lower Pennsylvanian ammonoids from the Bloyd Formation of Arkansas and Oklahoma: Geological Society of America Special Paper 96, 123 p.

McCaleb, J. A.; and Furnish, W. M., 1964, The Lower Pennsylvanian ammonoid genus *Axinolobus* in the southern Midcontinent: Journal of Paleontology, v. 38, p. 249–255.

Miller, M. S., 1985, Ammonoid biostratigraphy of the Gene Autry Shale (Morrowan–Pennsylvanian), south-central Oklahoma: University of Arkansas unpublished M.S. thesis, 130 p.

Nassichuk, W. W., 1975, Carboniferous ammonoids and stratigraphy in the Canadian Arctic Archipelago: Geological Survey of Canada Bulletin 237, 240 p.

Quinn, J. H.; and Carr, L. C., 1963, New Pennsylvanian *Diaboloceras* from northwest Arkansas: Oklahoma Geology Notes, v. 23, p. 111–118.

Saunders, W. B.; Manger, W. L.; and Gordon, M., Jr., 1977, Upper Mississippian and Lower and Middle Pennsylvanian ammonoid biostratigraphy of northern Arkansas, *in* Sutherland, P. K.; and Manger, W. L. (eds.), Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, p. 117–137.

Sutherland, P. K. (ed.), 1982, Lower and Middle Pennsylvanian stratigraphy in south-central Oklahoma: Oklahoma Geological Survey Guidebook 20, 44 p.

______ 1988, Ardmore and Marietta basins, *in* Southern Midcontinent region, *chapter 12 of* Sloss, L. L. (ed.), Sedimentary cover—North American craton; U.S.: Geological Society of America, The Geology of North America, v. D-2, p. 326–331.

Sutherland, P. K.; Grayson, R. C., Jr.; and Zimbrick, G. D., 1978, Reevaluation of Morrowan–Atokan Series boundary in northwestern Arkansas [abstract]: American Association of Petroleum Geologists Bulletin, v. 62, p. 566.

Tomlinson, C. W.; and McBee, W., Jr., 1959, Pennsylvanian sediments and orogenies of Ardmore district, Oklahoma, *in* Petroleum geology of southern Oklahoma, vol. 2: American Association of Petroleum Geologists, p. 3–52.

Waddell, D. E., 1966, Pennsylvanian fusulinids in the Ardmore basin, Love and Carter Counties, Oklahoma: Oklahoma Geological Survey Bulletin 113, 128 p.

Plate

PLATE 1

All specimens from the Gene Autry Shale. OU designates those reposited at the University of Oklahoma. SUI designates those reposited at the University of Iowa. Details of localities may be found in Miller (1985).

Figure 1.—*Gaitherites solidum* (Quinn), OU 9916, locality 2, Johnston County, x1.3.

Figure 2.—*Gastrioceras fittsi* Miller and Owen, SUI 49896, locality 9, Carter County, x1.3.

Figure 3.—*Wiedeyoceras smithi* McCaleb, OU 9917, locality 12, Carter County, x1.3.

Figure 4.—*Proshumardites morrowanus* Gordon, OU 9918, locality 12, Carter County, x1.3.

Figures 5,9.—*Diaboloceras neumeieri* Quinn and Carr. **5**, OU 9919, locality 13, Johnston County, x1.8. **9**, OU 9920, locality 3, Johnston County, x0.75.

Figure 6.—*Gastrioceras araium* McCaleb, OU 9921, locality 13, Johnston County, x1.

Figures 7,11.—*Axinolobus quinni* McCaleb and Furnish. **7**, OU 9922, locality 13, Johnston County, x1.3. **11**, OU 9923, locality 13, Johnston County, x0.50.

Figure 8.—*Gastrioceras attenuatum* McCaleb, OU 9924, locality 13, Johnston County, x1.25.

Figure 10.—*Phaneroceras compressum* (Hyatt), OU 9925, locality 13, Johnston County, x0.50.

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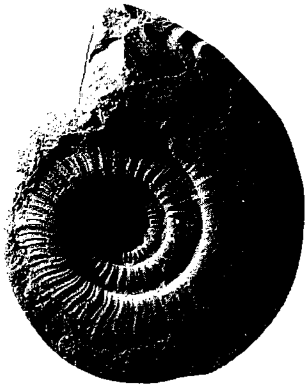

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Oklahoma Geological Survey Circular 94, 1992

Atokan and Basal Desmoinesian Conodonts from Central Iowa, Reference Area for the Desmoinesian Stage

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University of Iowa

ABSTRACT.—Atokan conodonts and palynomorphs are present in strata at the base of the type Des Moines succession in central Iowa, the reference area for the Desmoinesian Stage. The Atokan portion of that section comprises the Kilbourn Formation and part of the overlying Kalo Formation. Based on palynomorphs, the Iowa Geological Survey Bureau places the Atokan/Desmoinesian boundary at the base of the Cliffland coal within the Kalo Formation. This boundary interval can also be recognized by the appearance of the conodonts *Idiognathodus amplificus* n. sp., *I. obliquus*, and *Neognathodus caudatus* n. sp.

INTRODUCTION

The Atokan and Desmoinesian Stages are chronostratigraphic units loosely based on poorly defined type regions, rather than on precisely measured and described type sections. Keyes (1893) originally designated the Des Moines Formation (now Supergroup) to include rocks representing the entire Lower and Middle Pennsylvanian Series (*sensu* Zeller, 1968). Later studies eventually recognized the Morrowan and Atokan as separate units, reducing the duration of the Desmoinesian and restricting its "type region" to the outcrop belt along the Des Moines River within the boundaries of Iowa.

Moore and Thompson (1949) characterized the Desmoinesian Stage as coincident with the Zone of *Beedeina* from the standard North American fusulinid zonation. The ancestral form, *Beedeina insolita* (Thompson), was recovered in association with *Fusulinella iowensis* from the lower Des Moines Supergroup in a locality where its basal units thin by non-deposition over a Mississippian high. Because a significantly more complete lower section is exposed in various localities along the type Des Moines outcrop belt, the concept of the basal Desmoinesian boundary remains unsettled. These older deposits yield fossils regarded as representing the basal Desmoinesian by some workers, but as Atokan by others, rendering recognition of the Atokan/Desmoinesian boundary problematical. This and other Atokan/Desmoinesian boundary problems have been discussed recently by Lambert and Heckel (1991); the present paper is a preliminary examination of conodonts from the basal type Des Moines Supergroup.

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LITHOSTRATIGRAPHY

Pennsylvanian lithostratigraphic nomenclature in the Midcontinent historically follows that of Searight and others (1953), although most states have initiated some revisions. Landis and Van Eck (1965) summarized the terminology used in Iowa to that time. Based on palynological studies, Ravn and others (1984) further revised the Pennsylvanian nomenclature for Iowa as part of an assessment of coal resources in the state. Their nomenclatural revisions have been adopted by the Iowa Geological Survey Bureau and are followed herein. Their concepts provide greater resolution of the lower Des Moines (Cherokee Group) section, although application beyond the boundaries of Iowa awaits further study and refinement.

Ravn and others (1984) considered the Des Moines to be a Supergroup comprising the Cherokee and overlying Marmaton Groups. The Cherokee Group designates a predominantly nonmarine succession punctuated by variable and often laterally discontinuous marine horizons. Ravn and others (1984) proposed (in ascending order) the Kilbourn, Kalo, Floris, and Swede Hollow Formations as divisions based on stratigraphic packages recognized by palynological characteristics of their contained coals. Although developed from a subsurface data base, each unit was recognized in outcrop and a type section was designated.

The Kilbourn Formation records initial Pennsylvanian deposition on a heavily dissected, Mississippian karst surface. The Mississippian/Pennsylvanian boundary is typically marked by a thick residual paleosol or a conglomerate containing Mississippian clasts. The Kilbourn is characterized by various discontinuous nonmarine and marginal marine lithofacies. It developed principally as a topographic fill and ranges from zero across Mississippian highs to 220 ft thick over lows.

The succeeding Kalo Formation was defined by Ravn and others (1984) as the interval from the base of the Blackoak coal to the base of the Laddsdale coal complex. Kalo lithofacies are similarly variable, but two persistent coals were recognized as "members": the Blackoak coal at the base of the formation, and the Cliffland coal near its top. Marginal marine deposits dominate the Kalo, reflecting periodic transgressions over widespread coal swamps that developed in response to rising baselevel. Small, constructive deltas represent local clastic influxes that markedly thicken the interval at some localities. The Kalo Formation has a maximum thickness of 120 ft.

Although the lithostratigraphic framework is based on objective core analysis and type sections, most correlations reflect palynomorph assemblages, and some problems exist. The Blackoak and Cliffland coals contain palynomorph assemblages that differ only slightly, suggesting peat accumulation under similar paleoenvironmental conditions. The succeeding Laddsdale coals in the Floris Formation are also not considerably different (Ravn, 1986). The Cliffland is differentiated from the Blackoak primarily by decreasing miospore diversity, notably the extinction of *Dictyotriletes bireticulatus* (Ibrahim) among other forms (Ravn and others, 1984). The Laddsdale coals contain the earliest common occurrences of *Microreticulatisporites sulcatus* (Wilson and Kosanke) and *Mooreisporites inusitatus* (Kosanke). Presence or absence of particular miospores strongly influenced the stratigraphic assignment of coals in several cores, although Ravn (1979) had emphasized the influence of ecology on the palynomorphs preserved within any given coal.

CONODONT BIOSTRATIGRAPHY

Previous Investigations

In contrast to relatively extensive study of conodont assemblages from the upper Cherokee and Marmaton Groups, the only published record of lower Cherokee conodonts from the type Des Moines was the fortuitous recovery of several specimens from a coal ball collected near Oskaloosa, Iowa (Mamay and Yochelson, 1962). Theses by Gleim (1955) and Kastler (1958) were used to identify measured sections with conodont-bearing horizons and to help with the resolution of nomenclatural lithostratigraphic problems. Gregory (1982) processed samples from the section between the Blackoak and Cliffland coals in four cores from the study area, but conodonts were rare and mostly juveniles. Collections from cores drilled for the coal resources project by the Iowa Geological Survey Bureau were extensively used for the present investigation.

Element Occurrence and Diversity

Conodont elements were most abundant in fissile, black shales and dense, fetid limestones. Light-colored marine shales and limestones generally had lower abundances, whereas conodonts were largely absent in marginal marine to nonmarine clastics. The lower Kilbourn Formation contains only rare marine intervals that produced poorly preserved, depauperate assemblages, often dominated by juvenile elements. The upper Kilbourn and lower Kalo reflect greater marine influence, with marine horizons that usually produce abundant and diverse conodont assemblages. Complete ontogenetic suites were common, and ramiform elements were sometimes well represented. The marine horizon overlying the Blackoak coal is an exception, producing predominantly juvenile faunas (duplicating Gregory's results). A significant lacuna is suggested by a well-developed paleosol and dramatic change in conodont assemblages between underlying strata and the marine interval overlying the Cliffland coal. Conodonts from the latter are frequently the most abundant and diverse of the entire lower Cherokee Group.

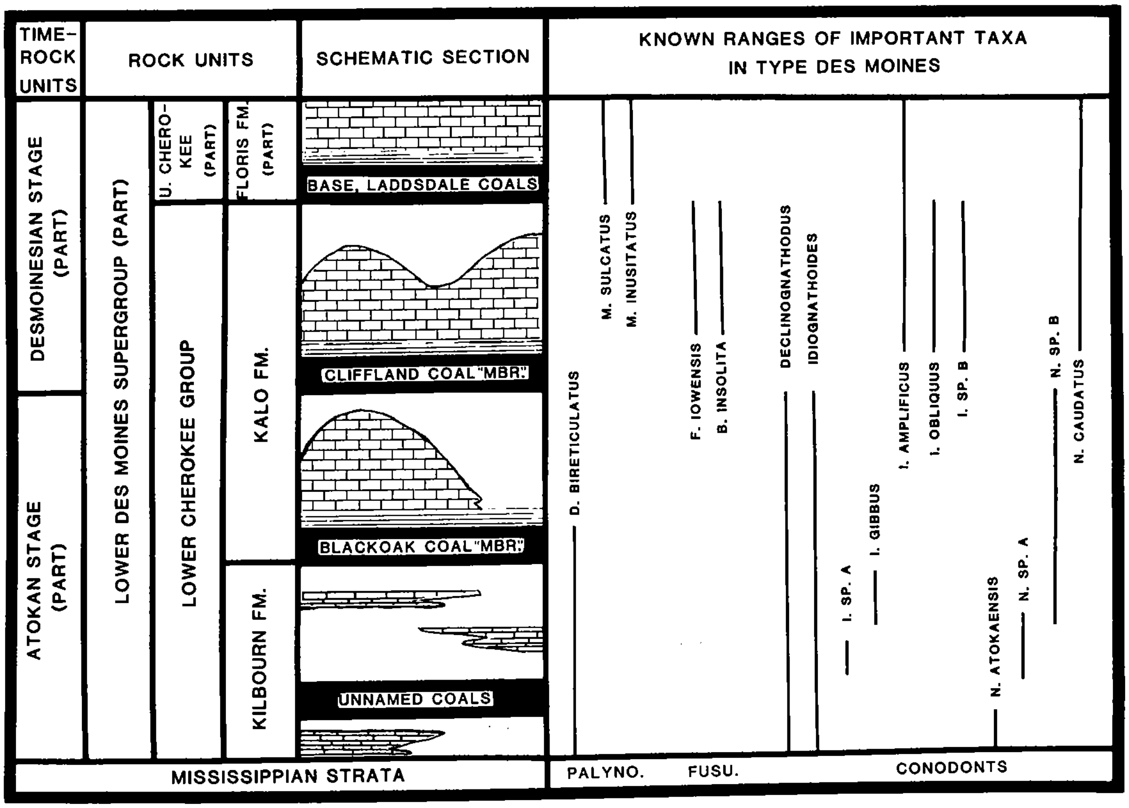
Most known Middle Pennsylvanian genera occur in the lower Cherokee, and most species present can be referred to known taxa. *Diplognathodus* occurs sporadically, but when present it is usually abundant. *Gondolella* was recovered from a single core sample and is represented by several broken platforms and a few ramiform elements. These specimens may represent *G. gymna* Merrill and King. Biostratigraphically significant genera are *Idiognathoides, Declinognathodus, Neognathodus,* and *Idiognathodus. Idiognathoides* and *Declinognathodus* are abundantly represented in the Kilbourn and lower Kalo Formations, but do not occur above the marine interval overlying the Blackoak coal.

Neognathodus is represented through the Kilbourn and Kalo Formations by four distinctive species. Rare specimens of *N. atokaensis* Grayson occur in the lower Kilbourn. *Neognathodus* sp. A, a form previously assigned to either *N. bothrops* Merrill or *N. bassleri* (Harris and Hollingsworth), first occurs in the upper Kilbourn and exhibits characters transitional with the succeeding *N.* sp. B (*N. "bothrops" sensu* Grayson, 1990). *Neognathodus* sp. B is the only neognathodid recovered from the marine interval above the Blackoak coal, although elsewhere it has greater range overlap with *N.* sp. A.

Neognathodus caudatus n. sp. appears in the marine strata overlying the Cliffland coal (upper Kalo Formation).

Specimens of *Idiognathodus* are too poorly represented in the lower Kilbourn to provide good control on correlation or more than general stratigraphic position. However, *I.* sp. A is a distinctive form recovered from two cores in association with *Neognathodus* sp. A low in the upper Kilbourn. *Idiognathodus gibbus* n. sp. occurs in the uppermost Kilbourn in three of the 11 cores studied, two of which also record its range overlapping that of both *Neognathodus* spp. A and B (Text-fig. 1). No species of *Idiognathodus* can be confidently recognized in the stratigraphic interval directly above the Blackoak coal, due to the predominance of juveniles and an inconsistent distribution of adult morphotypes.

The Atokan/Desmoinesian boundary is placed at the base of the Cliffland coal "member" of the Kalo Formation by the Iowa Geological Survey Bureau. The horizon selected is based on palynological data derived from surface and subsurface coals and their correlatives in the Illinois basin (Ravn and others, 1984), and closely approximates the level proposed by Thompson (1934) for the basal Desmoinesian based on fusulinids. The marine interval above the Cliffland coal produces a distinctive *Idiognathodus* fauna. *Idiognathodus amplificus* n. sp., *I. obliquus* Kossenko and Kozitskaya, and *I.* sp. B. all appear at the base of those marine deposits, along with *Neognathodus caudatus* n. sp. *Idiognathodus amplificus* n. sp. is a distinctively elongated species that ranges higher in the Cherokee Group. *Idiognathodus obliquus* Kossenko and Kozitskaya is important for correlation purposes. It was originally described from the Donets basin, U.S.S.R., and may also have been recovered from Japan (Koike, 1967). Because of its short range and wide distribution, this taxon could serve for delineating the Atokan/Desmoinesian boundary (Lambert and Heckel, 1991). *Idiognathodus* sp. B. is another distinctive form with a limited range. The ranges of biostratigraphically important taxa are summarized in Text-figure 1; outcrop/core localities are listed in the Appendix.



Text-figure 1. Stratigraphic and biostratigraphic summary for lower Des Moines Supergroup. Fine horizontal lines represent unnamed black shales; regular carbonate symbol represents persistent marine deposits; irregular carbonate symbol represents impersistent, impure marine deposits; unpatterned area represents marginal and nonmarine deposits in schematic section. "Palyno." underlies palynomorph species ranges, and "Fusu." underlies fusulinid ranges. "*I.*" stands for *Idiognathodus*, "*N.*" for *Neognathodus*, "*F.*" for *Fusulinella* and "*B.*" for *Beedeina*.

SYSTEMATIC PALEONTOLOGY

Genus **Idiognathodus** Gunnell, 1931

Type species.—Idiognathodus claviformis Gunnell, 1931 (original designation).

Idiognathodus amplificus n. sp.
Pl. 2, Figs. 1–12

Diagnosis.—A species characterized by Pa elements with an elongated, posterior platform bearing disrupted transverse ridges. The basal cavity is extremely narrow and pointed posteriorly, but anteriorly flares into a broad cavity. In left elements, the short anterior platform is composed of accessory lobes that drape posteriorly around the distal platform. Right elements are distinguished by less developed accessory lobes that do not extend posteriorly and a gently steepening distal platform in side view.

Description.—Elements display Class IIIb symmetry (Lane, 1968). Free blade intersects central platform, and carina is extremely short, if present. Adcarinal ridges are extremely short, with abrupt anterior termination. An accessory lobe is well developed on the inner side, extending from adcarinal ridges to halfway down the platform, consisting mostly of discrete nodes but also nodes coalesced into short ridges. Nodes develop on the broad outer margin from the posterior end of the adcarinal ridges, extending halfway along the transverse ridge complex, often half again as far as the inner accessory lobe. Platform margins are rounded. The transverse ridge complex comprises two-thirds to greater than three-fourths of the platform and is commonly disrupted, dividing into nodes parallel with ridge axes. An incipient trough may develop across posterior-most ridges in large specimens. The posterior end is sharply to bluntly pointed.

In right elements the adcarinal ridges are slightly longer and do not parallel the carina, producing an open slot on the inner side, closed on the outer side. The accessory lobes are less well developed, and the outer lobe does not extend as far posteriorly. Thus rights appear more symmetrical anteriorly and more elongate overall.

Platforms high, with height/width <1. Posterior slope is gentle at ~20°, with no sharp breaks. Right elements display increasing curvature, up to 45°.

Basal cavity deep, asymmetrical, with sub-central deepest point in anterior one-third, and furrow straight to only slightly deflected. Inner cup margin extends from posterior tip about halfway to deepest point at 20–30°, then swings to 40–60° to form base of oblate semicircle beneath inner accessory lobe. Outer cup margin extends a short distance at 30°, then abruptly turns to 40–45° to form larger semicircle, although sometimes extending from posterior tip at a constant 30° until meeting the base of the semicircle. Both sides distinctively flattened anteriorly, with greater area beneath the outer cup margin.

Remarks.—*Idiognathodus amplificus* is a highly variable taxon but maintains its integrity within the confines listed above. It is easily differentiated from *I. obliquus* by its elongated, relatively straight platform and its high height/width ratio. It is differentiated from *I.* sp. B by its well-developed accessory lobes and by its disproportional basal cavity.

Etymology.—Latin *amplifico,* meaning enlarge, increase, extend; in reference to the extended distal platform.

Material.—66 specimens; 27 lefts, 39 rights.

Holotype.—SUI 61563, Plate 2, no. 1.

Idiognathodus gibbus n. sp.
Pl. 2, Figs. 23–33

?*Idiognathodus delicatus* Gunnell. Rabe, 1977, p. 181–187, pl. 3, figs. 18–21 and 23–26 only.

?*Streptognathodus* sp. aff. *S. wabaunsensis* Gunnell. Grubbs, 1984, p. 71–72, pl. 1, figs. 8–11.

Diagnosis.—A species characterized by a Pa element with moderately long carina and adcarinal ridges that meet the distal platform at a pronounced angle. The transverse ridges exhibit a subtle bulge to the outer side and commonly bear an incipient posterior trough. The distinctively shallow basal cavity exhibits a subtle bulge to the outer side corresponding with that of the overlying transverse ridges.

Description.—Species possesses Class IIIb symmetry. Free blade meets platform centrally, continues onto platform as a moderately long carina, terminating at anterior-most transverse ridge. Long, nodose adcarinal ridges parallel carina for a considerable distance (one-third to one-half total platform length) and are usually of equal length, or inner ridge slightly longer. Adcarinal grooves are shallow. Nodose accessory lobe developed on inner side adjacent to carina and proximal transverse ridges. A few nodes may be present on the outer side. Proximal platform margins are rounded. Transverse ridge complex is broader on outer side (middle to middle-upper one-third of transverse ridges), producing a bulging appearance. An incipient sub-central trough may form across the transverse ridges. Posterior terminus bluntly pointed to sharply rounded.

Right elements are similar, but adcarinal ridges and carina in right elements may be shorter. They show greater variation in the ridge complex and often lack the outer bulge. The outer margin of the transverse ridge complex is sometimes aligned with the outer adcarinal ridge, producing a continuous prominence marking the outer platform margin.

The proximal platform is nearly horizontal; the distal platform is inclined 30–40°. Platform is low, height/width <<1. Right element sometimes slightly thicker (h/w still <1).

Basal cavity is shallow, very slightly asymmetrical, with deepest point centered in proximal one-third of straight furrow. Basal cavity expands anteriorly to inner side (coinciding with accessory lobe), posteriorly to outer side (coinciding with bulge of transverse ridges). Right element cavity less shallow, furrow sometimes deflected to inner side in distal one-half of platform.

Remarks.—Idiognathodus gibbus is a short ranging taxon. It can be compared with several forms illustrated previously, but none can be confidently assigned because all characters are not described or figured. Several specimens from Bucaramanga sample 43 of Rabe (1977) could represent *I. gibbus*, but may also represent an earlier species. Without lower and side views, no definitive comparison can be made. Similarly, Grubbs (1984) illustrated four specimens in oblique view that could belong to *I. gibbus*. Those specimens, *Idiognathodus gibbus*, and several Morrowan forms may belong to a discrete lineage of idiognathodids first represented by *Streptognathodus expansus* Igo and Koike.

Etymology.—Latin *gibber*, meaning humped, protuberant, swollen; in reference to the bulge in the outer transverse ridges and in the basal cavity.

Material.—170 specimens; approximately equal right and left representation.

Holotype.—SUI 61580, Plate 2, no. 26.

Idiognathodus obliquus,
Kossenko and Kozitskaya, 1978
Pl. 2, Figs. 13–22

?*Idiognathodus delicatus* Gunnell. Koike, 1967, p. 48–49, pl. 2, figs. 19,21–23 only.
?*Idiognathodus delicatus* Gunnell. Rabe, 1977, p. 181–187, pl. 3, figs. 5,7,9 only.
Idiognathodus obliquus Kossenko and Kozitskaya, 1978, p. 51–53, pl. 22, figs. 6–9.
Idiognathodus sp. morphotype B. Lambert and Heckel, 1991, fig. 5-4.

Diagnosis.—A species characterized by a sigmoidal silhouette in oblique inner view, because of pronounced deflection of the posterior platform and its inward curvature ("oblique" transverse ridges). Long adcarinal ridges flare anteriorly and downwards, conspicuous in both upper and lower views.

Description.—Elements display Class II symmetry. Free blade meets platform centrally. Nodose adcarinal ridges parallel carina in anterior platform; at proximal margin they sweep outwards and downwards, to continue as an anterior flange up to one-third the length of the free blade (equivalent to the anterior platform length). Proximal platform bears well-developed accessory lobes on both sides, inner slightly more well developed, outer generally following the platform curvature. Anterior margins rounded at the accessory lobes. Transverse ridges are well developed and distinctly segregated from the accessory lobes. Distal platform distinctly deflected inwards at 120–150°, with a variable but pronounced downward curvature. Posterior terminus is sharply pointed.

Elements very low (h/w <<1). No intervening area is exposed between the upper surface and cup. Anteriorly directed adcarinal ridges continue first as a flange, then as a bulge on the side of free blade, forming a thickened region that does not fade except beneath the anterior-most few denticles. Distal platform curves smoothly downward and inward at 40–50°; the upper distal platform surface lies within the inner view.

The basal cavity is moderately deep and is roughly equidimensional around the furrow, although slightly broader on the outer side due to a straight portion on the distal inner side. The deepest point lies in the anterior one-third of furrow. The adcarinal ridges produce a distinctive thickening of the lower posterior free blade.

Remarks.—Although several species could be selected to mark the Atokan/Desmoinesian boundary, *I. obliquus* is a promising candidate because it is striking morphologically and appears to be both widespread and short-ranging. Its distinctive upper surface curvature and conspicuously flared adcarinal ridges serve to distinguish it from both *I. amplificus* and *I.* sp. B.

Idiognathodus sp. A
Pl. 1, Figs. 28–30

Remarks.—The Pa element of *Idiognathodus* sp. A is characterized by anastomosing ridges on the proximal platform. No true accessory lobes are developed, but rare small nodes occur on the outer margin of a juvenile specimen. The carina and adcarinal ridges rapidly lose their integrity posteriorly, grading into the anastomosing ridges which in turn grade into transverse ridges. The platform is divided into roughly equal proximal and distal halves. The inner margin of the transverse ridge complex is poorly defined. The element is moderately high (h<w). The distinctly asymmetrical basal cavity has a convex outer margin and a sigmoidal inner margin. Although there are insufficient specimens to delineate variability in *I.* sp. A, it is important as a distinctive morphotype that precedes *I. gibbus* in the Kilbourn Formation. It occurs with *Neognathodus* sp. A and denotes the lower part of the upper Kilbourn Formation. Manger and Sutherland (1984) illustrated a similar specimen (their pl. 1, fig. 17) from the upper Morrowan of central Texas.

Idiognathodus sp. B
Pl. 1, Figs. 17, 21–23

Remarks.—The Pa element of *Idiognathodus* sp. B bears a relatively short carina and adcarinal ridges. The latter terminate abruptly anteriorly, producing a distinctive right angle in side view. A marginal accessory lobe is present on the inner anterior platform; nodes are rare on the outer margin. *Idiognathodus* sp. B is characterized by inconsistently spaced transverse ridges that split into small nodes on the posterior platform margins. The transverse ridges are most widely spaced at the posterior platform flexure, giving an impression of rapid upward growth that stretched the middle ridge complex. This impression is reinforced by continuation of the transverse ridges across the steep posterior tip in a few specimens. In most specimens the transverse ridge complex does not extend far below the posterior flexure, producing a squared posterior end in side view. In upper view, the posterior end is bluntly rounded, except that of the former specimens, which are more pointed. Elements are high (h=w), and bear basal cavities with expanded outer posterior margins. *Idiognathodus* sp. B is easily recognized by its bloated appearance, and is represented mostly by juveniles and subadults.

Genus **Neognathodus** Dunn, 1970

Type species.—*Polygnathus bassleri* Harris and Hollingsworth, 1933.

Neognathodus caudatus n. sp.
Pl. 1, Figs. 1–16

Gnathodus bassleri (Harris and Hollingsworth). Mamay and Yochelson, 1962, pl. 34, fig. 45.
Neognathodus sp. A. Lambert and Heckel, 1991, figs. 5-1, 5-2.

Diagnosis.—A species characterized by a flat-topped, wedge-shaped Pa element with a conspicuously elevated carina extending to the posterior tip, often nodose at the distal end. In lower view the basal cavity has an angular proximal inner margin and pointed posterior terminus.

Description.—Elements exhibit Class II symmetry. Free blade meets wedge-shaped platform centrally, but often at an angle up to 20°, or rarely in an arc; the point of inflection is always the node over the deepest point in basal cavity. The free blade continues to posterior terminus as a pronounced carina. Transverse ridges extend from platform margins to carina. Adcarinal grooves absent; platform characterized by flattened upper surface, with carina elevated above or equivalent with the margins. Inner, outer, and anterior platform margins steep. Inner transverse ridges extend downward onto anterior platform margin, usually deflected away from free blade ~30°. Outer transverse ridges usually do not extend as far anteriorly; often they terminate at, or just past, a perpendicular transverse node centered over the outer anterior basal cavity. New posterior transverse ridges begin as nodes on the side of the carina, then grow towards margins. This results in a nodose posterior carina in many specimens, reminiscent of the Mississippian genus *Gnathodus*.

Neognathodus caudatus comprises two morphotypes. In one, the inner and outer margins meet just to the anterior of the carinal posterior terminus. Transverse ridges maintain their integrity over the entire platform surface. The other morphotype typically maintains a wedge shape and long carina, but the transverse ridges may abruptly terminate anywhere along the platform margin, never approaching the posterior terminus or, in some cases, the proximal margin. Ornamentation varies considerably in these forms, including coalescing transverse ridges and unusual locations for perpendicular nodes.

Elements are high (h/w =1 at mid-platform), and proximal ornamentation is often highlighted in side view.

The basal cavity is deep, with the deepest point just to the posterior of the free blade. It is distinctly asymmetrical and sharply pointed posteriorly. The outer margin is smoothly triangulate anteriorly, sharply pointed posteriorly (at <10° out from the posterior tip, flaring into a smoothed equilateral triangle anteriorly). The inner margin is more angular, forming a near right-angle anteriorly, dropping 160–170° to mid-platform, then almost paralleling the furrow to the posterior tip. Basal furrow is usually straight, rarely deflected slightly at the distal end.

Remarks.—There are few intermediate specimens between the two morphotypes. These morphotypes have the same range, occur in approximately equivalent numbers, and both have left and right elements. A complete ontogenetic suite of *N. caudatus* is present in several samples, and differentiation into one or the other morphotype occurs relatively late in ontogeny. The basal cavity geometry is the most stable character of this species and does not vary significantly. All of these factors support erecting a single species for these disparate specimens.

Neognathodus caudatus differs from *N.* sp. A by the lack of pronounced adcarinal troughs and by expansion of the proximal platform. It differs from *N.* sp. B by those same characters, in addition to the latter's more symmetrical platform and basal cavity, and carina that does not extend to the posterior end. *Neognathodus caudatus* is another excellent taxon for marking the Atokan/Desmoinesian boundary (Lambert and Heckel, 1991) and, like *Idiognathodus obliquus* and *I. amplificus*, has been recovered from a wide spectrum of lithofacies. *Neo-*

gnathodus caudatus is relatively abundant, and ranges from the upper Kalo through the lower Floris Formations.

Etymology.—Latin *cauda,* meaning tail, appendage; in reference to the distal carina, where incipient nodes make it resemble *Gnathodus.*

Material.—227 specimens; approximately equal right and left representation.

Holotype.—SUI 61543, Plate 1, no. 1-3.

Neognathodus sp. A
Pl. 1, Figs. 24–27

Remarks.—The Pa element of *Neognathodus* sp. A is characterized by an asymmetrical platform with pronounced adcarinal troughs separating the carina from more highly elevated platform margins. The margins may lose that elevation posteriorly, with the posterior terminus formed by the carina meeting the inner margin, or sometimes both margins. The basal cavity is extremely asymmetrical, with a small, straight inner margin. The inner margin expands increasingly in later forms until it almost matches that of the outer margin. This species probably evolved from *N. atokaensis,* but its precise relationship to *N.* sp. B cannot be determined with the present material. It differs from *N. caudatus* by possessing adcarinal troughs, higher platform margins, and a less flared basal cavity.

Neognathodus sp. B
Pl. 1, Figs. 18–20

Remarks.—Pa elements of *Neognathodus* sp. B are characterized by a symmetrical or nearly symmetrical upper surface and by a symmetrical basal cavity. Adcarinal troughs are pronounced in juvenile specimens, but adults tend to lose them. The margins maintain their elevation above the carina only in the proximal platform, losing distal elevation more rapidly than *N.* sp. A. The margins enclose the carina posteriorly. *Neognathodus* sp. B has been confused with several Morrowan species of *Neognathodus,* as well as with *N. bothrops* Merrill. It differs from both *N. caudatus* and *N.* sp. A by having an enclosed carina and a symmetrical lower and upper outline.

REFERENCES CITED

Gleim, D. T., 1955, Stratigraphy and paleontology of the lower Pennsylvanian rocks in southeastern Iowa: University of Iowa unpublished Ph.D. dissertation, 212 p.

Grayson, R. C., Jr., 1990, Canyon Creek: a significant exposure of a predominantly mudrock succession recording essentially continuous deposition from the Late Devonian through the Middle Pennsylvanian, *in* Ritter, S. M. (ed.), Early to Middle Paleozoic conodont biostratigraphy of the Arbuckle Mountains, southern Oklahoma: Oklahoma Geological Survey Guidebook 27, p. 85–105.

Gregory, J. L., 1982, The stratigraphy and depositional environments of the Kalo Formation (Pennsylvanian) of south-central Iowa: University of Iowa unpublished M.S. thesis, 218 p.

Grubbs, R. K., 1984, Conodont platform elements from the Wapanucka and Atoka Formations (Morrowan–Atokan) of the Mill Creek Syncline, central Arbuckle Mountains, Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 65–79.

Kastler, N. B., 1958, Conodonts from Middle Pennsylvanian of southeastern Iowa: University of Iowa unpublished M.S. thesis, 66 p.

Keyes, C. R., 1893, The geological formations of Iowa: Iowa Geological Survey Annual Report, 1892, v. 1, p. 13–144.

Koike, T., 1967, A Carboniferous succession of conodont faunas from the Atetsu Limestone in southwest Japan: Tokyo Kyoiku Daigaku Science Reports, Section C, Geology, Mineralogy, and Geography, v. 9, p. 279–318.

Lambert, L. L.; and Heckel, P. H., 1991, The Atokan/Desmoinesian boundary in North America: preliminary considerations for selecting a boundary horizon, *in* Brenckle, P. L.; and Manger, W. L. (eds.), Intercontinental division and correlation of the Carboniferous System: Courier Forschungsinstitut Senckenberg 130, p. 307–318.

Landis, E. R.; and Van Eck, O. J., 1965, Coal resources of Iowa: Iowa Geological Survey Technical Paper 4, 141 p.

Lane, H. R., 1968, Symmetry in conodont element-pairs: Journal of Paleontology, v. 42, p. 1258–1263.

Mamay, S. H.; and Yochelson, E. L., 1962, Occurrence and significance of marine animal remains in American coal balls: U.S. Geological Survey Professional Paper 354-I, p. 193–224.

Manger, W. L.; and Sutherland, P. K., 1984, Preliminary conodont biostratigraphy of the Morrowan–Atokan boundary (Pennsylvanian), eastern Llano uplift, central Texas, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 115–121.

Moore, R. C.; and Thompson, M. L., 1949, Main divisions of Pennsylvanian Period and System: American Association of Petroleum Geologists Bulletin, v. 33, p. 275–302.

Rabe, E. H., 1977, Contributions to the stratigraphy of the East-Andean area of Columbia [in German]: Giessener Geologische Schriften 11, 223 p.

Ravn, R. L., 1979, An introduction to the stratigraphic palynology of the Cherokee Group (Pennsylvanian) coals of Iowa: Iowa Geological Survey Technical Paper 6, 117 p.

______1986, Palynostratigraphy of the Lower and Middle Pennsylvanian coals of Iowa: Iowa Geological Survey Technical Paper 7, 245 p.

Ravn, R. L.; Swade, J. W.; Howes, M. R.; Gregory, J. L.; Anderson, R. R.; and Van Dorpe, P. E., 1984, Stratigraphy of the Cherokee Group and revision of Pennsylvanian stratigraphic nomenclature in Iowa: Iowa Geological Survey Technical Information Series 12, 76 p.

Searight, W. V.; Howe, W. B.; Moore, R. C.; Jewett, J. M.; Condra, G. E.; Oakes, M. C.; and Branson, C. C., 1953, Classification of Desmoinesian (Pennsylvanian) of northern Mid-Continent: American Association of Petroleum Geologists Bulletin, v. 37, p. 2747–2749.

Thompson, M. L., 1934, The fusulinids of the Des Moines series of Iowa: Iowa University Studies in Natural History, v. 16, p. 272–332.

Zeller, D. E. (ed.), 1968, The stratigraphic succession in Kansas: State Geological Survey of Kansas Bulletin 189, 81 p.

APPENDIX: Localities

This appendix lists only localities and cores from which specimens are figured.

Locality 1.—Abandoned strip mine, SW¼NW¼ sec. 29, T. 74 N., R. 16 W., Mahaska County, Iowa. Most of the Kalo Formation is exposed by this strip pit, and both the Blackoak and Cliffland coals were mined. Although a complete suite of Kalo samples has been taken, specimens are only illustrated from the marine interval above the Cliffland coal. Those samples are measured from the top of the Cliffland coal: A, 0–7.5″ (basal black shale); B, 7.5–12.5″ (uppermost black shale); C, 24.5–32.5″ (lower limestone); D, 48–56″ (upper limestone); E, 60–66″ (basal dark shale).

Cores.—Detailed core logs with unit depths are available from the Iowa Department of Natural Resources, Geological Survey Bureau, Iowa City, Iowa.

CP 6. NW¼SW¼NW¼ sec. 8, T. 70 N., R. 12 W., Davis County, Iowa.

CP 22. SE¼SW¼SE¼ sec. 36, T. 70 N., R. 19 W., Appanoose County, Iowa.

CP 24. NW¼NE¼SE¼ sec. 1, T. 67 N., R. 14 W., Davis County, Iowa.

CP 37. NE¼SE¼NE¼ sec. 2, T. 72 N., R. 26 W., Clarke County, Iowa.

CP 47. SE¼NE¼NE¼ sec. 4, T. 79 N., R. 25 W., Polk County, Iowa.

CP 54. SW¼NE¼NW¼ sec. 29, T. 72 N., R. 18 W., Monroe County, Iowa.

Plates

PLATE 1

Magnifications all ×32. All figures are of Pa elements. All views upper unless noted otherwise. Specimens reposited at the Department of Geology, University of Iowa (SUI).

Figures 1–16.—*Neognathodus caudatus* n. sp. **1–3**, upper, oblique, and lower views of SUI holotype 61543; **4**, SUI 61544; **9**, SUI 61545; **11**, SUI 61546; **16**, SUI 61547, Locality 1-A. **5**, SUI 61548; **6**, SUI 61549, CP 6-unit 15. **7**, SUI 61550, Locality 1-E. **8**, SUI 61532; **10**, SUI 61551; **13**, SUI 61552; **14**, SUI 61533, Locality 1-B. **12**, SUI 61553; **15**, SUI 61554, Locality 1-D.

Figures 17,21–23.—*Idiognathodus* sp. B. **17,22**, inner side and upper view, SUI 61561; **21**, SUI 61562, Locality 1-C. **23**, SUI 61534, Locality 1-B.

Figures 18–20.—*Neognathodus* sp. B. **18**, SUI 61555; **19**, SUI 61537, CP 37-unit 65. **20**, SUI 61556, CP 37-unit 63.

Figures 24–27.—*Neognathodus* sp. A. **24**, SUI 61557, CP 24-unit 37. **25**, SUI 61558; **27**, SUI 61559, CP 22-unit 56. **26**, SUI 61560, CP 54-unit 21.

Figures 28–30.—*Idiognathodus* sp. A. **28**, SUI 61589; **29**, SUI 61590, CP 47-unit 43. **30**, SUI 61591, CP 22-unit 56.

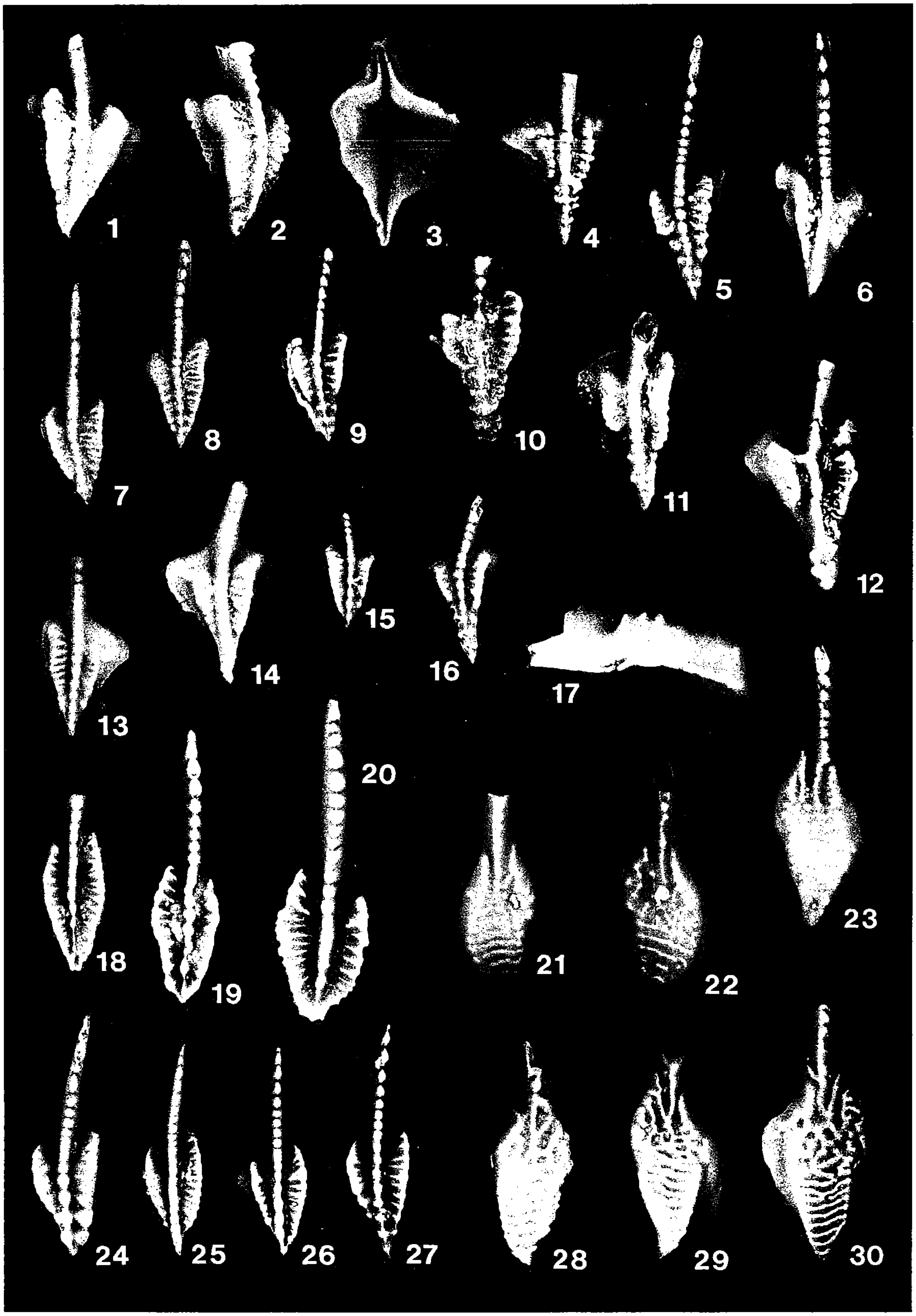

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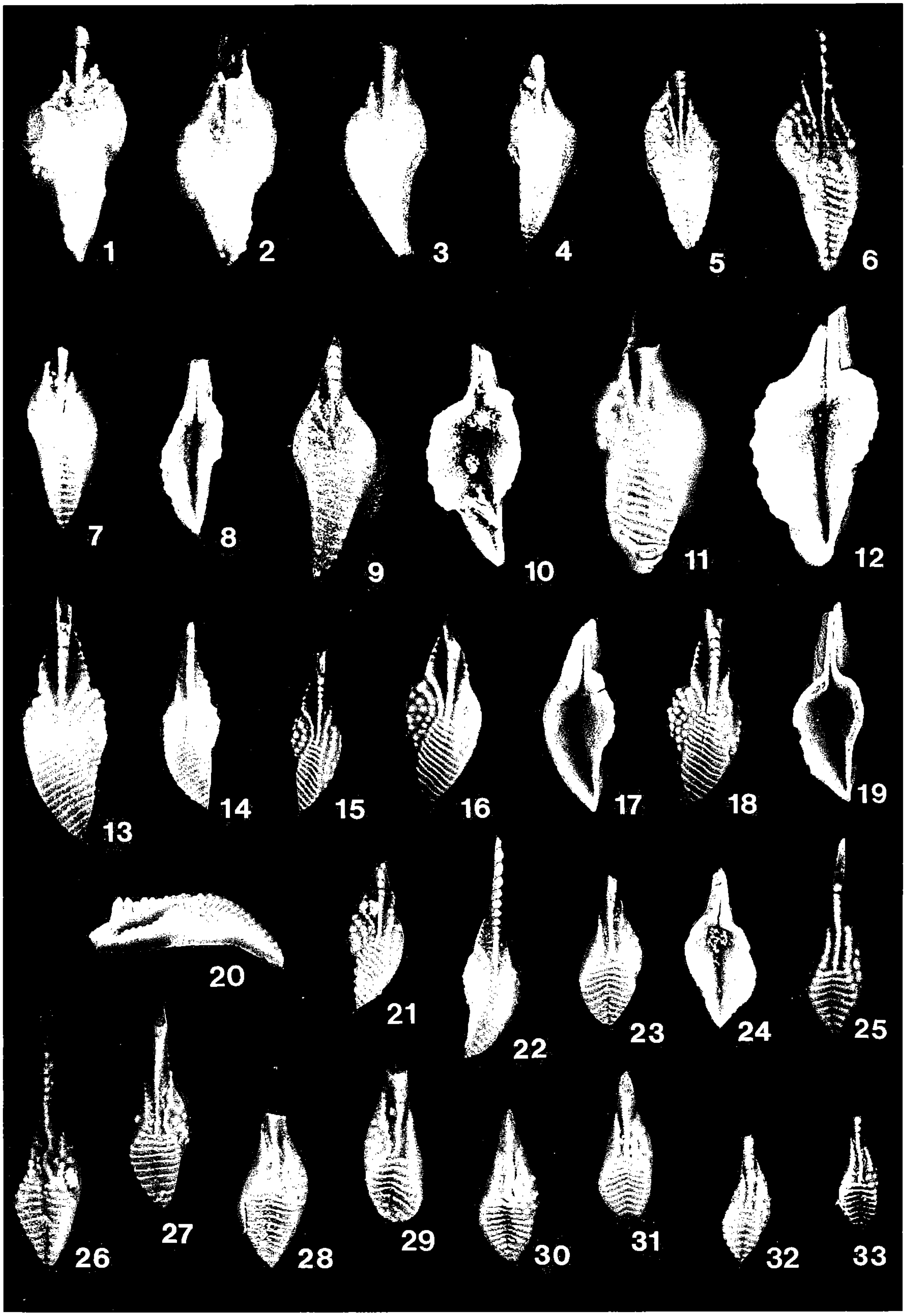
PLATE 2

Magnifications all ×32. All figures are of Pa elements. All views upper unless noted otherwise.

Figures 1–12.—*Idiognathodus amplificus* n. sp. **1**, SUI holotype 61563; **2**, SUI 61564, Locality 1-A. **3**, SUI 61565, Locality 1-E. **4**, SUI 61566; **5**, SUI 61567; **7,8**, upper and lower view, SUI 61568; **9,10**, upper and lower view, SUI 61569; **11,12**, upper and lower view, SUI 61570, Locality 1-B. **6**, SUI 61571, Locality 1-D.

Figures 13–22.—*Idiognathodus obliquus* Kossenko and Kozitskaya. **13**, SUI 61572, Locality 1-B. **14**, SUI 61573; **21**, SUI 61574; **22**, SUI 61575, Locality 1-A. **15**, SUI 61576, CP 6-unit 15. **16,17**, upper and lower view, SUI 61577; **18–20**, upper, lower, inner side views, SUI 61535, Locality 1-D.

Figures 23–33.—*Idiognathodus gibbus* n. sp. **23,24**, upper and lower view, SUI 61578; **25**, SUI 61579; **26**, upper view, SUI holotype 61580; **28**, SUI 61581; **29**, SUI 61582; **30**, SUI 61583; **31**, SUI 61584; **32**, SUI 61585; **33**, SUI 61586, CP 6-unit 25. **27**, SUI 61587, CP 37-unit 65.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Oklahoma Geological Survey Circular 94, 1992

Middle Pennsylvanian Fusulinid Biostratigraphy in South-Central New Mexico and South-Central Oklahoma

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ABSTRACT.—The Derryan Series, at its type section in south-central New Mexico, is represented by a 38-m fusulinid-bearing, carbonate-dominated sequence. Lithologies range from carbonate mudstone to packstone with interbedded shale. Fusulinids are abundant and well preserved. The Zones of *Millerella, Eoschubertella, Profusulinella* and *Fusulinella* are recognized. The Zone of *Beedeina* (= *Fusulina*) and the first stratigraphic occurrence of *Wedekindellina* are found above Derryan strata.

The Atokan Series of south-central and southeastern Oklahoma is represented by dominantly clastic lithologies reaching >2,000 m thick (Atoka Formation). Atoka Formation units yielding fusulinids are thin, poorly exposed, and discontinuous. In addition, fusulinid specimens tend to be abraded or broken. Atoka Formation fusulinid faunas, however, include several species found in Derryan strata.

Derryan fusulinids comparable to Atoka Formation species include: *Profusulinella* cf. *fittsi* (Thompson), *Fusulinella acuminata* Thompson, *Fusulinella devexa* Thompson, and *Fusulinella famula* Thompson. Desmoinesian forms similar to species from southeastern Oklahoma include *Beedeina insolita* (Thompson) and *Beedeina* aff. *hayensis* Ross and Sabins. Fusulinid zonation has been used for biostratigraphic correlation between these markedly different, but nearly time equivalent, lower Middle Pennsylvanian sequences.

INTRODUCTION

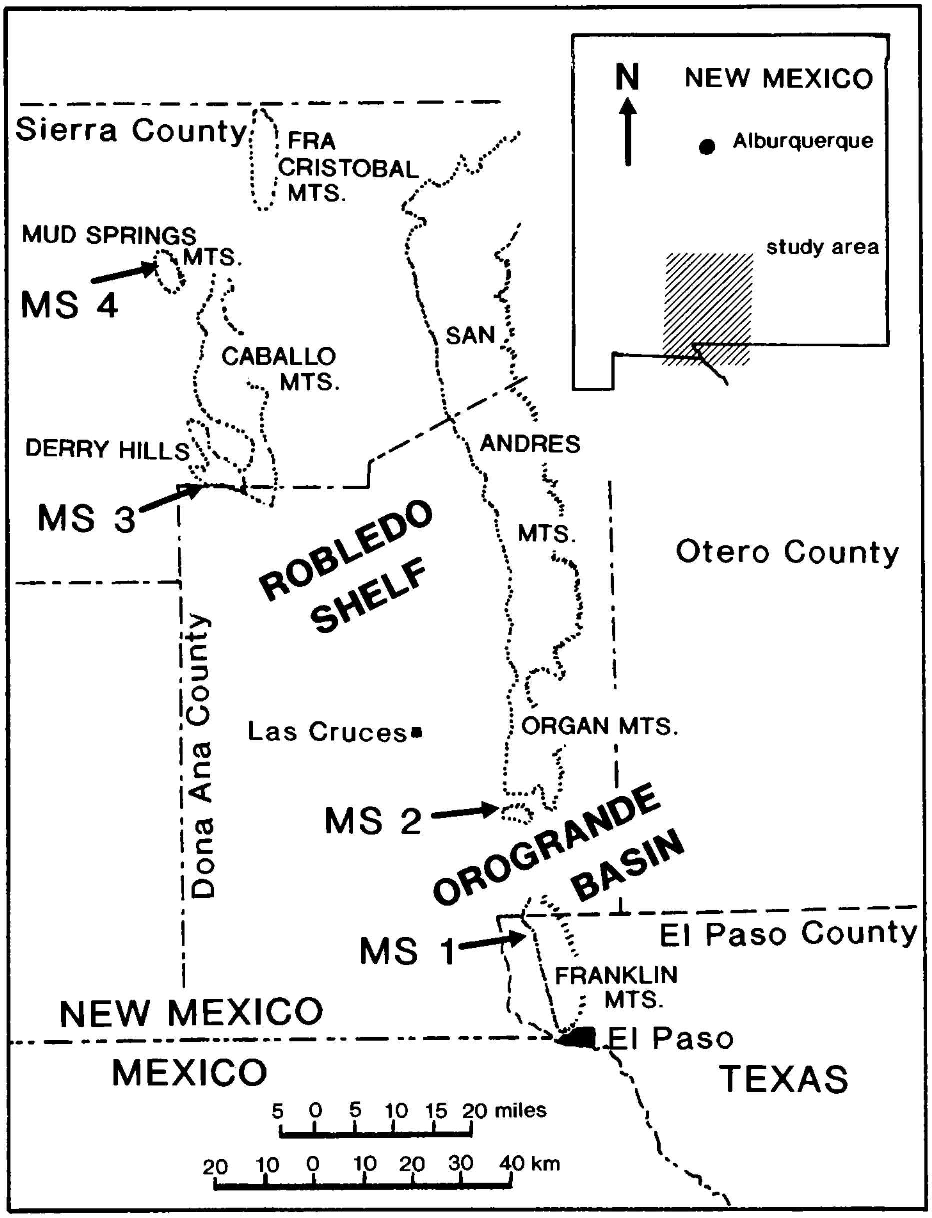
The Atokan Series (lower Middle Pennsylvanian System), as used in the Midcontinent, and the Derryan Series, as described in south-central New Mexico, include approximately the same stratigraphic interval. Thompson (1942) proposed the Derryan Series for all Pennsylvanian rocks below the base of the Desmoinesian Series (upper Middle Pennsylvanian System) in the central to the extreme south-central areas of New Mexico (Text-fig. 1). As originally defined, the Derryan Series significantly overlaps the basal Pennsylvanian Morrowan Series (Clopine, 1988). Thompson (1948) later amended the Derryan Series to include all rocks between the Morrowan and Desmoinesian Series, but the type section was not redefined.

Thompson (1942,1948) rejected all previous classification systems for Pennsylvanian strata in central and southern New Mexico due to the lack of adequately described boundaries between most lithologic units. He proposed an entirely new system for lithostratigraphic classification based on measured sections mainly from the Mud Springs Mountains and the Derry Hills of Sierra County (Text-fig. 2). Unfortunately, the type sections for many of Thompson's "formations" are also not adequately described, and their upper and lower boundaries are not well defined. Many units are difficult or impossible to map, and subdivisions are based, at least in part, on fusulinid occurrences. In addition, although informally referred to in the literature (e.g., Thompson, 1942,1948; Lane and others, 1972; King, 1973; Groves, 1986), fusulinid occurrences at the type Derryan section have not been adequately documented.

Spivey and Roberts (1946) proposed the name Atokan to include a post-Morrowan and pre-Desmoinesian series in south-central and southeastern Oklahoma represented by fusulinids more advanced than *Millerella*, but below the appearances of *Beedeina* (= *Fusulina*) and *Wedekindellina*. They rejected the term Derryan because at its type locality it "is only about 40 m (130 ft) thick, whereas the Atoka beds reach 2,133 m (7,000 ft) in thickness" (Spivey and Roberts, 1946, p. 186). The Atokan Series has gained widespread acceptance in the literature despite the fact that no type section was proposed, its base is unconformable in its type area, and later workers (Douglass and Nestell, 1984) have identified Desmoinesian fusulinids within "Atokan" strata.

This paper documents fusulinid occurrences at the Derryan type section, discusses fusulinid biostratigraphic correlations in southern New Mexico and westernmost Texas, and compares those faunas to previously described fusulinids from the Atoka Formation in its type area of south-central Oklahoma.

Type Derryan area fusulinid occurrences presented here are based on remeasurement and recollection of Thompson's (1942) Derry and Whiskey Canyon sections (MS 3 and 4 of this author) and two additional sections to the southeast. All figured specimens were recovered at the type Derryan section (Text-fig. 1) and are deposited in the Oklahoma Museum of Natural History, Invertebrate Paleontology Repository (OU). Samples were collected in each lithologic unit where fusulinaceans were observed or at a maximum of 1.5-m intervals where lithology did not change or fusulinaceans were not observed. Complete measured section data are included in Clopine (1990).



Text-figure 1. Locality map, with the approximate position of the Orogrande basin and Robledo shelf. Measured Section 1—Vinton Canyon, northern Franklin Mountains; MS 2—Bishop Cap, southern end of Organ Mountains; MS 3—Type Derryan, Derry Hills; MS 4—Whiskey Canyon, northern Mud Springs Mountains (modified from Verville and others, 1986).

	SERIES	FUSULINID BIOZONE	Gordon 1907		Needham 1937	Nelson 1940		Thompson 1942, 1948			Kelley & Silver 1952	This Paper
PENNSYLVANIAN SYSTEM	Virgilian	Zone of *Triticites*	Magdalena Group	Madera Limestone	Magdalena Formation	Magdalena Formation	unnamed upper member		Fresnal Group			Not Studied
										Bruton Fm.		
									Keller Group	Moya Formation		
										Del Cuerto Formation		
	Missourian								Hansonburg Group	Story Fm.	Bar B Formation	
										Burrego Fm.		
									Veredas Group	Council Spring Limestone		
							Bishop Cap Member			Adobe Formation		
										Coane Fm.		
	Desmoinesian	Zone of *Fusulina*							Bolander Group		Nakaye Formation	Lower Desmoinesian Series
				Sandia Formation					Armendaris Group	Garcia Fm.		
										Whiskey Canyon		
										Elephant Butte Fm.		
							Berino Member			Warmington		
	Atokan	Zone of *Fusulinella*						Derry Series	Mud Springs Group	Cuchillo Negro	Red House Formation	Atokan Series
										Fra Cristobal Formation		
		Zone of *Profusulinella*							Green Canyon Group	Apodaca Formation		
		Eoschubertella					La Tuna Member			Arrey Formation		
	Morrowan	Zone of *Millerella*										Morrowan Series

Text-figure 2. Lithostratigraphic subdivisions used by several authors in central and south-central New Mexico and westernmost Texas with approximate chronostratigraphic and biostratigraphic correlations.

Acknowledgments

This study was supported by National Science Foundation Grant EAR-8517591 to Patrick K. Sutherland, and by grants from Amoco Research Laboratory in Tulsa, Oklahoma, the University of Oklahoma School of Geology and Geophysics, and the New Mexico Bureau of Mines and Mineral Resources to Clopine. Interpretations benefited from observations made in the field by Patrick K. Sutherland, Walter L. Manger, and George J. Verville. The paper also benefited from the critical review of an earlier version by Patrick K. Sutherland, John R. Groves, George J. Verville, and George A. Sanderson.

TYPE DERRYAN AREA

Stratigraphic Setting and Fusulinid Occurrence

The Robledo shelf and Orogrande basin areas of south-central New Mexico and westernmost Texas were the site of broad shallow carbonate shelf deposition in Early and Middle Pennsylvanian time (Text-fig. 1). Subtidal deposits are dominant, but interbedded intertidal and supratidal facies also occur. The thickest sequences occur near the axis of the Orogrande basin. Basal northwestward onlap of Morrowan strata and lower subsidence rates progressively reduce the thickness of Lower and Middle Pennsylvanian strata onto the Robledo Shelf.

The type Derryan section is located on the central Robledo Shelf. Upper Morrowan and Atokan strata occur within the Derryan type section (Thompson, 1942, p. 32–33, section #19). Limestones immediately above type Derryan strata include an abundant, well-preserved, Early Desmoinesian fusulinid fauna.

Sutherland and Manger (1984), on close examination of a new road cut exposure immediately adjacent to the original type section, identified a well-preserved late Morrowan brachiopod fauna within basal Derryan beds (lower nodular portion of Thompson's [1942] unit 1). At the same horizon, Groves (1986) identified the primitive fusulinaceans *Millerella extensa* Marshall and *Eostaffella* spp., both long ranging, to the exclusion of more-advanced forms (Pl. 1). Manger and others (1987) placed the Morrowan/Atokan boundary at an unconformity within a thin shale interval <1 m above the base of the Pennsylvanian System. Morrowan strata thicken dramatically southeast of the type Derryan section, reaching 75 m in the northern Franklin Mountains (Clopine, 1990, MS 1).

Thompson (1942, p. 33) reported *Profusulinella* in the "Arrey Formation" at the type Derryan section 5 m above the base of the Pennsylvanian System (unit 3 of Thompson, 1942; unit 3-10 of this author). Thompson (1948, p. 82) altered his earlier interpretation stating that further study seemed to indicate that specimens from his unit 3, formerly referred to as *Profusulinella*, may be a primitive form of *Eoschubertella*. Thompson's latter view, fully supported by this author, has apparently not been noted by other workers.

Lane and others (1972, p. 553) reported *Profusulinella* 4.5 m above the base of the type Derryan section (same as Thompson's [1942] lowest "*Profusulinella*" horizon). However, these specimens have been reexamined by G. J. Verville (personal communication, 1989) who stated that they should be assigned to *Eoschubertella*. King (1973, p. 2) reported *Profusulinella* in bed 2 of Thompson (1942; unit 3-7 of this author). In addition, King (1973, p. 2) suggested from "unidentifiable fragments" that the lowest *Profusulinella* occurrence may be in Thompson's bed 1. King's material is no longer available for study, however, and he now believes that those specimens were probably *Eoschubertella* (W. E. King, personal communication, 1989).

Eoschubertella mexicana Thompson and *Pseudoendothyra* sp. are the most advanced fusulinids found in the "Arrey Formation" at the type Derryan section (Pl. 1). *Eoschubertella*? occurs 0.25 m above the Morrowan/Atokan unconformity (1.2 m above the base of the Pennsylvanian System). Unequivocal *Eoschubertella* first occurs 1.2 m higher in the section. *Eoschubertella mexicana* becomes more common and occurs with *Pseudoendothyra* slightly higher in the section (Pl. 1, Figs. 1–7). The interval including *Eoschubertella* and/or *Pseudoendothyra* below the appearance of *Profusulinella* is here referred to as the Zone of *Eoschubertella* and included in the lower part of the Atokan Series (see Clopine, 1990, p. 17–22, for further discussion of the Zone of *Eoschubertella*). Unfortunately, the rarity of occurrence of *Eoschubertella* and other early Atokan indicators within the lower part of their range, limits their biostratigraphic usefulness for regional correlation.

The only documented occurrence of *Profusulinella* at the type Derryan section is in a previously unreported horizon 2.7 m above the base of the "Apodaca Formation" (13.65 m above the lowest Atokan strata) where rare *Profusulinella* cf. *fittsi* have been recovered (Pl. 2, Figs. 8,9). This horizon is tentatively correlated with the lower portion of the Zone of *Profusulinella* based on the rare occurrence of similar specimens associated with common *Profusulinella copiosa* Thompson at MS 1 and Bishop Cap (Clopine, 1990, MS 2).

Fusulinella is very well represented in the type Derryan section. There, two major subdivisions within the Zone of *Fusulinella* have been recognized. These are informally termed the *Fusulinella acuminata* Thompson lineage zone and the *Fusulinella devexa* Thompson interval zone (ascending order). *Fusulinella acuminata* appears in the upper portion of the "Apodaca Formation" and ranges into the upper portion of the "Fra Cristobal Formation." Thompson (1948) indicated the occurrence of *Profusulinella apodacensis* Thompson in the upper "Apodaca Formation." In this study, however, the "Apodaca Formation" yielded only members of the genus *Fusulinella*. *Fusulinella fugax* Thompson and *Fusulinella proxima* Thompson (Pl. 2) occur with *F. acuminata* in the lower portion of the latter's range at the type Derryan section (MS 3).

Fusulinella acuminata (Pl. 3, Figs. 7–9) appears to be the direct ancestor of *Fusulinella devexa* (Pl. 3, Figs. 1–6). These forms occur together at the base of the *Fusulinella devexa* interval zone. The *Fusulinella devexa* interval zone includes the upper portion of the "Fra Cristobal Formation" and the "Cuchillo Negro Formation" at the type Derryan section. The upper boundary of the *F. devexa* interval zone corresponds to the top of the type Derryan Series, which occurs at the base of a 0.5-m-thick dolomite immediately below the first occurrence of *Beedeina insolita* Thompson.

At MS 4, the *Fusulinella devexa* interval zone includes the upper portion of the "Fra Cristobal Formation," the "Cuchillo Negro Formation," and the lower "Elephant Butte Formation" (Thompson, 1942). It is interesting to note that there is little lithologic basis for the subdivision of these "formations." In addition, they include somewhat different fusulinid faunas from the same "formations" at the type Derryan section. As a result, none of them have significant lithostratigraphic or biostratigraphic use.

Thompson (1948) reported rare *Beedeina insolita* associated with common *Fusulinella devexa* at Whiskey Canyon (*Beedeina insolita* type locality). This reported occurrence is 2 m below the lowest *Beedeina insolita* recovered at MS 4 in this study. This slightly younger *Beedeina insolita* occurrence, however, is still within the upper range of *Fusulinella devexa*. The first occurrence of *Beedeina insolita* is above the range of *Fusulinella devexa* at MS 3, suggesting a minor unconformity at this locality. Alternatively, the barren dolomite bed immediately above type Derryan strata and immediately below the lowest *Beedeina insolita* could represent the interval of overlap (Pl. 4).

In the study area, several species of *Beedeina* occur within the interval between the base of the Zone of *Beedeina* and the first occurrence of *Wedekindellina* (Pl. 5). This interval is assigned to the lower Desmoinesian Series. Its lower portion is characterized by *Beedeina insolita*, a primitive member of the genus. The somewhat more advanced *Beedeina* aff. *hayensis* and *Beedeina* aff. *joyitaensis* occur slightly higher stratigraphically. Each of these species is relatively primitive compared to congeneric forms found above the first occurrence of *Wedekindellina*. This interval is 16.9 m thick at MS 3 and gradually increases to 51.6 m to the southeast at MS 1.

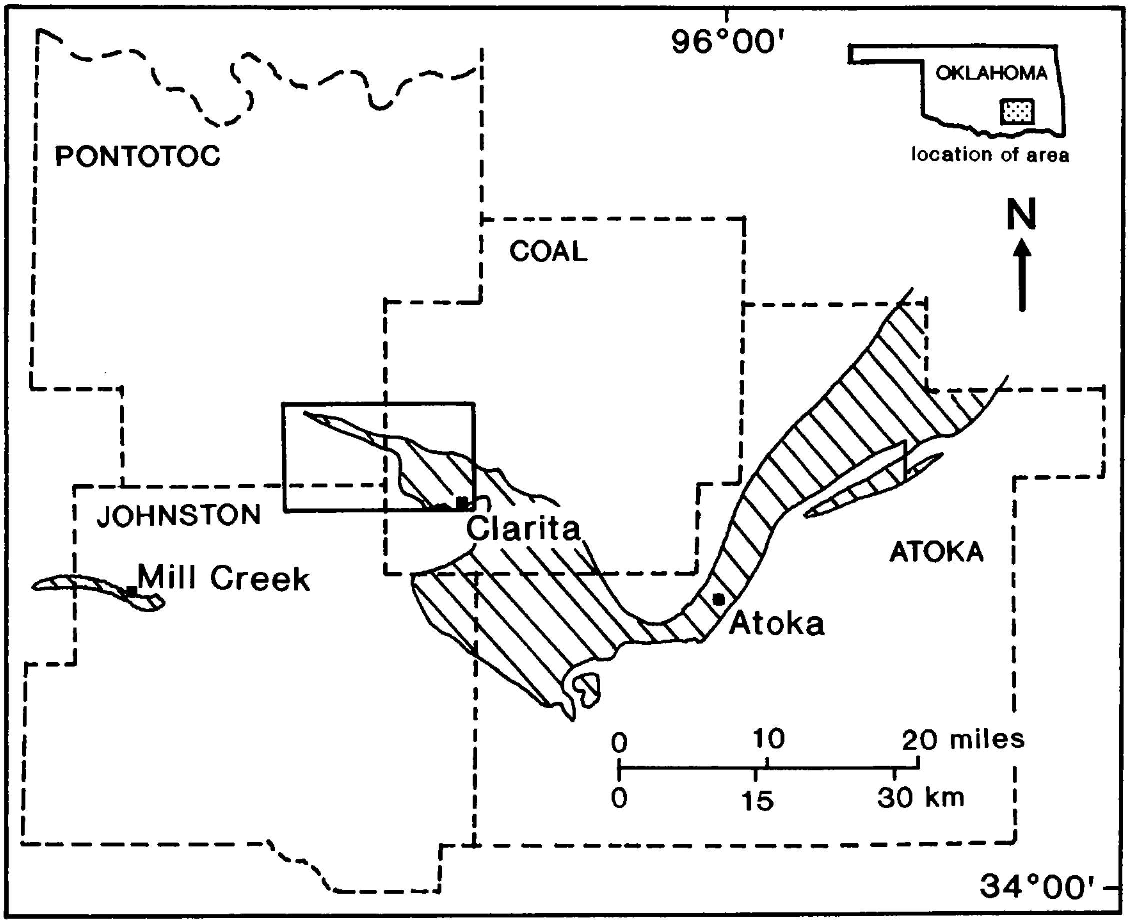
CORRELATIONS TO TYPE ATOKAN AREA

Lower and Middle Pennsylvanian fusulinid foraminifers from south-central New Mexico and westernmost Texas are comparable to faunas of similar age from the Midcontinent. Douglass and Nestell (1984) documented fusulinid occurrences in the Atoka Formation (Lower–Middle Pennsylvanian) in the Atokan type area of south-central Oklahoma (Text-fig. 3). That fauna is generally poorly preserved, specimens tend to be abraded or broken, and limestones yielding fusulinids are thin, discontinuous, and poorly exposed. Nevertheless, several important correlations are possible.

Douglass and Nestell (1984) found the zones of *Profusulinella*, *Fusulinella*, and *Beedeina* represented in the Atoka Formation. Groves and Grayson (1984) recovered *Eoschubertella* in the upper limestone of the Wapanucka Formation in the frontal Ouachita Mountains. The unconformity at the base of the Atoka Formation on the northeastern flank of the Arbuckle Mountains, and subsequent poorly or nonfossiliferous intervals, make correlations to the type Atokan rather tenuous. However, the basal Atokan Zone of *Eoschubertella*, recognized at the type Derryan section and across the study area, appears correlative with the upper Wapanucka Formation in the frontal Ouachita Mountains. The Zone of *Eoschubertella* is not recognized on the northeastern flank of the Arbuckle Mountains due to the significant basal Atokan unconformity in this area (Zachry and Sutherland, 1984, fig. 3).

The Zone of *Profusulinella* is the oldest fusulinid zone recognized in the Atoka Formation. It is poorly represented, however, and only two species have been recovered. The most primitive form is *Profusulinella fittsi* (Thompson). The type locality of this species is its only previously reported occurrence. However, locality information given by Thompson (1935) is incorrect, and the precise stratigraphic position of the type specimens is not known (Douglass and Nestell, 1984).

Profusulinella millcreekensis Douglass and Nestell in the Arbuckle Mountains area is the only other profusulinellid reported in the Atoka Formation. *Profusulinella millcreekensis* has not been recovered in southern New Mexico or westernmost Texas. However, it appears similar to *Profusulinella* sp., occurring at MS 1. This form is a morphologic intermediate between the primitive ovoid species *Profusulinella copiosa* and more advanced forms such as *Profusulinella munda* Thompson, *Profusulinella decora* Thompson, and *Profusulinella apodacensis*. Groves (1991) recognized an even more advanced form, *Profusulinella regia* Thompson above *P. apodacensis* in the upper Marble Falls and basal Smithwick Formations in central Texas. *Profusulinella regia*, along with *P. kentuckyensis* Thompson and Riggs from Butler County, Kentucky, and



Text-figure 3. Outcrop of Atoka Formation in southwestern Arkoma basin, Oklahoma (= northeastern flank of Arbuckle Mountains), the box marking the type area of the Atokan Series as used by Sutherland and Manger (1984). Atokan exposure in the Mill Creek syncline in Johnston County is also marked (modified from Douglass and Nestell, 1984).

P. whitetailensis Ross and Sabins from the Horquilla Limestone in southeastern Arizona are among the youngest and morphologically most advanced profusulinellids (Groves, 1991). Those forms are more advanced than any *Profusulinella* recovered from the Derryan type area or the Atokan type area.

The Zone of *Fusulinella* is well represented in southern New Mexico and west Texas and in the Atoka Formation of Oklahoma. The informal subdivision of the Zone of *Fusulinella* in southern New Mexico into the older *Fusulinella acuminata* lineage zone and the younger *Fusulinella devexa* interval zone is not possible in the Atoka Formation, where no well-preserved evolutionary sequence can be identified. However, the highly advanced *Fusulinella* aff. *leyi* found near the top of the Atoka Formation in the Mill Creek syncline area in Johnston County, Oklahoma (Text-fig. 3), is very comparable to *Fusulinella famula* Thompson. *Fusulinella famula* is associated with *Beedeina insolita* and highly advanced *Fusulinella devexa* in basal Desmoinesian strata in southern New Mexico (Pl. 4). *Fusulinella* aff. *leyi* without associated Desmoinesian forms suggests latest Atokan age for highest Atoka Formation beds in the Mill Creek syncline.

The Zone of *Beedeina* is well developed below the first occurrence of *Wedekindellina* in southern New Mexico and westernmost Texas. This is in contrast to south-central Oklahoma, where the lowest stratigraphically documented occurrences of *Beedeina* and *Wedekindellina* in the Mill Creek syncline fall at the same horizon (Douglass and Nestell, 1984). This horizon is above an unconformity at the Atokan/Desmoinesian boundary and immediately above a basal Desmoinesian conglomerate. The unconformity apparently represents the interval noted in southern New Mexico and westernmost Texas between the first occurrence of *Beedeina* and the first occurrence of *Wedekindellina*.

Beedeina grileyi Douglass and Nestell occurs in the absence of *Wedekindellina* just below the "Griley" Limestone on the northeast flank of the Arbuckle Mountains in Coal County, Oklahoma (Text-fig. 3). This informal lithologic unit has been included in the upper Atokan Formation (Sutherland and others, 1982). That interpretation suggests that the highest part of the Atoka Formation, at least locally, is early Desmonesian in age and correlates to the interval in southern New Mexico that lies within the Zone of *Beedeina* below the first occurrence of *Wedekindellina*. P. K. Sutherland (personal communication, 1990) pointed out that the "Griley" Limestone and its associated shales and sandstones apparently represent a down-dropped block of strata preserved between the Jack Hills Fault and the Franks Fault (see Sutherland and others, 1982, fig. 7), and thus cannot be related stratigraphically either to the main part of the Atoka Formation below or to the Desmoinesian Hartshorne and McAlester Formations above. The significance of this apparent Desmoinesian age of the upper part of the Atoka Formation in this area cannot be evaluated elsewhere in the region because of the general paucity of diagnostic fossils.

Thompson (1934) described *Beedeina* and *Wedekindellina* near the base of the Pennsylvanian System in the type Desmoinesian area of Iowa. There is no specific type section for the Desmoinesian Series, however, and the Desmoinesian type area has obvious limitations as a reference standard (Sanderson and West, 1981). Thompson (1934) reported the occurrence of *Fusulinella iowensis* from ~6 m above the basal Pennsylvanian unconformity and 27 m below the first occurrence of *Wedekindellina*. Shaver (1984) included in the Desmoinesian Series this *F. iowensis*-bearing interval that lies below the first occurrence of *Beedeina* and *Wedekindellina*. Sanderson and West (1981) and most other authors consider the Desmoinesian to be essentially equivalent to the Zone of *Beedeina*, and they included the *Fusulinella iowensis* interval in the uppermost Atokan. The latter position is supported by Ravn (1986, p. 10), who regarded the lowest portion of the Des Moines Supergroup to be Atokan in age based on palynomorphs from coals in the Kilbourn and lower Kalo Formations. Thompson (1948) found *Fusulinella* cf. *iowensis* associated with *Beedeina* in southern New Mexico, and it is clear that *Fusulinella* ranges across the Atokan/Desmoinesian boundary. For these reasons, the first occurrence of *Beedeina* remains the best basal Desmoinesian indicator, and the base of the Pennsylvanian System in the Desmoinesian type area should not be regarded as basal Desmoinesian in age (see Lambert, this volume, for further discussion).

REFERENCES CITED

Clopine, W. W., 1988, Fusulinid zonation and stratigraphic analysis of the Derryan Series (Middle Pennsylvanian) in southern New Mexico–West Texas [abstract]: Geological Society of America Abstracts with Programs, v. 20, no. 7, p. 122.

______1990, Middle Pennsylvanian fusulinid biostratigraphy of the type Derryan area in southern New Mexico and westernmost Texas: University of Oklahoma unpublished Ph.D. dissertation, 271 p.

Douglass, R. C.; and Nestell, M. K., 1984, Fusulinids of the Atoka Formation, Lower–Middle Pennsylvanian, south-central Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 19–39.

Gordon, C. H., 1907, Mississippian (Lower Carboniferous) formations in the Rio Grande Valley: American Journal of Science [series 4], v. 24, no. 174, p. 58–64.

Groves, J. R., 1986, Foraminiferal characterization of the Morrowan–Atokan (lower Middle Pennsylvanian) boundary: Geological Society of America Bulletin, v. 97, p. 346–353.

______1991, Fusulinacean biostratigraphy of the Marble Falls Limestone (Pennsylvanian), western Llano region, central Texas: Journal of Foraminiferal Research, v. 21, no. 1, p. 67–95.

Groves, J. R.; and Grayson, R. C., Jr., 1984, Calcareous foraminifers and conodonts from the Wapanucka Formation (Lower–Middle Pennsylvanian), frontal Ouachita Mountains, southeastern Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 81–89.

Kelley, V. C.; and Silver, C., 1952, Geology of the Caballo Mountains: University of New Mexico Publications in Geology No. 4, 286 p.

King, W. E., 1973, Fusulinids *Millerella* and *Eostaffella* from the Pennsylvanian of New Mexico and Texas: New Mexico Bureau of Mines and Mineral Resources Memoir 26, 34 p.

Lane, H. R.; Sanderson, G. A.; and Verville, G. J., 1972, Uppermost Mississippian–basal Middle Pennsylvanian conodonts and fusulinids from several exposures in the south-central and southwestern United States: 24th International Geological Congress, Montreal, Section 7, p. 549–555.

Manger, W. L.; Sutherland, P. K.; and Clopine, W. W., 1987, Morrowan–Atokan boundary relations, type "Derryan" section, New Mexico [abstract]: Geological Society of America Abstracts with Programs, v. 19, no. 3, p. 173.

Needham, C. E., 1937, Some New Mexico Fusulinidae: New Mexico School of Mines, State Bureau of Mines and Mineral Resources Bulletin 14, 59 p.

Nelson, L. A., 1940, Paleozoic stratigraphy of Franklin Mountains, West Texas: American Association of Petroleum Geologists Bulletin, v. 24, p. 157–172.

Ravn, R. L., 1986, Palynostratigraphy of the Lower and Middle Pennsylvanian coals of Iowa: Iowa Geological Survey Technical Paper No. 7, p. 245.

Sanderson, G. A.; and West, R. R., 1981, Fusulinids from the Desmoinesian type area [abstract]: Geological Society of America Abstracts with Programs, v. 13, no. 7, p. 545.

Shaver, R. H., 1984, Atokan Series concepts with special reference to the Illinois basin and Iowa, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 101–113.

Spivey, R. C.; and Roberts, T. G., 1946, Lower Pennsylvanian terminology in central Texas: American Association of Petroleum Geologists Bulletin, v. 30, p. 181–186.

Sutherland, P. K.; and Manger, W. L., 1984, Morrowan brachiopods in the type "Derryan" Series (Pennsylvanian), New Mexico [abstract]: Geological Society of America Abstracts with Programs, v. 16, no. 2, p. 115.

Sutherland, P. K.; Archinal, B. E.; and Grubbs, R. K., 1982, Morrowan and Atokan (Pennsylvanian) stratigraphy in the Arbuckle Mountains area, Oklahoma, *in* Sutherland, P. K. (ed.), Lower and Middle Pennsylvanian stratigraphy in south-central Oklahoma: Oklahoma Geological Survey Guidebook 20, p. 1–18.

Thompson, M. L., 1934, The fusulinids of the Des Moines Series of Iowa: University of Iowa Studies, Studies in Natural History, New Series No. 284, v. 16, p. 277–332.

______1935, Fusulinids from the Lower Pennsylvanian Atoka and Boggy Formations of Oklahoma: Journal of Paleontology, v. 9, p. 291–306.

______1942, Pennsylvanian System in New Mexico: New Mexico Bureau of Mines and Mineral Resources Research Bulletin 17, 92 p.

______1948, Studies of American fusulinids: University of Kansas Paleontological Contributions, Protozoa, Article 1, p. 1–184.

Verville, G. J.; Sanderson, G. A.; and Madsen, M. E., 1986, Pennsylvanian fusulinids from the Fra Cristobal Range, Sierra County, New Mexico: New Mexico Geological Society Guidebook, 37th Field Conference, p. 215–223.

Zachry, D. L.; and Sutherland, P. K., 1984, Stratigraphy and depositional framework of the Atoka Formation (Pennsylvanian), Arkoma basin of Arkansas and Oklahoma, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 9–17.

Plates

PLATE 1

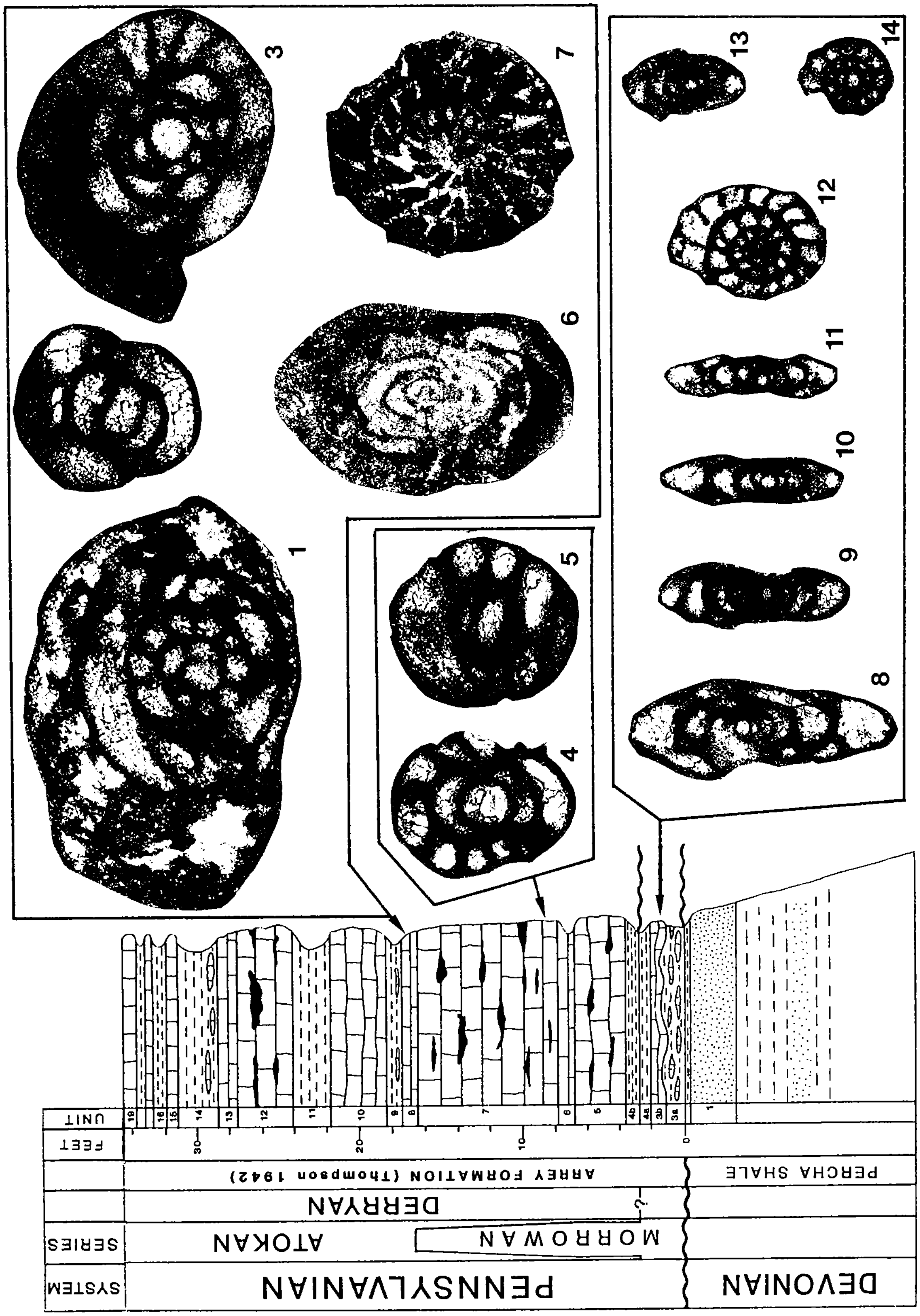
"Arrey Formation," Derryan type section, MS 3 (figures ×100 except as noted).

Figures. 1–5.—*Eoschubertella mexicana* Thompson. **1**, (OU 10898) axial section, unit 3-8 top. **2**, (OU 10896) axial section of juvenile, unit 3-8 top. **3**, (OU 10897) axial section, unit 3-8 top. **4**, (OU 10892) axial section of juvenile, unit 3-7 base. **5**, (OU 10893) tangential section of juvenile, unit 3-7 base.

Figures. 6,7.—*Pseudoendothyra* sp. **6**, (OU 10886) axial section ×40, unit 3-8 top. **7**, (OU 10887) sagittal section ×40, unit 3-8 top.

Figures. 8–12.—*Millerella* spp. **8**, (OU 10866) axial section, unit 3-3. **9**, (OU 10868) axial section, unit 3-3. **10**, (OU 10870) axial section, unit 3-3. **11**, (OU 10869) axial section, unit 3-3. **12**, (OU 10878) sagittal section, unit 3-3.

Figures. 13,14.—*Eostaffella* sp. **13**, (OU 10879) axial section, unit 3-3. **14**, (OU 10883) sagittal section, unit 3-3.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SYSTEM
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DEVONIAN
ATOKAN
MORROWAN
DERRYAN
?
ARREY FORMATION (Thompson 1942)
PERCHA SHALE
0
10
20
30

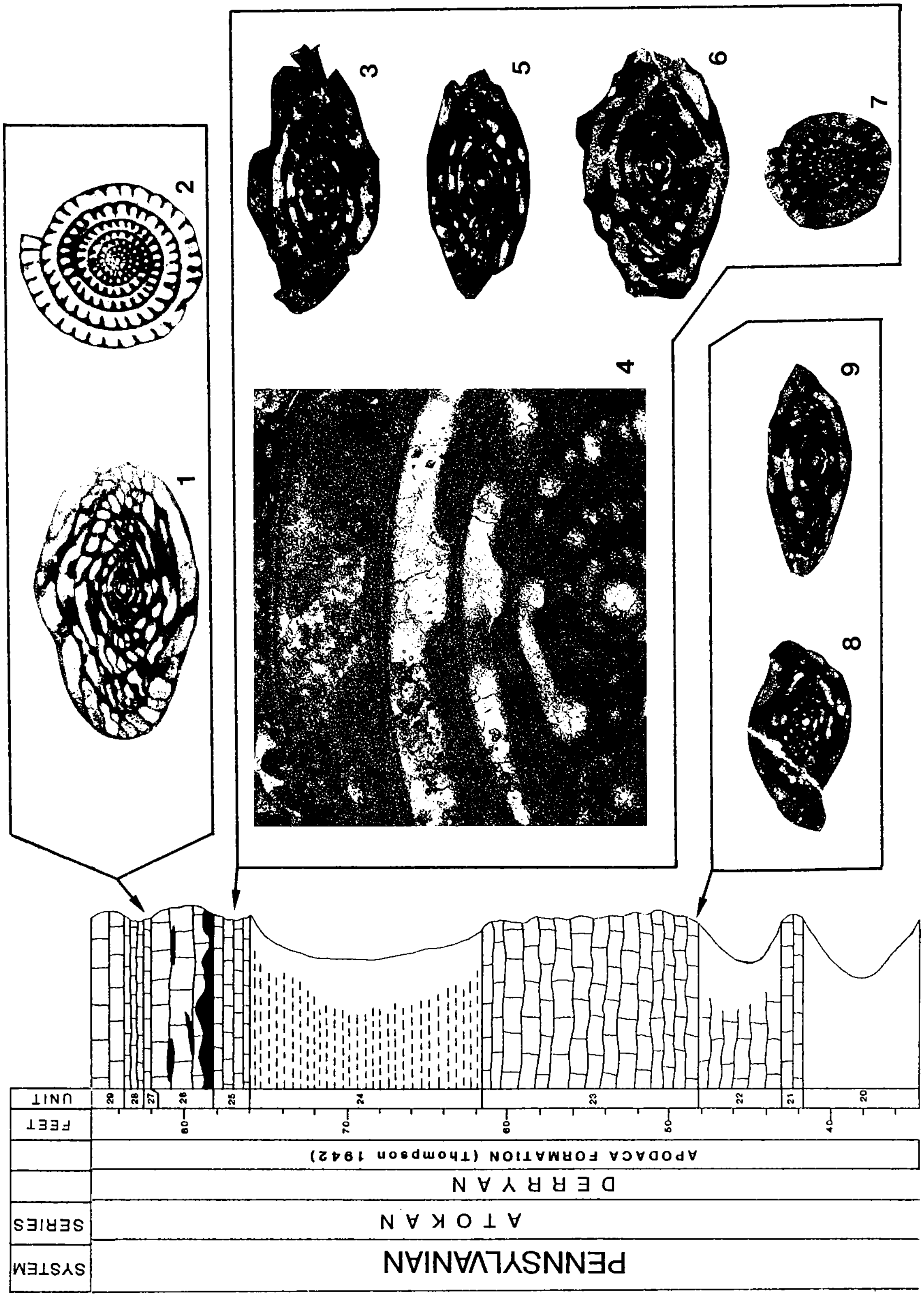
PLATE 2

"Apodaca Formation," Derryan type section (figures ×20 except as noted).

Figures 1,2.—*Fusulinella proxima* Thompson. **1**, (OU 10991) axial section, unit 3-27 base. **2**, (OU 10992) sagittal section, unit 3-27 base.

Figures 3–7.—*Fusulinella fugax* Thompson. **3**, (OU 10960) axial section, unit 3-25 base. **4**, (OU 10960) axial section ×100 showing thin but consistent diaphanotheca development in fourth, fifth and sixth volutions. **5**, (OU 10961) axial section, unit 3-25 base. **6**, (OU 10967) axial section, unit 3-25 base. **7**, (OU 10968) sagittal section, unit 3-25 top.

Figures 8,9.—*Profusulinella* cf. *fittsi*. **8**, (OU 11057) axial section, unit 3-23 0.5 ft above base. **9**, (OU 11058) axial section, unit 3-23 0.5 ft above base.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FEET
UNIT
PENNSYLVANIAN
ATOKAN
DERRYAN
APODACA FORMATION (Thompson 1942)
20
21
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23
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27
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40
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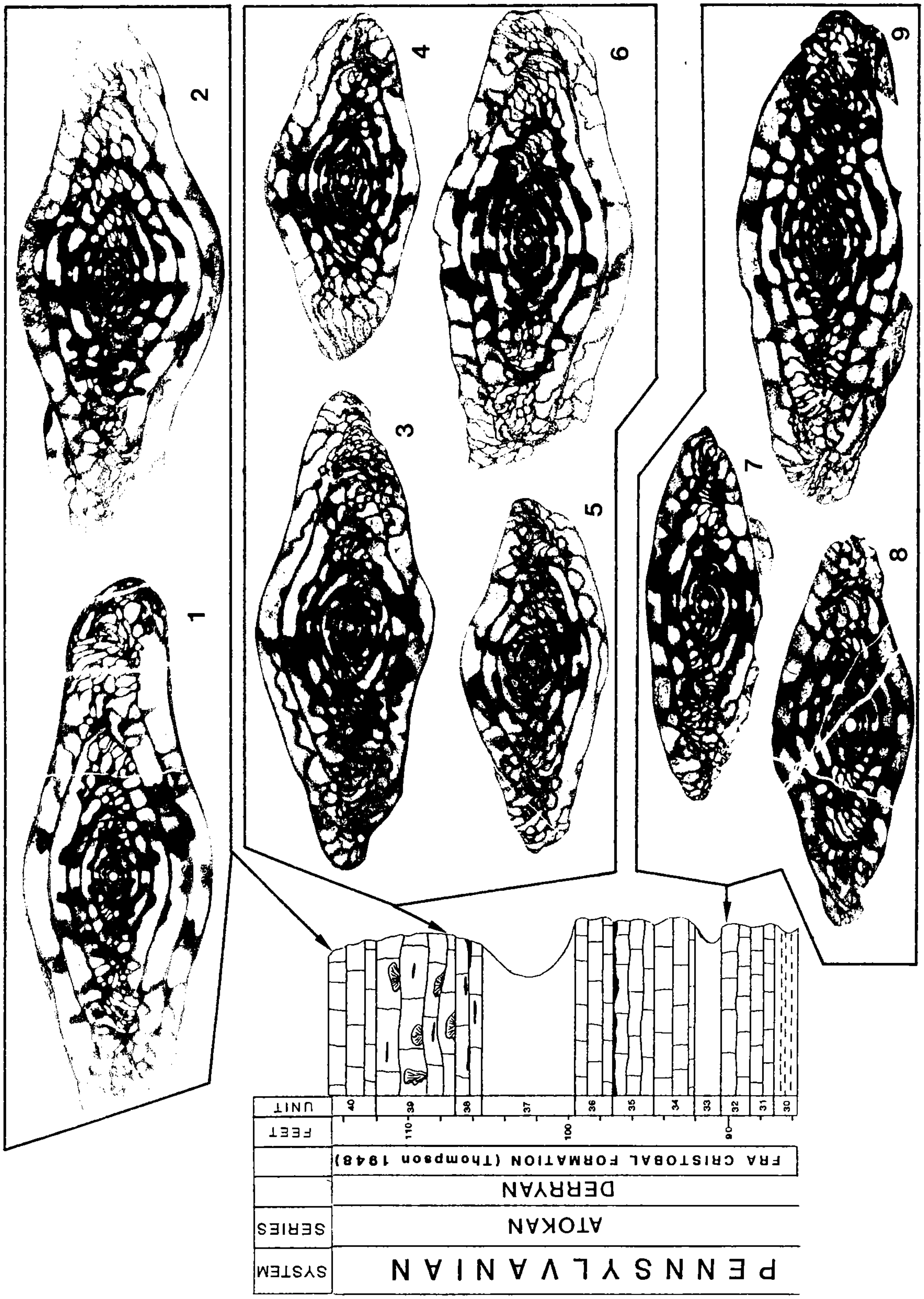
60
70
80

PLATE 3

"Fra Cristobal Formation," Derryan type section (all figures ×20).

Figures 1–6.—*Fusulinella devexa* Thompson. **1**, (OU 9183) axial section, unit 3-40 0.5 ft below top. **2**, (OU 9184) axial section, unit 3-40 0.5 ft below top. **3**, (OU 11007) axial section, unit 3-39 base. **4**, (OU 11009) axial section, unit 3-39 base. **5**, (OU 9185) axial section, unit 3-39 base. **6**, (OU 11008) axial section, unit 3-39 base.

Figures 7–9.—*Fusulinella acuminata* Thompson. **7**, (OU 10978) axial section, unit 3-32 0.5 ft below top. **8**, (OU 10975) axial section, unit 3-32 0.5 ft below top. **9**, (OU 10971) axial section, unit 3-32 0.5 ft below top.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SYSTEM
PENNSYLVANIAN
SERIES
ATOKAN
DERRYAN
FRA CRISTOBAL FORMATION (Thompson 1948)
FEET
90
100
110
UNIT
30
31
32
33
34
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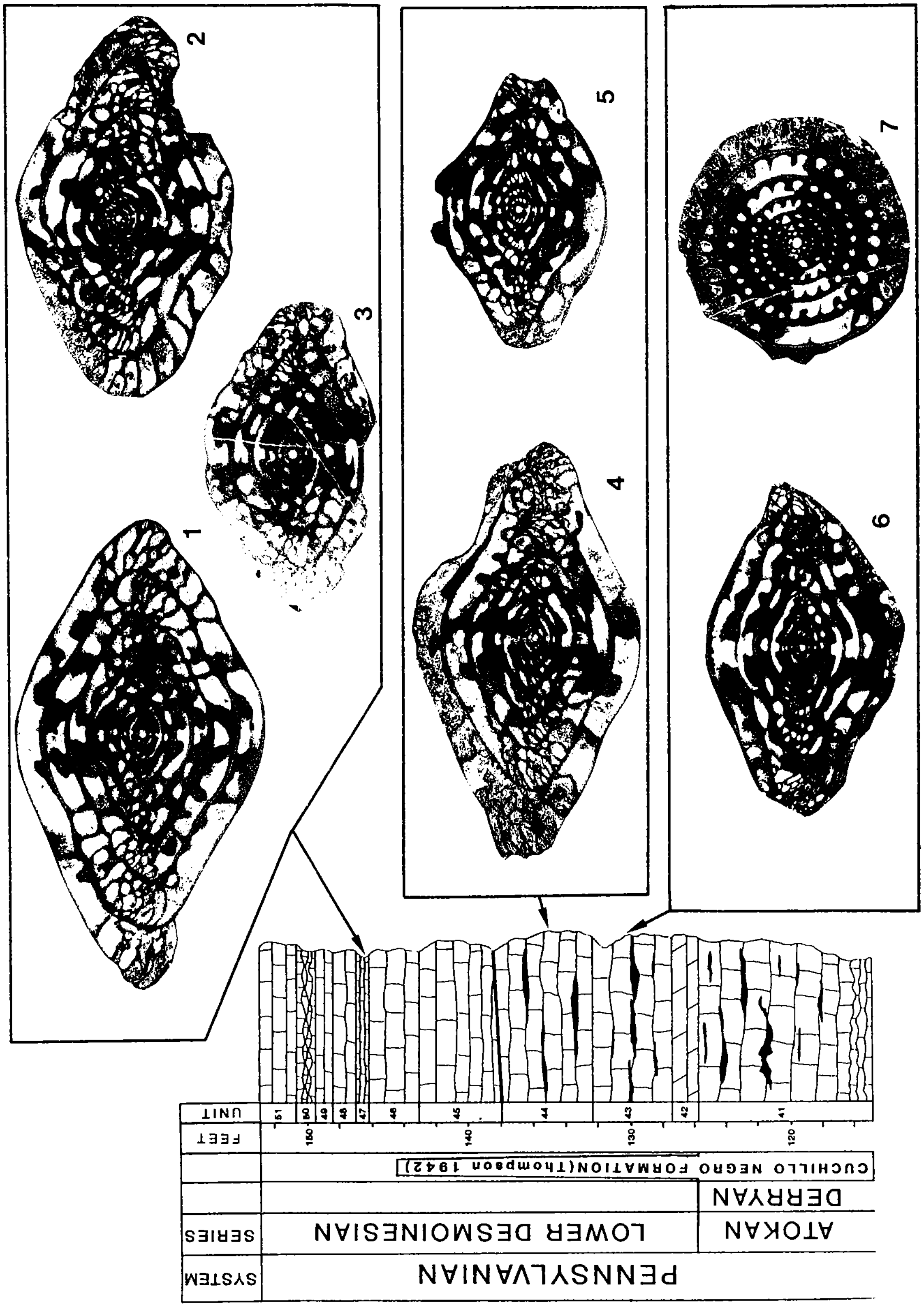
38
39
40

PLATE 4

"Cuchillo Negro Formation" and younger (basal Desmoinesian) strata immediately above type Derryan section (all figures ×20).

Figures 1–5.—*Beedeina insolita* (Thompson). **1**, (OU 11021) axial section, unit 3-47 base. **2**, (OU 11028) axial section, unit 3-47 base. **3**, (OU 11022) axial section, unit 3-47 base. **4**, (OU 11027) axial section, unit 3-44 3 ft above base. **5**, (OU 11024) axial section, unit 3-44 3 ft above base.

Figures 6,7.—*Fusulinella famula* Thompson. **6**, (OU 11020) axial section, unit 3-43 2.7 ft above base. **7**, (OU 11017) sagittal section, unit 3-43 2.7 ft above base.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SYSTEM
PENNSYLVANIAN
SERIES
ATOKAN
LOWER DESMOINESIAN
DERRYAN
CUCHILLO NEGRO FORMATION(Thompson 1942)
FEET
120
130
140
150
UNIT
41
42
43
44
45
46
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48
49
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51

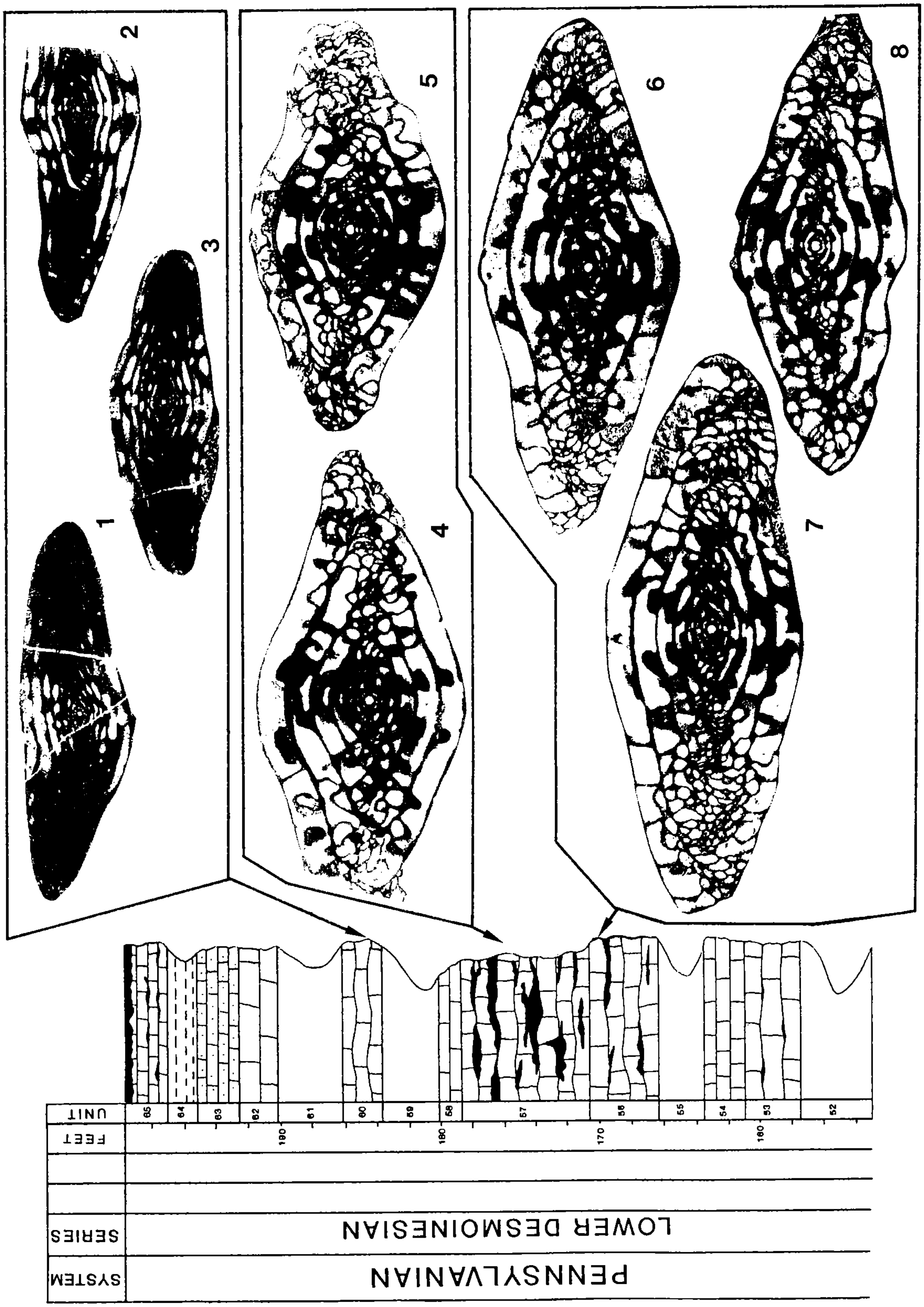
PLATE 5

Lower Desmoinesian Series above type Derryan section (all figures ×20).

Figures 1–3.—*Wedekindellina excentrica* Roth and Skinner. **1**, (OU 11062) axial section, unit 3-60 base. **2**, (OU 11064) axial section, unit 3-60 base. **3**, (OU 11063) axial section, unit 3-60 base.

Figures 4,5.—*Beedeina* aff. *hayensis*. **4**, (OU 11034) axial section, unit 3-57 top. **5**, (OU 11035) axial section, unit 3-57 5 ft above base.

Figures 6–8.—*Beedeina*? aff. *rockymontana*. **6**, (OU 11086) axial section, unit 3-56 top. **7**, (OU 11090) axial section, unit 3-56 top. **8**, (OU 11088) axial section, unit 3-56 top.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SYSTEM
SERIES
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PENNSYLVANIAN
LOWER DESMOINESIAN
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Oklahoma Geological Survey Circular 94, 1992

Stratigraphic Distribution of Non-Fusulinacean Foraminifers in the Marble Falls Limestone (Lower–Middle Pennsylvanian), Western Llano Region, Central Texas

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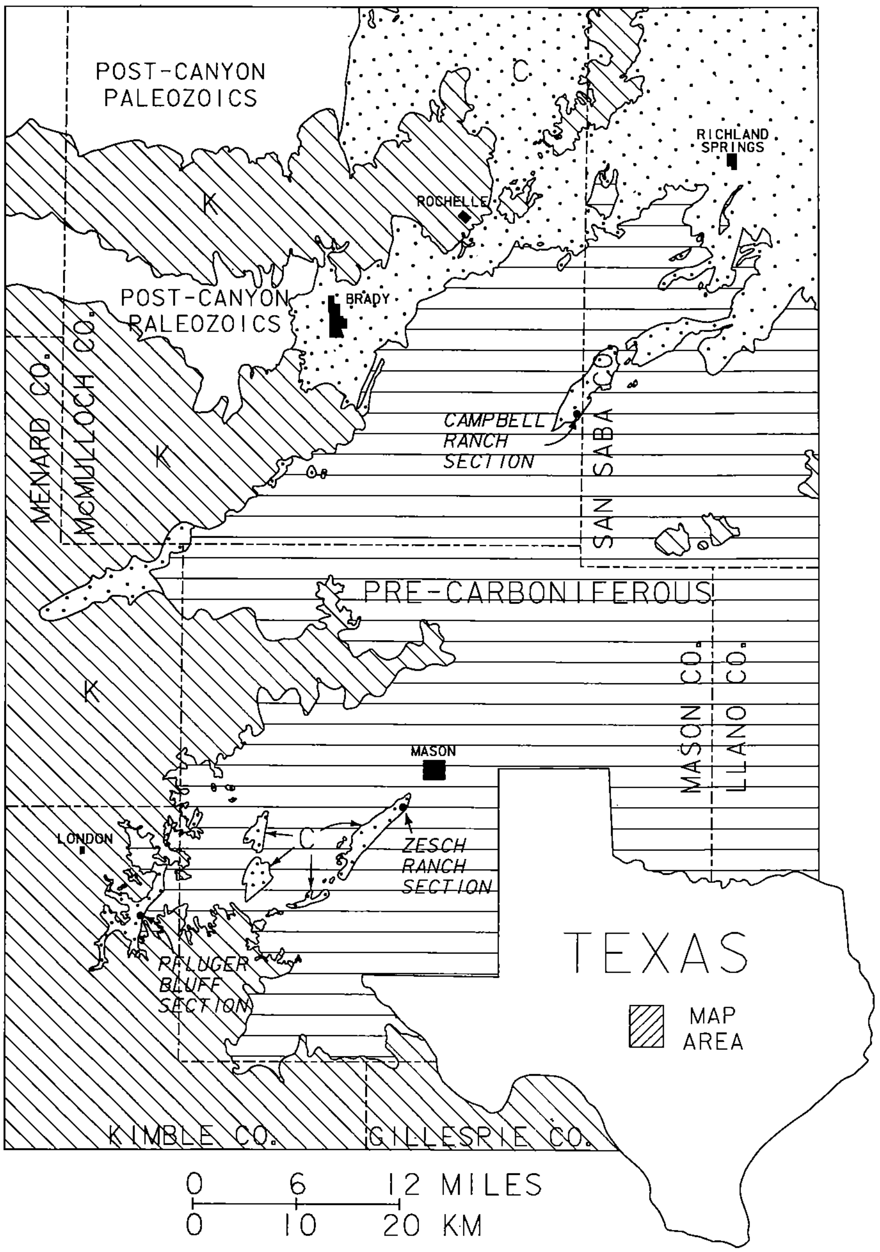
ABSTRACT.—The Marble Falls Limestone (Morrowan–Atokan) in the western Llano region of central Texas yields 22 genus- and species-level taxa of smaller foraminifers whose distributions are summarized relative to an independently established local fusulinacean zonation. Most of the non-fusulinaceans are long-ranging and probably of only minor biostratigraphic value. Documentation of their occurrences is important, nevertheless, in that at least two species are new (albeit not formally described here), and several others are reported for the first time in North America. Still others, while known in North America, are demonstrated to range significantly lower than previously thought. The assemblage from the medial and late Atokan portion of the Marble Falls is remarkably similar to ones described from the Moscovian of the Soviet Union and other parts of the Eurasian–Arctic faunal realm. That similarity suggests a much higher degree of cosmopolitanism than is generally recognized.

INTRODUCTION

This article is a companion to that by Groves (1991) in which fusulinacean-based biostratigraphic correlations were used to help interpret the lithostratigraphy of the Marble Falls Limestone in the western Llano area, central Texas. Its purpose is to record the distribution of smaller, non-fusulinacean foraminifers with respect to the independently established fusulinacean zonation. The reader is referred to Groves (1991) for a discussion of regional lithostratigraphy and fusulinacean biostratigraphy.

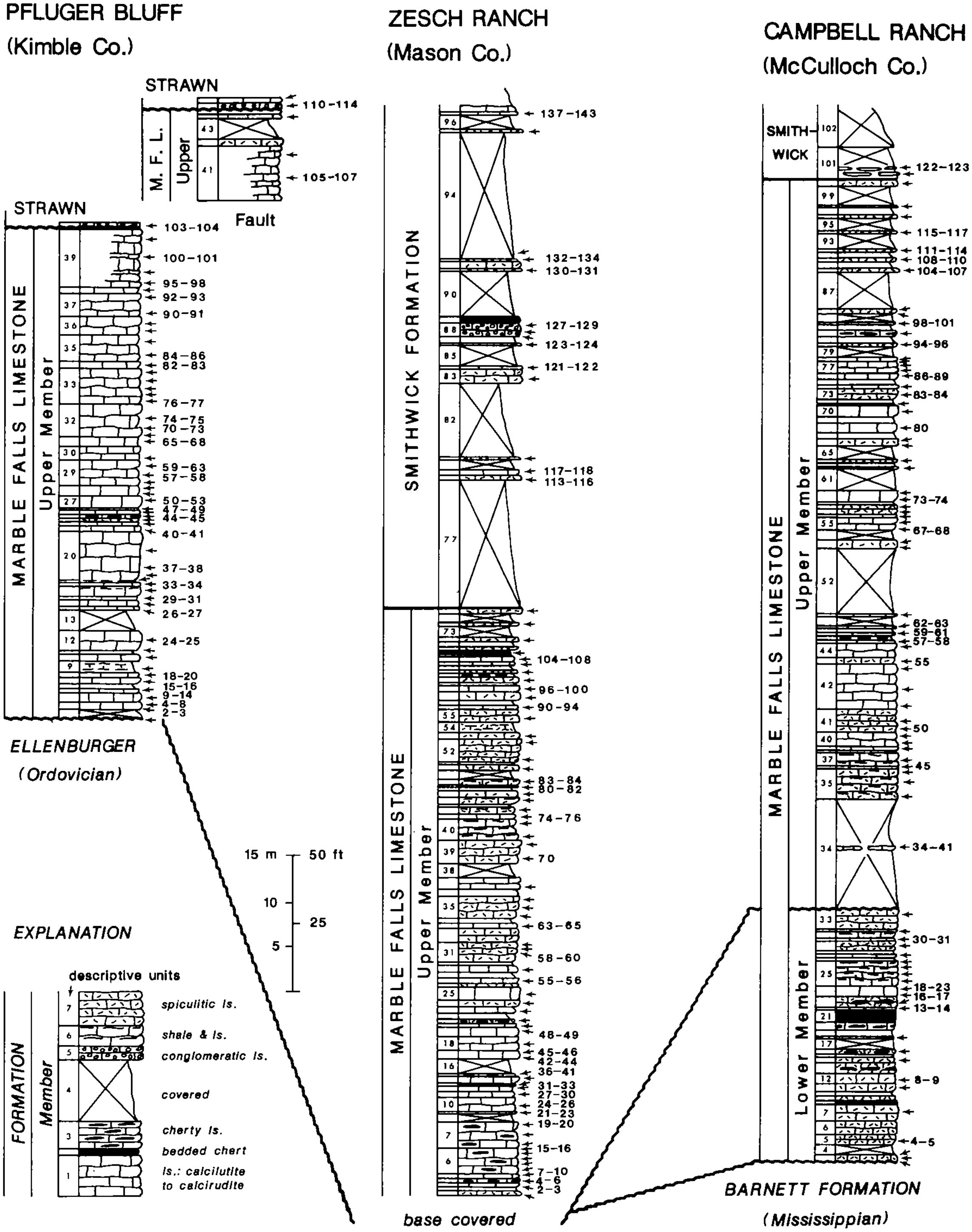
Foraminifer-bearing samples come from three measured sections in McCulloch, Mason, and Kimble Counties (Text-figs. 1,2). The Campbell Ranch section in easternmost McCulloch County consists of 360 ft of the Marble Falls overlain conformably by an undetermined thickness of the Smithwick Formation. There, the Marble Falls is divided into lower and upper members, and is interpreted to rest unconformably on the Upper Mississippian Barnett Formation. The members of the Marble Falls are separated by an unconformity that corresponds to the Morrowan–Atokan boundary. The lower member is Morrowan throughout its distribution across the Llano uplift. The upper member is entirely Atokan, but its age within the Atokan varies considerably, being older on the eastern and northern flanks of the uplift and progressively younger on the western and southwestern flanks. The Zesch Ranch section in central Mason County includes 215 ft of the upper member directly above the Barnett. The lower member is absent here and in all areas west of central McCulloch County as a result of nondeposition, removal by erosion, or some combination of both. The upper member at Zesch Ranch is overlain by 183 ft of Smithwick, which locally contains several thin, fossiliferous limestones. The Pfluger Bluff section in northeastern Kimble County includes 177 ft of upper Marble Falls bracketed unconformably by the Ordovician Ellenburger Group and the Desmoinesian upper Strawn Group. The Barnett and older post-Ellenburger strata were removed by pre-upper Marble Falls erosion, whereas the Smithwick and lower Strawn were stripped by pre-upper Strawn erosion.

Composited ranges of non-fusulinacean foraminifers are depicted on Text-figure 3 relative to lithostratigraphic units, the local fusulinacean zonation, and chronostratigraphic subdivisions of the Lower and Middle Pennsylvanian. Twenty-two taxa have been recognized, most of which are long-ranging and only minimally useful biostratigraphically. That notwithstanding, the fauna is noteworthy for containing at least two new species (not formally described) and several others which heretofore have not been documented from Morrowan or Atokan rocks in North America. Indeed, some taxa were previously known only from the Soviet Union and other areas of the Eurasian–Arctic faunal realm. Their presence in the Marble Falls suggests a greater degree of cosmopolitanism than is generally recognized, which in turn may be attributable in part to the dearth of studies of North American Pennsylvanian smaller foraminifers.

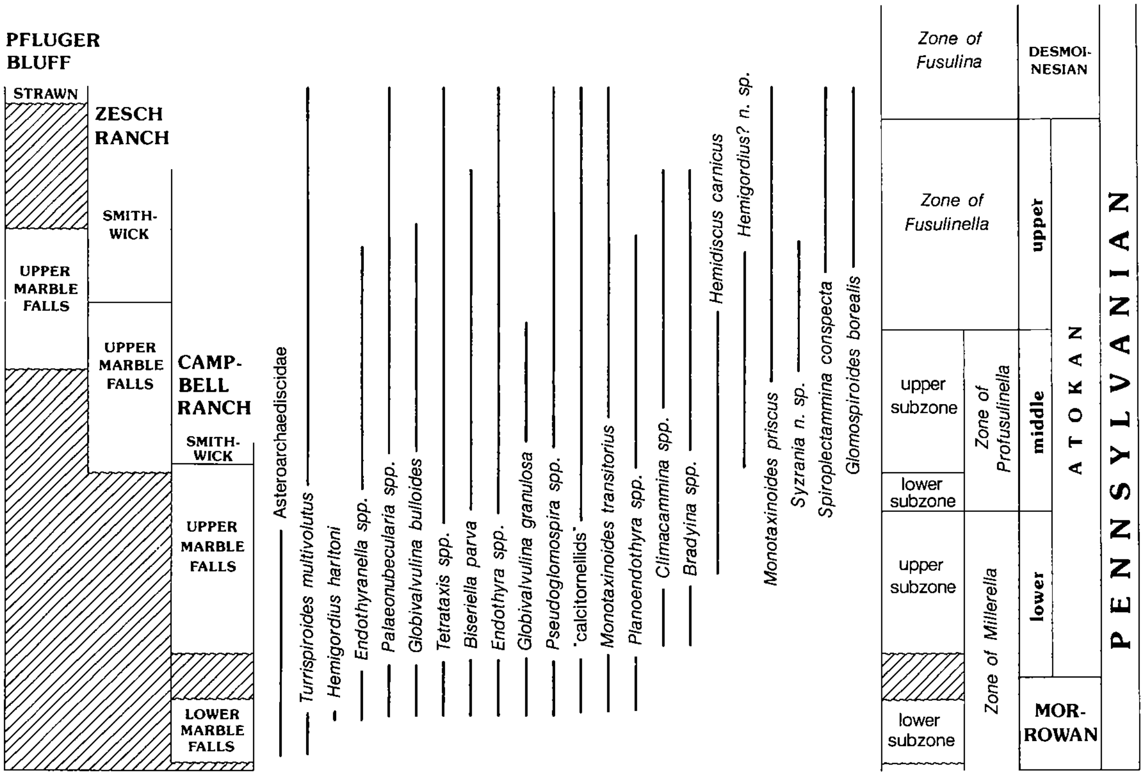


Text-figure 1. Index map to western Llano region and sampled localities. Horizontally ruled pattern = pre-Carboniferous sedimentary and crystalline formations; stippled pattern (C) = Chappel through Canyon Carboniferous formations; diagonally ruled pattern (K) = Cretaceous formations.

Syzrania n. sp. occurs in rocks assigned to the upper subzone of the Zone of *Profusulinella* and the Zone of *Fusulinella*. This species is morphologically primitive, and its medial to late Atokan age is the oldest North American occurrence of the genus. The test of *Hemigordius*? n. sp. is highly evolute, planispirally coiled, and nonseptate with rather coarsely granular wall structure. Unique secondary deposits are developed as microgranular fillings along the line of suture between the tubular chamber and the preceding volution. This combination of characters differs from that in all previously described genera including *Pseudoammodiscus* (which lacks secondary deposits) and true *Hemigordius* (in which coiling is initially skewed, with external, not internal, side thickenings). Specimens occur in the



Text-figure 2. Columnar stratigraphic sections and positions of microfossil samples at Campbell Ranch, Zesch Ranch, and Pfluger Bluff.



Text-figure 3. Local ranges of smaller foraminifers relative to fusulinacean zones, lithostratigraphic units, and North American chronostratigraphic subdivisions of the Pennsylvanian. Vertical dimension scaled approximately to stratigraphic thickness with some distortion.

upper subzone of the Zone of *Profusulinella* and the Zone of *Fusulinella*. *Glomospiroides borealis* is reported in North America for the first time in rocks assigned to the Zones of *Fusulinella* and *Fusulina*. The types are from the upper Moscovian Podolsky and Myachkovsky Horizons of the Russian Platform (Reitlinger, 1950). *Spiroplectammina conspecta*, also originally described from the Moscovian of the Russian Platform (Kashirsky and Podolsky Horizons) (Reitlinger, 1950), closely resembles forms reported from Virgilian rocks in north-central Texas and the Midcontinent (Cushman and Waters, 1928; Toomey, 1972). The present occurrences in the Zones of *Fusulinella* and *Fusulina* significantly lower the documented North American range. *Monotaxinoides priscus* ranges locally from the upper subzone of the Zone of *Profusulinella* into the Zone of *Fusulina*. This is the first North American report of the species, originally described from the Serpukhovian of the Donbass (Brazhnikova and Yartseva, 1956). Finally, *Hemidiscus carnicus* occurs in strata assigned to the upper subzone of the Zone of *Millerella* through a portion of the Zone of *Fusulinella*. This species is normally associated with Old World Permian assemblages.

Acknowledgments

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FAUNAL DOCUMENTATION

Occurrences and brief remarks (where appropriate) are given below for each of the encountered taxa. Identifications are based on examination of more than 3,000 photographed specimens. The collection is housed at the University of Oklahoma Museum of Natural History, Invertebrate Paleontology Repository (OU).

Syzrania n. sp.
Pl. 1, Figs. 22–31

Remarks and occurrence.—These specimens possess a relatively weakly developed outer hyaline wall layer. This differs from all other species in the genus in which the thickness of the hyaline layer greatly exceeds that of the inner micrograpular layer. The thin hyaline layer is morphologically primitive and suggests an evolutionary derivation from *Earlandia* Plummer. These are the oldest North American specimens in the genus. Specimens are from Zesch Ranch samples 71, 73, and 101; and Pfluger Bluff samples 2, 7, 8, 24, 36, 39, 43, 45, 46, 49, 54, 56, 58, 60, 64, 67, 68, 71, 73, 79, 80, 82, 83, 88, 95, and 99.

Pseudoglomospira Bykova, 1955

Pl. 1, Figs. 1–7

Occurrence.—Campbell Ranch samples 18, 20–24, 28, 30–32, 35, 38, 41, 45, 46, 56, 60, 61, 64, 70, 75, 90, 92, 101, 104–106, 108, 113, and 114; Zesch Ranch samples 2, 3, 7–11, 14, 19–23, 27–29, 35, 37–39, 63, 67–69, 74, 76, 78, 90–92, 94, 98–100, 104–106, 108, 109, 112, 113, 115, 117, 118, 121–124, 130, 132, 134, 137, 139, 140, 142, and 143; and Pfluger Bluff samples 2–8, 10–13, 15–19, 23, 24, 27, 29–32, 34, 45–48, 56, 58, 60, 64–66, 68, 70, 71, 73, 76, 77, 88, 89, 100, 104, 108, and 110–113.

"calcitornellids"

Pl. 1, Figs. 8–13

Remarks and occurrence.—These specimens include representatives of three nominal genera: *Calcitornella, Calcivertella,* and *Trepeilopsis*, all erected by Cushman and Waters (1928). All three "genera" probably are a single highly variable taxon in which the growth habit is determined as much by the nature of the substrate as by the genotype of the organism. Specimens are from Campbell Ranch samples 20, 27, 28, 35, 38–40, 43, 45, 57–59, 86, and 87; Zesch Ranch samples 19, 20, 52, 55, 80, 92, 93, 95, 103, 104, 108, 117, 118, 121, 123, 128, 130, 132, 134, and 137; and Pfluger Bluff samples 8, 30, 34, 45, 48, 54, 67–69, 78, 88, 104, 113, and 115.

Glomospiroides borealis (Reitlinger, 1950)

Pl. 1, Figs. 14–18

Occurrence.—Pfluger Bluff samples 44, 45, 48, 67, 68, and 115.

Palaeonubecularia Reitlinger, 1950

Pl. 1, Figs. 19–21

Occurrence.—Campbell Ranch samples 17, 21, 34–37, 57–60, 70, 88, 90–92, 106, 113, and 114; Zesch Ranch samples 7, 10, 21, 23, 36, 37, 39, 41, 69, 72, 91, 93, 94, 107, 117, 118, 122, and 125; and Pfluger Bluff samples 5, 8, 9, 11, 13, 27, 29, 31, 47, 49, 62, 64, 78, 91, 104, 110, and 112–115.

Climacammina Brady, *in* Etheridge, 1873

Pl. 2, Figs. 11–16

Occurrence.—Campbell Ranch samples 34, 36, 38, 39, 74, 83–85, 94, and 98; Zesch Ranch samples 4, 19, 43, 48, 59, 62, 69, 80, 96, 98, 107, 113, 114, and 137; and Pfluger Bluff samples 4, 7, 9, 12, 13, 20, 30, 62, 96, 97, and 102.

Tetrataxis Ehrenberg, 1854

Pl. 2, Figs. 5–10

Occurrence.—Campbell Ranch samples 17, 37, and 120; Zesch Ranch samples 3, 29, 43, 48, 49, 52, 66, 69, 84, 91–93, 96, 99, 100, 105, 113, 117, 118, 122–129, 137, 139, 140, and 142; and Pfluger Bluff samples 18–20, 27, 37, 39, 44, 45, 48, 55, 59–63, 66, 67, 69–71, 76–84, 87, 89, 91, 101, 106–108, 111, 113, and 114.

Spiroplectammina conspecta Reitlinger, 1950

Pl. 2, Figs. 17–24

Occurrence.—Zesch Ranch samples 117, 118, and 121–124; and Pfluger Bluff samples 60, 113, and 115.

Hemigordius harltoni Cushman and Waters, 1928

Pl. 2, Figs. 1–4

Occurrence.—Campbell Ranch samples 13, 16, and 17.

Hemigordius? n. sp.

Pl. 3, Figs. 1–16

Remarks and occurrence.—Planispiral coiling, moderately granular wall structure, and internal microgranular deposits along the line of suture between the tubular chamber and the preceding volution distinguish these specimens from all previously described forms. They should be assigned to a new species in a new genus. Specimens are from Campbell Ranch samples 117–120; Zesch Ranch samples 8, 22, 23, 30, 42, 44, 52, 55, 59, 60, 62, 64, 65, 69, 72, 73, 75, 78, 99, 105, and 117; and Pfluger Bluff samples 2, 5, 8, 18, 22, 23, 31, 47, 83, and 89.

Turrispiroides multivolutus (Reitlinger, 1949)

Pl. 3, Figs. 17–26

Occurrence.—Campbell Ranch samples 4, 10, 11, 13, 14, 16–28, 30–32, 34, 36–41, 43, 45, 46, 49, 52–54, 56–61, 67, 72, 74, 75, 80, 85–87, 90, 93, 94, 98–100, 104–106, 109, 112, 115, 116, 119–121, and 123; Zesch Ranch samples 2, 3, 6–10, 14, 16–18, 20–29, 35, 37, 38, 40, 47, 52, 55, 58–65, 69, 71, 72, 74–78, 80, 83, 84, 88, 90, 91, 93–98, 100, 102, 104–106, 109, 113–115, 117, 118, 121, 124, 130, 132, 134, and 143; and Pfluger Bluff samples 2–5, 7–12, 14, 18, 19, 22–25, 27, 29, 32–34, 36, 38, 39, 41–46, 48–52, 54, 57, 58, 65–70, 72, 73, 76, 83, 88, 103, 104, 108, 112, 114, and 115.

Monotaxinoides transitorius Brazhnikova and Yartseva, 1956

Pl. 3, Figs. 27–33

Occurrence.—Campbell Ranch samples 19, 22, 45–47, 49, 50, 52, 53, 56–61, 64, 66, 67, 81, 93–96, 99, 100, 103–108, 111, 115, 117, 120, and 121; Zesch

Ranch samples 13–18, 21–24, 27, 35, 42, 58–63, 65, 66, 69, 71–74, 77, 78, 80, 81, 83, 88, 94–97, 100–110, 113–115, 117, 119, 122–124, 130, 132–134, 137, and 141–143; and Pfluger Bluff samples 12, 14–18, 22, 24, 26, 27, 29, 32, 36–48, 51, 53, 56, 60, 64, 66, 67, 70, 71, 76, 78, 79, 83, 88, 89, 101, 103, 108, 112, 113, and 115.

Monotaxinoides priscus Brazhnikova and Yartseva, 1956

Pl. 3, Figs. 34–40

Occurrence.—Zesch Ranch samples 66 and 118; and Pfluger Bluff samples 54, 56, 58, 67, 71, 110, and 114.

Hemidiscus carnicus Schellwien, 1898

Pl. 4, Figs. 1–7

Occurrence.—Campbell Ranch samples 64, 114, and 117; Zesch Ranch samples 9, 60, 62, 63, 68, 69, 80–82, 84, and 94; and Pfluger Bluff samples 3, 30, 34, and 47.

Biseriella parva (Chernysheva, 1948)

Pl. 4, Figs. 8–13

Occurrence.—Campbell Ranch samples 16, 17, 20–22, 27, 28, 30, 31, 34–41, 43, 45–47, 49, 52, 53, 56–64, 69, 73–75, 86–89, 95, 99–101, 103–106, 108–111, 118–121, and 123; Zesch Ranch samples 2–12, 14, 17, 21–26, 28–30, 32, 33, 35, 37–39, 42, 44, 48, 49, 55, 58, 60, 62, 64, 67, 69, 72, 74, 75, 80, 82, 96, 104, 106, 107, 114, 115, 125, 128, and 137; and Pfluger Bluff samples 2–16, 19, 20, 22–34, 39, 42, 44–53, 60, 61, 63, 68, 70, 71, 73, 75–77, 80, 91, and 105–107.

Globivalvulina bulloides (Brady, 1876)

Pl. 4, Figs. 14–19

Occurrence.—Campbell Ranch samples 17–19, 21, 35, 98, 99, 110, and 114; Zesch Ranch samples 3, 7–10, 23–25, 28, 30, 35, 44, 50, 55, 67, 71, 72, 76, 78, 91, 93, 94, 124, and 125; and Pfluger Bluff samples 7, 8, 13, 26-29, 33, 51, 52, 66, and 76.

Globivalvulina granulosa Reitlinger, 1950

Pl. 4, Figs. 20–24

Occurrence.—Campbell Ranch samples 18, 89, 99, and 106; Zesch Ranch sample 99; and Pfluger Bluff samples 9, 13, 14, 18, and 19.

Bradyina Möller, 1878

Pl. 5, Figs. 11–17

Occurrence.—Campbell Ranch samples 34, 36–38, 41, 43, 57–59, 61, 86, 87, 94–96, 98–101, 104, 108, 118, 119, and 123; Zesch Ranch samples 3, 6, 27, 36, 42, 44, 91, 92, 98, 113–115, 122, 137, and 140; and Pfluger Bluff samples 2, 3, 5, 10, 11, 16, 18, 19, 27, 29, 30, 38, 60–62, 67, 74, and 75.

Endothyra Phillips, 1846, *emend.* Brady, 1876, *emend.* China, 1965

Pl. 5, Figs. 18–21

Occurrence.—Campbell Ranch samples 16, 18–21, 26, 31, 34, 36–39, 41, 43, 53, 59–61, 63, 67, 69, 74, 75, 77, 83–94, 96, 98, 101, 108, 109, 111, 113, 114, and 117–123; Zesch Ranch samples 2–4, 6, 7, 10, 13, 18, 19, 22, 23, 26, 27, 35, 38, 42–44, 47–49, 51, 52, 54, 58, 59, 64, 66, 67, 69, 72–76, 78, 82, 84, 91–93, 95, 96, 98, 100, 102, 104–107, 112–115, 117, 118, 121–124, 127–130, 133, 141, and 143; and Pfluger Bluff samples 2–14, 16, 18–20, 23, 27, 29, 31, 32, 34, 38, 40, 42, 44–46, 48, 53, 54, 56, 58, 60–64, 66, 68, 71, 73, 75–78, 80–83, 86, 89, 91, 95–97, 101, 104, 106, 107, and 112.

Endothyranella Galloway and Harlton, *in* Galloway and Ryniker, 1930

Pl. 5, Figs. 2, 5, 6

Occurrence.—Campbell Ranch samples 15, 22, 23, and 121; Zesch Ranch samples 21, 69, 96, 99, and 115; and Pfluger Bluff samples 43, 44, 46, 56, 58, 66, 67, 71, and 107.

Planoendothyra Reitlinger, 1959

Pl. 5, Figs. 1, 4, 7–9

Occurrence.—Campbell Ranch samples 24, 37, 94, 96, 101, and 120; Zesch Ranch samples 4, 7, 9, 13–15, 17, 19, 90–92, 94, 115, and 122; and Pfluger Bluff samples 5, 12, 13, 27, 41, 44, 46, 61, 62, 68, 83, 97, 99–102, and 107.

Asteroarchaediscidae (undifferentiated)

Pl. 5, Figs. 3, 10

Occurrence.—Campbell Ranch samples 3, 10, 12, 14, 17, 20, 22, 28, 29, 32, and 74.

REFERENCES CITED

Brady, H. B., 1876, A monograph of Carboniferous and Permian Foraminifera (the genus *Fusulina* excepted): Palaeontological Society, London, v. 30, 166 p.

Brazhnikova, N. E.; and Yartseva, M. V., 1956, The evolution of the genus *Monotaxis* [in Russian]: Akademiya Nauk SSSR, Voprosy Mikropaleontologii, no. 1, p. 62–72.

Bykova, E. V., 1955, On the Foraminifera, Radiolaria, and Ostracoda in the Devonian Period of the Volga–Ural region [in Russian]: Vsesoyuznyy Neftyanoy Nauchno-Issledovatel'skiy Geologorazvedochnyy Institut, Trudy 87, Mikrofauna SSR, no. 6, p. 1–141.

Chernysheva, N. E., 1948, Some new species of Foraminifera from the Visean of the Makarov District [in Russian]: Akademiya Nauk SSSR, Geologicheskiy Institut, Trudy 62, no. 19, p. 246–250.

China, W. E., 1965, Opinion 724. *Endothyra bowmani* Phillips (1846) (Foraminifera)—Validated under the plenary powers (International Commission of Zoological Nomenclature Opinion 724): Bulletin of Zoological Nomenclature, v. 22, pt. 1, p. 37–39.

Cushman, J. A.; and Waters, J. A., 1928, Some Foraminifera from the Pennsylvanian and Permian of Texas: Contributions from the Cushman Laboratory for Foraminiferal Research, v. 4, no. 2, p. 31–55.

Ehrenberg, C. G., 1854, Mikrogeologie: L. Voss, Leipzig, 374 p.

Etheridge, R., Jr., 1873, Notes on certain genera and species mentioned in the foregoing lists: Scotland Geological Survey Memoir, Explanation of Sheet 23, appendix 2, p. 93–107.

Galloway, J. J.; and Ryniker, C., 1930, Foraminifera from the Atoka Formation of Oklahoma: Oklahoma Geological Survey Circular 21, 37 p.

Groves, J. R., 1991, Fusulinacean biostratigraphy of the Marble Falls Limestone (Pennsylvanian), western Llano region, central Texas: Journal of Foraminiferal Research, v. 21, no. 1, p. 67–95.

Möller, V. von, 1878, Die spiral-gewundenen Foraminiferen des Russischen Kohlenkalks: Mémoires de l'Académie Impériale des Sciences de St. Pétersbourg, 7ème Série, v. 25, no. 9, 147 p.

Phillips, J., 1846, On the remains of microscopic animals in the rocks of Yorkshire: Geological and Polytechnic Society of West Riding, Yorkshire, v. 2, p. 274–285.

Reitlinger, E. A., 1949, An account of the smaller Foraminifera in the lower part of the Middle Carboniferous in the central Ural and Kama regions [in Russian]: Akademiya Nauk SSSR, Izvestiya Seriya Geologicheskaya, v. 6, p. 149–164.

______1950, Foraminifera from the Middle Carboniferous deposits of the central part of the Russian Platform (exclusive of the family Fusulinidae) [in Russian]: Akademiya Nauk SSSR, Instituta Geologicheskogo Nauk, Trudy, v. 126, no. 47, p. 1–127.

______1959, Order Endothyrida, *in* Rauzer-Chernousova, D. M.; and Fursenko, A. V. (eds.), Protozoa, *part 1 of* Orlov, Yu. A. (ed.), Fundamentals of paleontology [in Russian]: Izdatel'stvo Akademiya Nauk SSSR, p. 190–201.

Schellwien, R. J., 1898, Die Fauna des Karnischen Fusulinenkalks: Palaeontolographica, v. 44, pts. 5–6, p. 237–282.

Toomey, D. F., 1972, The biota of the Pennsylvanian (Virgilian) Leavenworth Limestone, Midcontinent region. Part 3: Distribution of calcareous Foraminifera: Journal of Paleontology, v. 46, p. 276–298.

PLATE 1

(All figures ×100)

Figures 1–7.—*Pseudoglomospira* spp., all random sections. **1**, CR spl. 61 (OU 8712g). **2**, CR spl. 31 (OU 8797a). **3**, CR spl. 104 (OU 8798a). **4**, PB spl. 8 (OU 8298b). **5**, PB spl. 113 (OU 8294b). **6**, CR spl. 106 (OU 8799a). **7**, PB spl. 2 (OU 8800a).

Figures 8–13.—"calcitornellids." **8**, longitudinal section, ZR spl. 134 (OU 8801a). **9**, horizontal section, CR spl. 40 (OU 8802a). **10**, longitudinal section, ZR spl. 95 (OU 8803a). **11**, vertical section, CR spl. 35 (OU 8728b). **12**, vertical section, CR spl. 39 (OU 8804a). **13**, longitudinal section, CR spl. 20 (OU 8635b).

Figures 14–18.—*Glomospiroides borealis* (Reitlinger, 1950), all random sections from PB. **14**, spl. 44 (OU 8306b). **15**, spl. 115 (OU 8805a). **16**, spl. 48 (OU 8651c). **17**, spl. 45 (OU 8647c). **18**, spl. 48 (OU 8651d).

Figures 19–21.—*Palaeonubecularia* spp., all random sections. **19**, PB spl. 115 (OU 8805b). **20**, CR spl. 17 (OU 8313g). **21**, ZR spl. 93 (OU 8806a).

Figures 22–31.—*Syzrania* n. sp., all longitudinal sections. **22**, PB spl. 88 (OU 10736a). **23**, PB spl. 54 (OU 10737a). **24**, PB spl. 80 (OU 10738a). **25**, ZR spl. 71 (OU 10739a). **26**, PB spl. 56 (OU 10740a). **27**, PB spl. 83 (OU 10741a). **28**, PB spl. 73 (OU 10742a). **29**, PB spl. 68 (OU 10743a). **30**, PB spl. 43 (OU 10744a). **31**, PB spl. 58 (OU 10745a).


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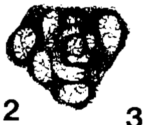
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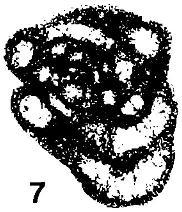
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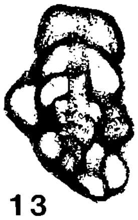
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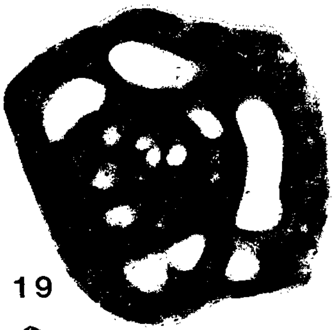
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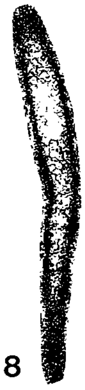
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PLATE 2

Figures 1–4.—*Hemigordius harltoni* Cushman and Waters, 1928, ×200; all from CR. **1,** oblique axial section, spl. 16 (OU 8639c). **2,** oblique sagittal section, spl. 13 (OU 10746a). **3,** oblique section, spl. 16 (OU 8639d). **4,** oblique section, spl. 17 (OU 8313h).

Figures 5–10.—*Tetrataxis* spp., ×40; all axial sections. **5,** PB spl. 27 (OU 8297d). **6,** PB spl. 18 (OU 10747a). **7,** PB spl. 101 (OU 8709b). **8,** ZR spl. 29 (OU 10748a). **9,** ZR spl. 124 (OU 8716b). **10,** PB spl. 44 (OU 8306c).

Figures 11–16.—*Climacammina* spp., ×20. **11,** transverse section, CR spl. 98 (OU 8786b). **12,** longitudinal section, ZR spl. 113 (OU 8329d). **13,** longitudinal section, PB spl. 9 (OU 10749a). **14,** longitudinal section, CR spl. 98 (OU 8786c). **15,** longitudinal section, ZR spl. 114 (OU 10750a). **16,** longitudinal section, PB spl. 96 (OU 8768b).

Figures 17–24.—*Spiroplectammina conspecta* Reitlinger, 1950, ×100; all longitudinal sections. **17,** ZR spl. 117 (OU 10751a). **18,** ZR spl. 122 (OU 8316b). **19,** ZR spl. 118 (OU 10752a). **20,** PB spl. 113 (OU 8294c). **21,** ZR spl. 117 (OU 10751b). **22,** ZR spl. 118 (OU 10752b). **23,** ZR spl. 118 (OU 10752c). **24,** ZR spl. 118 (OU 10752d).


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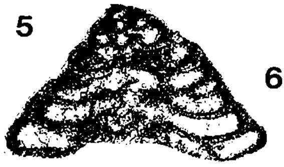
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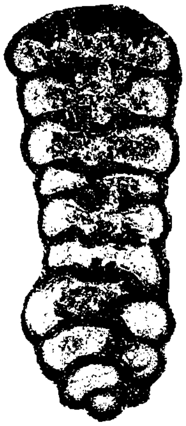
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PLATE 3

Figures 1–16.—*Hemigordius*? n. sp., ×100; 1–8, 10–14 axial and oblique axial sections; 9,15,16 sagittal sections. **1,** ZR spl. 60 (OU 8630c). **2,** ZR spl. 44 (OU 8300c). **3,** ZR spl. 42 (OU 10753a). **4,** ZR spl. 62 (OU 10754a). **5,** ZR spl. 72 (OU 8317f). **6,** ZR spl. 105 (OU 8765d). **7,** ZR spl. 8 (OU 8308d). **8,** ZR spl. 69 (OU 8785b). **9,** ZR spl. 55 (OU 8296c). **10,** PB spl. 31 (OU 10755a). **11,** CR spl. 117 (OU 10756a). **12,** ZR spl. 42 (OU 10753b). **13,** ZR spl. 23 (OU 10757a). **14,** ZR spl. 78 (OU 10758a). **15,** ZR spl. 64 (OU 10759a). **16,** ZR spl. 30 (OU 10760a).

Figures 17–26.—*Turrispiroides multivolutus* (Reitlinger, 1949), ×200; 17–20, 22–26 axial sections; 21 sagittal section. **17,** ZR spl. 100 (OU 8727b). **18,** CR spl. 32 (OU 8636b). **19,** PB spl. 58 (OU 10745b). **20,** CR spl. 30 (OU 10761a). **21,** ZR spl. 121 (OU 10762a). **22,** CR spl. 32 (OU 8636c). **23,** CR spl. 41 (OU 8319b). **24,** ZR spl. 60 (OU 8630d). **25,** CR spl. 28 (OU 10763a). **26,** PB spl. 29 (OU 8305d).

Figures 27–33.—*Monotaxinoides transitorius* Brazhnikova and Yartseva, 1956, ×200; 27–30,32,33 axial sections; 31 sagittal section; all from ZR. **27,** spl. 62 (OU 10754b). **28,** spl. 96 (OU 10764a). **29,** spl. 63 (OU 8299b). **30,** spl. 107 (OU 8725e). **31,** spl. 60 (OU 8630e). **32,** spl. 110 (OU 10765a). **33,** spl. 63 (OU 8299c).

Figures 34–40.—*Monotaxinoides priscus* Brazhnikova and Yartseva, 1956, ×200; all axial and oblique axial sections. **34,** PB spl. 114 (OU 10766a). **35,** ZR spl. 66 (OU 10767a). **36,** PB spl. 54 (OU 10737b). **37,** PB spl. 56 (OU 10740b). **38,** PB spl. 58 (OU 10745c). **39,** PB spl. 56 (OU 10740c). **40,** PB spl. 58 (OU 10745d).

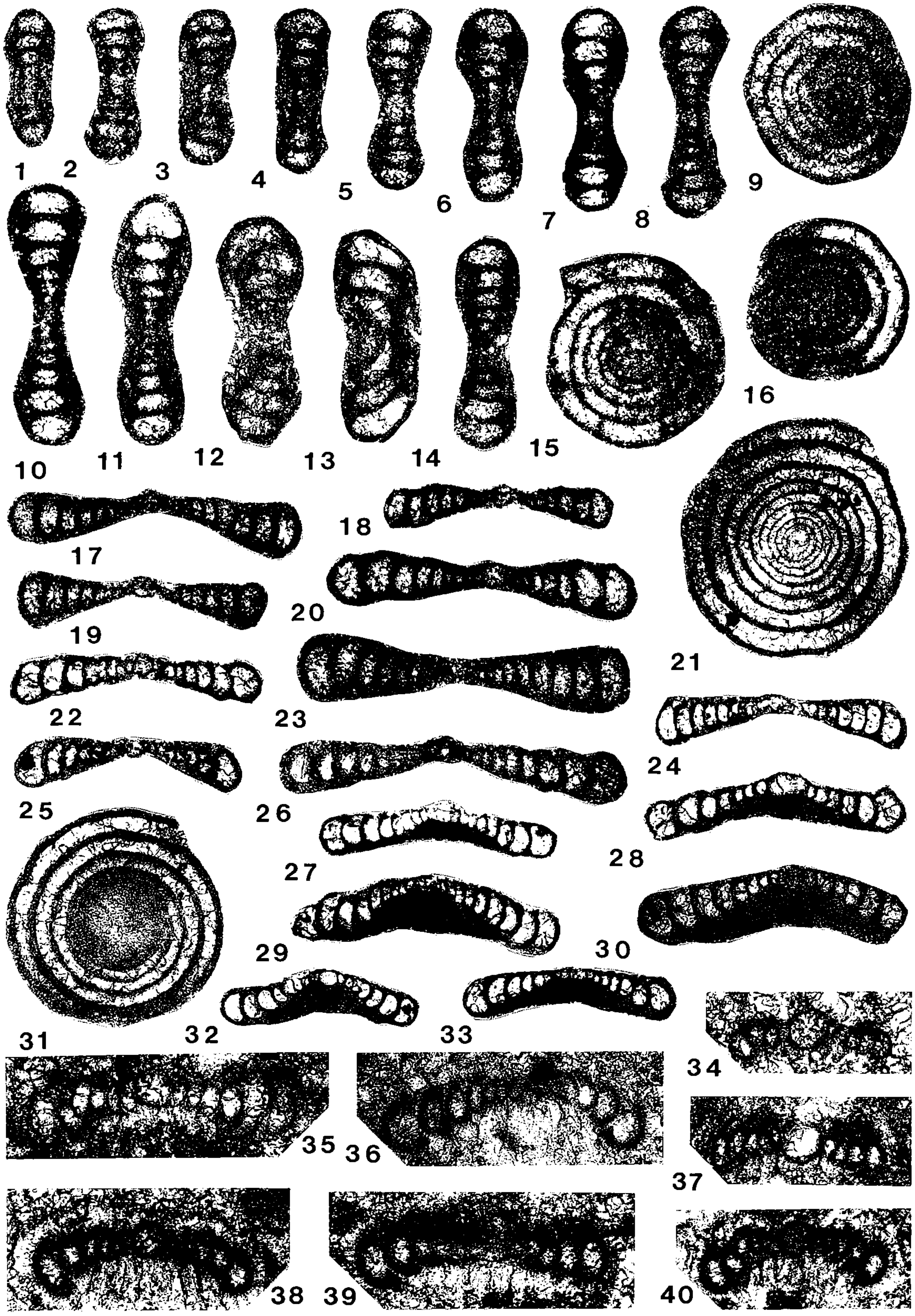
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PLATE 4

(All figures ×100)

Figures 1–7.—*Hemidiscus carnicus* Schellwien, 1898; 1 sagittal section; 2–7 axial and tangential axial sections. **1,** CR spl. 64 (OU 10768a). **2,** PB spl. 34 (OU 8770b). **3,** CR spl. 114 (OU 8629h). **4,** PB spl. 34 (OU 8770c). **5,** ZR spl. 80 (OU 10769a). **6,** ZR spl. 81 (OU 10770a). **7,** PB spl. 34 (OU 8770d).

Figures 8–13.—*Biseriella parva* (Chernysheva, 1948), all sagittal sections from CR. **8,** spl. 34 (OU 8793b). **9,** spl. 22 (OU 8637b). **10,** spl. 17 (OU 8313i). **11,** spl. 36 (OU 10771a). **12,** spl. 37 (OU 10772a). **13,** spl. 36 (OU 10771b).

Figures 14–19.—*Globivalvulina bulloides* (Brady, 1876); 14–18 sagittal sections; 19 oblique axial section. **14,** PB spl. 27 (OU 8297e). **15,** PB spl. 66 (OU 10773a). **16,** ZR spl. 35 (OU 8302c). **17,** PB spl. 7 (OU 8731b). **18,** ZR spl. 30 (OU 10760b). **19,** PB spl. 7 (OU 8731c).

Figures 20–24.—*Globivalvulina granulosa* Reitlinger, 1950, all variably oblique and tangential axial sections. **20,** CR spl. 18 (OU 10774a). **21,** CR spl. 106 (OU 8799b). **22,** CR spl. 99 (OU 8330e). **23,** CR spl. 18 (OU 10774b). **24,** PB spl. 19 (OU 8646b).


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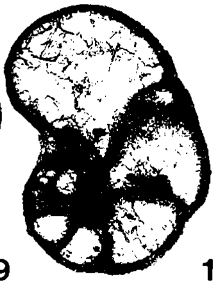
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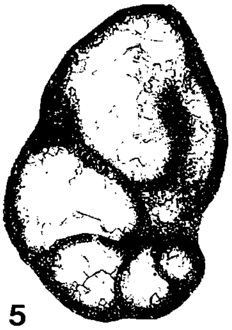
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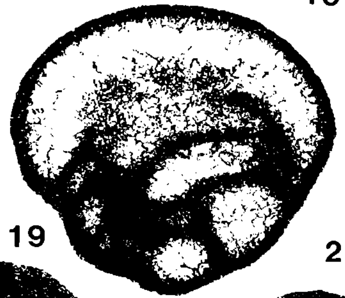
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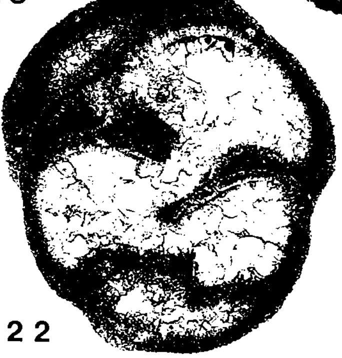
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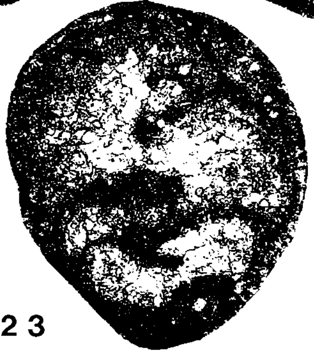
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PLATE 5

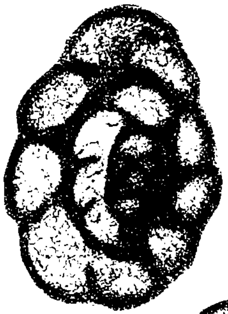
Figures 1,4,7–9.—*Planoendothyra* spp., ×100; 1 sagittal section; 4,7–9 axial and tangential axial sections. **1,** PB spl. 8 (OU 8298c). **4,** PB spl. 44 (OU 8306d). **7,** ZR spl. 4 (OU 10775a). **8,** PB spl. 99 (OU 10776a). **9,** CR spl. 24 (OU 10777a).

Figures 2,5,6.—*Endothyranella* spp., ×100; all longitudinal sections. **2,** CR spl. 121 (OU 10778a). **5,** ZR spl. 69 (OU 8785c). **6,** ZR spl. 99 (OU 10779a).

Figures 3,10.—Asteroarchaediscidae (undifferentiated), ×200; both axial sections from CR. **3,** spl. 28 (OU 10763b). **10,** spl. 17 (OU 8313j).

Figures 11–17.—*Bradyina* spp., ×20. **11,** oblique section, ZR spl. 44 (OU 8300d). **12,** oblique section, CR spl. 99 (OU 8330f). **13,** axial section, PB spl. 10 (OU 10780a). **14,** axial section, CR spl. 104 (OU 8798b). **15,** sagittal section, CR spl. 100 (OU 10781a). **16,** sagittal section, ZR spl. 6 (OU 10782a). **17,** sagittal section, CR spl. 61 (OU 8712h).

Figures 18–21.— *Endothyra* spp., ×100; 18,20,21 axial sections; 19 sagittal section. **18,** ZR spl. 122 (OU 8316c). **19,** ZR spl. 3 (OU 8726b). **20,** CR spl. 38 (OU 10783a). **21,** ZR spl. 3 (OU 8726c).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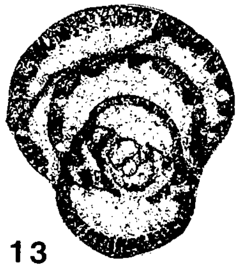
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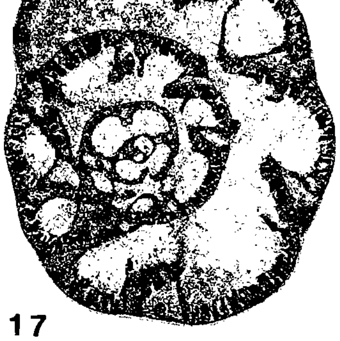
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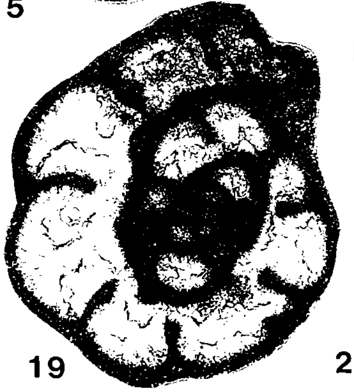
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Oklahoma Geological Survey Circular 94, 1992

Chaetetes (Demospongiae): Its Occurrence and Biostratigraphic Utility

R. R. West
Kansas State University

ABSTRACT.—The coralline demosponge *Chaetetes* has had a number of taxonomic "homes" and is widely distributed in Carboniferous rocks of the world. The ease of recognition and wide distribution of *Chaetetes* in North America has led to a widely held belief that it is confined to a specific subdivision of the Pennsylvanian in that region. Despite reports of *Chaetetes* from rocks of Missourian age and a reported range from ?Silurian to Permian (Hill, 1981), the view has persisted that *Chaetetes* is an index, guide, or zone fossil for the Lower Pennsylvanian (Morrowan, Atokan, and Desmoinesian).

Current knowledge of the spatial and temporal distribution of *Chaetetes* in North America supports recognition of an abundance biozone (as defined by the 1983 North American Stratigraphic Code); this biozone includes Atokan and Desmoinesian strata and neither requires nor implies extinction of the genus. Clear, concise, and current science requires clarification of the biostratigraphic utility of *Chaetetes* as an abundance biozone, not as a range zone or chronozone.

INTRODUCTION

Chaetetes has long been considered a useful biostratigraphic taxon for recognition of part of the Pennsylvanian System. Moore and others (1944, p. 670, 673) stated that "a number of fossils, . . . and *Chaetetes* among the corals, . . . are restricted to a certain subdivision" and that "the unconformity at the base of the Missourian series marks the upper limit of known occurrences of *Chaetetes* in the Pennsylvanian of North America." Moore and others (1944, p. 673) also suggested that Desmoinesian species of *Chaetetes* are different from those of the Morrowan and Atokan (Lampasan) by stating that "colonies of *Chaetetes* that are common in some Morrowan and Lampasan rocks of Arkansas, Oklahoma, and Texas are specifically distinct from the so-called *C. milleporaceus*, which is very common in Desmoinesian deposits."

The view that *Chaetetes* is an index, guide, or zone fossil for the Lower Pennsylvanian (Morrowan, Atokan, and Desmoinesian) and that it became extinct at the end of the Desmoinesian has persisted (Heckel, 1988), despite the fact that Missourian occurrences of this genus have been reported (Bassler, 1950; Shelton, 1953; Lane, 1962; Brown and Wermund, 1969; Sando, 1985; and West, 1988). Indeed, Hill (1981, p. F508) gave the range of *Chaetetes* as ?Silurian to Permian.

Hill (1981) listed six Paleozoic families (four of these are queried) and two post-Paleozoic families within the Order Chaetetida. The combined range of the six Paleozoic families (i.e., Paleozoic chaetetids) is given by Hill (1981) as Ordovician to Upper Permian, nearly the same as that of the genus *Chaetetes*. Although Hill (1981, p. F519) recognized six post-Paleozoic genera (three in each of the two families, Chaetetidae and Acanthochaetetidae) that "may be reasonably referred to the Chaetetida," she separated the two groups stratigraphically. This may have been done to preserve the Tabulata as a Paleozoic subclass because Hill listed the range for the Tabulata (p. F506) as Ordovician to Permian. Actually, the range of the Order Chaetetida would have been more properly Ordovician to Recent, because Hartman and Goreau (1975) described *Acanthochaetetes wellsi* from the western Pacific, and *Acanthochaetetes* is one of the three genera Hill included in the Family Acanthochaetetidae. Apparently, Hill was unaware of the 1975 paper by Hartman and Goreau, because it is not cited in her 1981 study.

The systematics of the Chaetetida has been in a state of flux, since the recognition of extant sclerosponges. Indeed, chaetetids, stromatoporoids, and other groups have more in common with the Porifera than with the Coelenterata. Readers interested in these problems should consult Wood and others (1989).

Acknowledgments

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BASIC BIOSTRATIGRAPHIC UNITS

The 1983 edition of the North American Stratigraphic Code recognizes three principal kinds of biostratigraphic units (p. 862): interval, assemblage, and abundance biozones. An interval biozone is the body of strata between the lowest and highest documented occurrence of a single taxon. An assemblage biozone is a biozone identified on the basis of three or more taxa. An abundance biozone is characterized by quantitatively distinctive maxima of the relative abundance of one or more taxa. Three types of interval biozones are recognized: taxon range zone, concurrent range zone (or a partial range zone if used to partition the range of a third taxon), and lineage zone. There are two types of assemblage zones (see North American Stratigraphic Code, 1983, p. 863).

To establish the biostratigraphic "value" of *Chaetetes*, we need to examine the stratigraphic range and abundance of this genus and its association with other taxa. It will then be possible to determine which, if any, of the three biozones might be applicable for this genus.

RANGE AND ABUNDANCE OF *CHAETETES*

General

Chaetetes, a manuscript name by Fischer von Waldheim, was included in a work by Eichwald (1829, p. 197) resulting in some confusion as to the genotype, but the matter was clarified by Oakley's (1936) subsequent designation of *Chaetetes cylindraceus* as genotype.

Many species originally assigned to *Chaetetes* have been shown to be Bryozoa by further study and more modern techniques. This can be seen easily by comparing any late-19th-century list of Paleozoic fossils with a similar list from the latter part of the 20th century. Additionally, a large number of the 19th- and early-20th-century species came from rocks of Ordovician age (see Bassler, 1950). As Ordovician species became bryozoans, the stratigraphic range of *Chaetetes* shrank, and the present range is given (Hill, 1981) as ?Silurian to Permian. In contrast, Kato and Kamada (1977) reported a Triassic occurrence of *Chaetetes* in the Kitakami Mountains of Japan.

Current studies of species of *Chaetetes* suggest that further revision may be required. According to West (1989a,b), the seven species of Carboniferous *Chaetetes* in North America show few morphological differences; i.e., the characters currently used to speciate them are not accurate discriminators of species.

Most of what follows deals with the Carboniferous of North America, but it is also necessary to summarize briefly the temporal and spatial range and abundance of *Chaetetes* in pre- and post-Carboniferous rocks to put the Carboniferous forms in proper perspective.

Pre-Carboniferous

As noted earlier in this paper, species of *Chaetetes* from Ordovician rocks proved to be bryozoans, and I know of no valid Ordovician species or occurrences of this genus.

A specimen (from Carl Brett) referred to "*Chaetetes*" *lycoperdon* occurs in the Middle Silurian Clinton Group of New York. The specimen is "essentially identical" to specimens identified as chaetetids by Gillette (1947) (Brett, personal communication). The specimen is a tabular colony ~5 mm thick having an area of ~80 mm^2. Preliminary examination of an etched surface indicates that the calicles are small and the walls and tabulae are thin. Other specimens from the Late Silurian (Keyser Formation) in Pennsylvania are rather small domical to columnar colonies (2–2.5 cm in diameter and 2–4 cm high). Thus, I am satisfied, pending further study, that the genus *Chaetetes* is present in rocks of Silurian age.

Don Kissling provided me with illustrations of Middle Devonian specimens of *Chaetetes* from cores of the Winnipegosis Formation in Saskatchewan, Canada. These specimens are very well preserved, high domical colonies associated with stromatolite boundstones and peloid packstones/grainstones in a stromatolite crest facies of pinnacle reefs. These specimens are believed to represent valid occurrences of *Chaetetes* in the Devonian.

Carboniferous–Mississippian

Most occurrences of *Chaetetes* in the United Kingdom and Europe are in rocks of Lower Carboniferous age. Occurrences in reef mounds of middle and upper Viséan age in the United Kingdom are reported by Wolfenden (1958) and Jameson (1980,1987), and I have examined numerous other specimens of *Chaetetes depressus*, *C. radians*, and *C. septosus* from the Lower Carboniferous of England, Scotland, Wales, and the European continent.

Numerous occurrences of *Chaetetes* have been reported in the Carboniferous of China (Lin and others, 1988; X. Zhang, personal communication, 1989) and Japan (Minato and Kato, 1974). A particularly well-known occurrence in Japan is in the reef carbonates of the Akiyoshi Limestone (Ota, 1968, 1977), where *Chaetetes* occurs in rocks that range from Namurian A (middle Chesterian) to Westphalian C (lower Desmoinesian) in age. *Chaetetes cylindricus* has been reported from the Lower Carboniferous of Peru (Bassler, 1950).

The only described species of *Chaetetes* known to me from the Mississippian of North America is *C. wyomingensis* Sando (1975). This species was recorded in the *Caninia* Zone of the Moffat Trail Member of the Amsden Formation in western Wyoming by Sando (1975), now known as Zone V in the western interior coral zonation of Sando and Bamber (1985). Other possible Mississippian occurrences of the genus were reported by Stouder (1938), Duncan (1965,1966), and Gutschick (1965) from the Lower Kinkaid Limestone of western Kentucky and eastern Illinois. Trace and McGrain (1985) stated that Duncan (personal communication to them, 1964) identified the chaetetid in these rocks as *Chaetetella*, but Gutschick, who corresponded with Duncan on several occasions, called specimens from this area and stratigraphic unit *Chaetetes* in his 1965 paper. My examination of Gutschick's collections supports assignment to *Chaetetes*. The occurrence of this chaetetid is restricted to the Lower Kinkaid in this area, is useful in mapping (Trace and McGrain, 1985), and thus has local biostratigraphic significance.

Other Mississippian occurrences are in Arrow Canyon, Nevada, where I have collected *Chaetetes* from several intervals assigned to the Mississippian by foraminifers and conodonts (P. Brenckle and R. Lane, personal communication). I also have specimens of *Chaetetes* from the Mississippian Monte Cristo Limestone in Nevada and from talus below the lower part of the Great Blue Formation (Meramecian/Chesterian) from the Wellsville Mountains, Utah.

Carboniferous–Pennsylvanian

Morrowan

Morrowan occurrences of *Chaetetes* in North America are: Bird Springs Formation of Nevada (possible chaetetids occur in 22 horizons in Arrow Canyon, dated as Morrowan, Atokan, and Desmoinesian by foraminifers and conodonts, P. Brenckle and R. Lane, personal communication); Braggs Member of the Sausbee Formation in Oklahoma and Arkansas (Sutherland and Henry, 1977; Sutherland and Manger, 1977,1979); La Tuna Formation in West Texas (Connolly, 1985; Connolly and others, 1989). The occurrence in the Braggs Member of the Sausbee Formation also has local biostratigraphic utility (Sutherland and Henry, 1977; Sutherland and Manger, 1977,1979).

Atokan

Numerous occurrences of *Chaetetes* are reported (Einor and others, 1984), particularly from the Ural Mountains, and I have collected a single, small, domical colony from the Lower Moscovian at the Kilim River section.

Chaetetes is rather conspicuous in rocks of Atokan age in North America. These occurrences are commonly as biostromes and small bioherms, particularly in the Great basin, and especially in Nevada (see West, 1988, for references to Atokan occurrences of *Chaetetes*) and have some local biostratigraphic significance there. A *Chaetetes-Profusulinella* faunizone is reported by Dott (1954, 1955) in northeastern Nevada, where most of the occurrences of *Chaetetes* are persistent biostromes. Using data from Dott (1955), West (1988, p. 156) calculated that some of these biostromes cover as much as 4,500 km^2. Mollazal (1961), Bissell (1962, 1964), and Lane (1962) also recognized a *Chaetetes–Profusulinella* (Lower Atokan) association in east-central Nevada and eastern and western Utah, respectively. However, Webster (1969) found no relationship between *Chaetetes* and *Profusulinella* in southern Nevada, but reported an Upper Atokan *Fusulinella–Chaetetes* faunal zone. In the Akiyoshi Limestone Group of Japan (Ota, 1968) and in the Llano area of Texas (Winston, 1963), chaetetid reef mounds occur in rocks containing both *Profusulinella* and *Fusulinella*.

Unfortunately, for detailed regional biostratigraphy, the number of chaetetid biostromes varies from section to section in the Great basin. Nelson and Langenheim (1980) reported three such stratigraphic occurrences in Arrow Canyon, four at Dry Canyon, and eight in the Las Vegas Range. Webster (1969) reported from three to five in southern Nevada, and Brill (1963) noted five *Chaetetes* zones in the Spring Mountains. My own work (unpublished) in Arrow Canyon records 22 beds of possible chaetetids from the Morrowan through the Desmoinesian. Of importance to biostratigraphy are the observations by Rich (1969, p. 356), who noted that *Chaetetes*-rich zones are not consistent between stratigraphic sections, that a number of different zones occur at different stratigraphic positions at a number of different locations, and that some cyclicity is apparent. On the other hand, Sando (1985, p. 348) included *Chaetetes* as one of the 14 genera characteristic of the Atokan in the Western Interior Region of the United States.

The Marble Falls Formation in the Mason area of central Texas contains some impressive chaetetid colonies (Winston, 1963; Sutherland, 1984). Winston (1963) recorded colonies >3 m tall and 1 m in diameter. Chaetetid biostromes are key beds that mark boundaries between cyclic units in the same area (Winston, 1963, p. 31).

In West Texas (El Paso area), chaetetid biostromes are reported from the Atokan part of the Magdalena Formation (Lambert and Stanton, 1986; Lambert and others, 1986; Lambert, 1989). According to Lambert (1989, p. 117–118), these biostromes were noted by Beede (1918), who recorded their occurrence in a prominent stratum 3–8 ft thick that was traceable for 3–4 mi.

Other occurrences in the Atokan of the southwestern United States are in Arizona (Reid, 1968;

Sando, 1985), New Mexico (Wilson and others, 1969; Sando, 1985), and western Colorado (Sando, 1985; Connolly, 1990). Wilson and others (1969) also noted chaetetids in Atokan to Desmoinesian rocks (Horquilla limestone) in the Palomas Mountains, Chihauhua, Mexico. Sando (1985) reported *Chaetetes* from the Keeler Canyon Formation, the lower part of which is thought to be Atokan or Desmoinesian, in the Panamint Range of California (Hall and MacKevett, 1958). An early Moscovian, probably Atokan, occurrence (very similar to Desmoinesian occurrences in southeastern Kansas) is reported by Stemmerik (1989) from the Kap Jungersen Formation in Holm Land of eastern North Greenland.

Desmoinesian

Chaetetid occurrences in the Desmoinesian, particularly in North America, are more widespread than in any of the other subdivisions of the Pennsylvanian. They are reported from all the areas noted above for the Atokan (Arizona, Utah, Nevada, New Mexico, Colorado, and the surface and subsurface of Texas), as well as from Oklahoma, Kansas, Missouri, Iowa, Illinois, Indiana, and Wyoming. Moore and others (1944, p. 701) also reported *Chaetetes milleporaceus* from South Dakota and Idaho; Vanderwilt (1935) reported this species from the Hermosa of Colorado; and one of the two type localities for this species is Cumberland Mountain in Tennessee. However, there is some doubt as to the validity of this latter occurrence (see Lane and Martin, 1966). The details of some of these occurrences were considered by West (1988). Willman and others (1975, p. 183) stated that *Chaetetes* is seldom found above the top of the Desmoinesian in Illinois. Reported occurrences in western Illinois (Wanless, 1958) may be spurious (Weibel, personal communication, 1990). According to Lane and Martin (1966), at least two Desmoinesian horizons contain chaetetids in Indiana.

In the subsurface of Texas, Oklahoma, eastern Colorado, and possibly western Kansas, chaetetid-bearing units are known to produce hydrocarbons. Some chaetetid occurrences in southeastern Kansas are stained black and almost "drip" oil. Occurrences in the Honaker Trail Formation of the Paradox basin are also associated with hydrocarbon production. Chaetetids are associated with three levels of bioherm development in the Paradox basin (Wengerd, 1951,1955,1963), but the rather widespread biostromes noted in the Atokan are apparently absent in that area. In North America, the biomass of chaetetids appears to have been greatest during the Desmoinesian (West, 1988).

Missourian

There is a distinct decrease in the abundance of *Chaetetes* in rocks of Missourian age, but the genus does not become extinct in the Missourian. It has been reported from the Graford Group (Bassler, 1950) and the Winchell Limestone in north-central Texas (Shelton, 1953; Brown and Wermund, 1969; West, 1988). Recently, I received chaetetids collected by D. Boardman from yet another Winchell Limestone locality in this area. Watney and others (1989, p. 96) reported chaetetid-like masses from the Argentine Limestone Member in east-central Kansas, and my collections from this locality suggest that these specimens are very similar to those from the Winchell Limestone of Texas. Other Missourian occurrences in North America are reported by Lane (1962) from the Ely Group in Nevada, by Sando (1985) from Utah, and by Bassler (1950) from Oklahoma and Missouri.

Virgilian

I know of no reported occurrences of *Chaetetes* from the Virgilian in North America. Indeed, few chaetetid occurrences are reported from the Upper Stephanian (Virgilian) in the world. Bassler (1950, p. 234–235) reported *Chaetetes cylindraceus radians* from the Uralian of Spitsburgen.

Post-Carboniferous

Hill (1981, p. F508) indicated Permian occurrences of *Chaetetes* from North America and Asia. Asian occurrences are reported by Bassler (1950) and Branson (1948) from China and Japan. Reinhardt (1988) noted chaetetids from Permian reefs in China. To date, I have been unable to locate Hill's source for the North American occurrences, but, given the Asian occurrence, I suspect these North American occurrences might be in the accreted terrains along the Pacific coast of the continent. *Chaetetes tabuliporoides* was described by Termier and others (1977) from Permian rocks of southern Tunisia. As noted earlier, Kato and Kamada (1977) recorded *Chaetetes* cf. *C. tenuiradiatus* from the Triassic of Japan. (*Chaetetes tenuiradiatus* was first described by Sokolov [1950] from the Viséan of the Moscow basin). Hagemeister (1988) noted chaetetids in the Triassic (Carnian) of Austria. Specimens collected by P. Reid from the Triassic of the Yukon appear, on preliminary examination, to be *Chaetetes*.

SUMMARY AND CONCLUSIONS

This current study indicates clearly that the genus *Chaetetes* does not become extinct at the Desmoinesian–Missourian boundary in North America or elsewhere. Indeed, the interval biozone of this genus is Silurian to Triassic. Inasmuch as three or more taxa are required for an assemblage biozone, that concept is also inappropriate. Thus, we are left with the abundance biozone (acme zone of the International Subcommission on Stratigraphic Classification, 1976, p. 59) concept. If we use an estimate of the biomass of *Chaetetes* as an indication of abundance (West, 1988), the abundance biozone of *Chaetetes* includes rocks of Atokan and Desmoinesian age in North America.

As noted earlier, Moore and others (1944, p. 673) suggested that the Desmoinesian species (*Chaetetes milleporaceus*) is different from Morrowan and Atokan species such as *Chaetetes favosus*, *C. eximus*, and *C. subtilis*. Moreover, Nelson and Langenheim (1980, p. 15) in their studies of *Chaetetes milleporaceus* and *C. favosus* in Nevada stated "that the two species perhaps are ecotypes rather than Mendelian populations." Current studies (West, 1989a,b) suggest that criteria used to differentiate species of *Chaetetes* are in need of revision; those studies are progressing.

REFERENCES CITED

Bassler, R. S., 1950, Faunal lists and descriptions of Paleozoic corals: Geological Society of America Memoir 44, 315 p.

Beede, J. W., 1918, Notes on the geology and oil possibilities of the northern Diablo Plateau in Texas: University of Texas Bulletin 1852, 37 p.

Bissell, H. J., 1962, Pennsylvanian–Permian Oquirrh basin of Utah: Brigham Young University Geology Studies, v. 9, pt. 1, p. 26–49.

______1964, Ely, Arcturus, and Park City Groups (Pennsylvanian–Permian) in eastern Nevada and western Utah: American Association of Petroleum Geologists Bulletin, v. 48, p. 565–636.

Branson, C. C., 1948, Bibliographic index of Permian invertebrates: Geological Society of America Memoir 26, 1049 p.

Brill, K. G., 1963, Permo–Pennsylvanian stratigraphy of western Colorado Plateau and eastern Great basin regions: Geological Society of America Bulletin, v. 74, p. 307–331.

Brown, L. F., Jr.; and Wermund, E. G. (eds.), 1969, A guidebook to the Late Pennsylvanian shelf sediments, north-central Texas: Dallas Geological Society, 69 p.

Connolly, M. W., 1985, Microfacies analysis, paleoecology, and environment of deposition of Morrowan shelf carbonates, Magdalena Limestone (lower division), Hueco Mountains, El Paso County, West Texas: Texas A&M University unpublished M.S. thesis, 437 p.

______1990, Paleobiogeography of *Chaetetes* (Chesterian–Missourian) in North America [abstract]: Geological Society of America Abstracts with Programs, v. 22, p. 9.

Connolly, M. W.; Lambert, L. L.; and Stanton, R. J., Jr., 1989, Paleoecology of Lower and Middle Pennsylvanian (Middle Carboniferous) *Chaetetes* in North America: Facies, v. 20, p. 139–168.

Dott, R. H., Jr., 1954, *Chaetetes*, important marker in Pennsylvanian of central Great basin [abstract]: Geological Society of America Bulletin, v. 65, p. 1245–1246.

______1955, Pennsylvanian stratigraphy of Elko and northern Diamond ranges, north-eastern Nevada: American Association of Petroleum Geologists Bulletin, v. 39, p. 2211–2305.

Duncan, H. M., 1965, Mississippian chaetetid from Kentucky, *in* Geological Survey Research in 1965: U.S. Geological Survey Professional Paper 525-A, p. A122.

______1966, Mississippian occurrence of *Chaetetes*, *in* Geological Survey Research 1966: U.S. Geological Survey Professional Paper 550-A, p. A112.

Eichwald, C. E. d', 1829, Zoologia specialis quam expositis animalibus tum vivis, tum fossilibus potissimum Rossiae in universum, et Poloniae in specie, in usum lectionum, J. Zawalski (Vilna): v. 1, 314 p.

Einor, O. L.; and others (eds.), 1984, Upper Paleozoic of Southern Urals: Excursion 047 of 27th International Geological Congress, Moscow, 135 p.

Gillette, T., 1947, The Clinton of western and central New York: New York State Museum Bulletin 341, 191 p.

Gutschick, R. C., 1965, *Pterotocrinus* from the Kinkaid Limestone (Chester, Mississippian) of Illinois and Kentucky: Journal of Paleontology, v. 39, p. 636–646.

Hagemeister, A., 1988, Kylische sedimentation auf einer stabilen Karbonatplattform: Die Raibler Schichten (Karn) des Drauzuges/Karten (Osterreich): Facies, v. 18, p. 83–122.

Hall, W. E.; and MacKevett, E. M., 1958, Economic geology of the Darwin Quadrangle, Inyo County, California: California Department of Natural Resources, Division of Mines, Special Report 51, 73 p.

Hartman, W. D.; and Goreau, T. F., 1975, A Pacific tabulate sponge, living representative of a new order of sclerosponges: Postilla, v. 167, 21 p.

Heckel, P. H., 1988, Preliminary conodont biostratigraphic characterization of Middle–Upper Pennsylvanian eustatic cyclic sequence in Mid-continent North America [abstract]: Geological Society of America Abstracts with Programs, v. 20, p. 347.

Hill, D., 1981, Rugosa and Tabulata, *in* Teichert, C. (ed.), Treatise on invertebrate paleontology, Part F, Coelenterata, Supplement 1: University of Kansas Press, Lawrence, p. F1–F762.

International Subcommission on Stratigraphic Classification, 1976, International stratigraphic guide: John Wiley and Sons, New York, 200 p.

Jameson, J., 1980, Depositional environments in the Petershill Formation, Bathgate, West Lothian: University of Edinburgh unpublished Ph.D. dissertation, 545 p.

______1987, Carbonate sedimentation on a mid-basin high: the Petershill Formation, Midland Valley of Scotland, *in* Miller, J.; Adams, A. E.; and Wright, V. P. (eds.), European Dinantian environments: John Wiley and Sons, New York, p. 309–327.

Kato, M.; and Kamada, K., 1977, On the occurrence of *Chaetetes* from the Triassic Inai Group in the Karakuwa Peninsula, southern Kitakami Mountains, Japan [in Japanese]: Journal of the Geological Society of Japan, v. 83, p. 250–251.

Lambert, L. L., 1989, Carbonate facies and biostratigraphy of the Middle Magdalena, Heuco Mountains, West Texas: Texas A&M University unpublished M.S. thesis, 280 p.

Lambert, L. L.; and Stanton, R. J., Jr., 1986, Environmental controls on *Chaetetes* growth morphology [abstract]: Geological Society of America Abstracts with Programs, v. 18, p. 665.

Lambert, L. L.; Connolly, W. M.; and Stanton, R. J., Jr., 1986, Preferred depth of abundant Pennsylvanian *Chaetetes* [abstract]: Fourth North American Paleontological Convention Abstracts, p. A27.

Lane, B. O., 1962, The fauna of the Ely Group in the Illipah area of Nevada: Journal of Paleontology, v. 36, p. 888–911.

Lane, N. G.; and Martin, R. A., 1966, Redescription of *Chaetetes milleporaceous* Milne-Edward & Haime, 1851: Journal of Paleontology, v. 40, p. 651–654.

Lin, Baoyu; and others, 1988, Monograph of Palaeozoic corals: tabulatomorphic corals: Geological Publishing House, Beijing, China, v. 1, 467 p., v. 2, 493 p.

Minato, M.; and Kato, M., 1974, Upper Carboniferous corals from the Nagaiwa Series, southern Kitakami Mountains, N.E. Japan: Journal of the Faculty of Science Hokkaido University, Series IV, Geology and Mineralogy, v. 16, p. 43–119.

Mollazal, Y., 1961, Petrology and petrography of Ely Limestone in part of eastern Great basin: Brigham Young University Geology Studies, v. 8, p. 3–35.

Moore, R. C.; and others, 1944, Correlation of Pennsylvanian formations of North America: Geological Society of America Bulletin, v. 55, p. 657–706.

Nelson, W. J.; and Langenheim, R. L., Jr., 1980, Ecological observation on *Chaetetes* in southern Nevada: Pacific Geology, v. 14, p. 1–22.

North American Commission on Stratigraphic Nomenclature, 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841–875.

Oakley, K. P., 1936, An Ordovician species of *Chaetetes*: Geological Magazine, v. 73, p. 440–444.

Ota, M., 1968, The Akiyoshi Limestone Group: a geosynclinal organic reef complex: Bulletin Akiyoshi-dai Science Museum No. 5, 44 p.

______1977, General geology of the Akiyoshi Limestone Group, *part 1 of* Geological studies of Akiyoshi: Bulletin Akiyoshi-dai Science Museum No. 12, 33 p.

Reid, A. M., 1968, Biostratigraphy of the Naco Formation (Pennslyvanian) in south-central Arizona: University of Arizona unpublished Ph.D. dissertation, Tucson, 299 p.

Reinhardt, J. W., 1988, Uppermost Permian reefs and Permo–Triassic sedimentary facies from the southeastern margin of Sichuan basin, China: Facies, v. 18, p. 231–288.

Rich, M., 1969, Petrographic analysis of Atokan carbonate rocks in central and southern Great basin: American Association of Petroleum Geologists Bulletin, v. 53, p. 340–366.

Sando, W. J., 1975, Coelenterata of the Amsden Formation (Mississippian and Pennsylvanian) of Wyoming: U.S. Geological Survey Professional Paper 848-C, 30 p.

______1985, Biostratigraphy of Pennsylvanian (Upper Carboniferous) corals, Western Interior Region, conterminous USA: Compte Rendu, Dixieme Congres International de Stratigraphie et de Geologie du Carbonifere, v. 2, p. 335–350.

Sando, W. J.; and Bamber, E. W., 1985, Coral zonation of the Mississippian System in the Western Interior Province of North America: U.S. Geological Survey Professional Paper 1334, 61 p.

Shelton, J. W., 1953, Studies of the Strawn–Canyon boundary, Pennsylvanian, in north-central Texas: University of Illinois unpublished Ph.D. dissertation, 120 p.

Sokolov, B. S., 1950, Carboniferous Chaetetidae of the northeast Ukraine and adjacent areas: Trudy vses. neft. naucho-issled. geol.-razv. Inst. N. S., v. 27, 101 p.

Stemmerik, L., 1989, Chaetetid bioherm, Upper Carboniferous, Holm Land, eastern north Greenland, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent areas: Canadian Society of Petroleum Geologists Memoir 13, p. 688–689.

Stouder, R. E., 1938, Chester rocks of Meade, Hardin, and Breckenridge Counties, Kentucky: American Association of Petroleum Geologists Bulletin, v. 22, p. 267–284.

Sutherland, P. K., 1984, *Chaetetes* reefs of exceptional size in Marble Falls Limestone (Pennsylvanian), central Texas: Fourth International Symposium of Fossil Cnidaria, Paleontographica Americana No. 54, p. 543–547.

Sutherland, P. K.; and Henry, T. W., 1977, Carbonate platform facies and new stratigraphic nomenclature of the Morrowan Series (Lower and Middle Pennsylvanian), northeastern Oklahoma: Geological Society America Bulletin, v. 88, p. 425–440.

Sutherland, P. K.; and Manger, W. L. (eds.), 1977, Upper Chesterian–Morrowan stratigraphy and the Mississippian–Pennsylvanian boundary in northeastern Oklahoma and northwestern Arkansas: Oklahoma Geological Survey Guidebook 18, 185 p.

______(eds.), 1979, Mississippian–Pennsylvanian shelf-to-basin transition, Ozark and Ouachita regions, Oklahoma and Arkansas: Oklahoma Geological Survey Guidebook 19, 81 p.

Trace, R. D.; and McGrain, P., 1985, The *Chaetetella* Zone in the Kinkaid Limestone (Mississippian): a useful stratigraphic marker along the southern rim of the Eastern Interior (Illinois) basin: Kentucky Geological Survey Information Circular 14, series 11, 9 p.

Termier, H.; Termier, G.; and Vachard, D., 1977, Monographie paleontologique des affleurements Permians du Djebel Tebaga (Sud Tunisien): Paleontographica A, v. 156, p. 1–109.

Vanderwilt, J. W., 1935, Stratigraphy of Pennsylvanian Hermosa Formation in Elk Mountains, Gunnison County, Colorado: American Association of Petroleum Geologists Bulletin, v. 19, p. 1668–1677.

Watney, W. L.; French, J. A.; and Franseen, E., 1989, Sequence stratigraphic interpretations and modeling of cyclothems: Kansas Geological Survey Guidebook, 211 p.

Wanless, H. R., 1958, Pennsylvanian faunas of the Beardstone, Glasford, Havana, and Vermont Quadrangles: Illinois State Geological Survey Report of Investigations 205, 59 p.

Webster, G. D., 1969, Chester through Derry conodonts and stratigraphy of northern Clark and southern Lincoln Counties, Nevada: University of California

Publication in Geological Sciences, v. 79, p. 1–121.

Wengerd, S. A., 1951, Reef limestones of Hermosa Formation, San Juan Canyon, Utah: American Association of Petroleum Geologists Bulletin, v. 35, p. 1038–1051.

______1955, Biohermal trends in Pennsylvanian strata of San Juan Canyon, Utah, *in* Geology of parts of the Paradox, Black Mesa, and San Juan basins: Four Corners Geological Society Guidebook, First Field Conference, p. 70–77.

______1963, Stratigraphic section at Honaker Trail, San Juan Canyon, San Juan County, Utah, *in* Bass, R. O.; and Sharp, S. L. (eds.), Shelf carbonates of Paradox basin: Four Corners Geological Society, Fourth Field Conference Symposium, p. 235–243.

West, R. R., 1988, Temporal changes in Carboniferous reef mound communities: PALAIOS, v. 3, p. 152–169.

______1989a, Carboniferous species of *Chaetetes* in North America [abstract]: Geological Society of America Abstracts with Programs, v. 21, p. 43.

______1989b, Variability in *Chaetetes* (Carboniferous of North America), is it environmental? [abstract]: Geological Society of America Abstracts with Programs, v. 21, p. A253.

Willman, H. B.; and others, 1975, Handbook of Illinois stratigraphy: Illinois State Geological Survey Bulletin 95, 261 p.

Wilson, J. L.; Madrid-Solis, A.; and Malpica-Cruz, R., 1969, Microfacies analysis of Pennsylvanian and Wolfcampian strata in southwestern U.S.A. and Chihuahua, Mexico, *in* Cordoba, D. A.; Wengerd, S. A.; and Shoemaker, J. (eds.), The Border region: New Mexico Geological Society Field Conference 20, p. 80–90.

Winston, D., II, 1963, Stratigraphy and carbonate petrology of the Marble Falls Formation, Mason and Kimble Counties, Texas: University of Texas, Austin, unpublished Ph.D. dissertation, 344 p.

Wolfenden, E. B., 1958, Paleoecology of the Carboniferous reef complex and shelf limestones in northwest Derbyshire, England: Geological Society of America Bulletin, v. 69, p. 871–898.

Wood, R.; Reitner, J.; and West, R. R., 1989, Systematic and phylogenetic implications of the haplosclerid stromatoporoid *Newellia mira* nov. gen.: Lethaia, v. 22, p. 85–93.

ADDED NOTE: It is now clear that what have been previously considered chaetetids, stromatoporoids, sclerosponges, and other demosponges are polyphyletic groups (see Wood, 1990), and existing classifications are currently being revised. Until all existing genera and species of chaetetids and stromatoporoids, and others, are carefully restudied, it is only appropriate to refer to them as possessing chaetetid-grade or stromatoporoid-grade skeletons. Thus, it is inappropriate, in most cases, to refer to the genus *Chaetetes*. The phrase chaetetid-grade skeleton is more appropriate and accurate.

This new information has some potential biostratigraphic implications. It now appears that demosponges with a chaetetid-grade skeleton disappear at the Desmoinesian–Missourian boundary, at least in North America. Careful examination of all specimens referred to as chaetetids from the Missourian of North America by West (1988) and Watney and others (1989), and specimens supplied by Boardman (see p. 166 of this paper), revealed a stromatoporoid-grade skeleton like that described by *Newellia mira* (Wood and others, 1989). Thus, the top of the acme zone (end of the Desmoinesian) of this paper may correspond with the demise of some coralline demosponges with a chaetetid-grade skeleton. However, there are chaetetid-grade demosponges in the Permian, Triassic, and Jurassic, as well as extant forms such as *Spirastrella* (*Acanthochaetetes*) and *Merlia*. Therefore, until all existing taxa are reevaluated, caution must be exercised in using chaetetid-grade demosponges for biostratigraphy.

Wood, R., 1990, Reef-building sponges: American Scientist, v. 78, p. 224–235.

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Upper Paleozoic Reefs and Their Biota in the Canadian Arctic Archipelago

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ABSTRACT.—Upper Carboniferous (Pennsylvanian) and Lower Permian shelf and slope carbonates and associated evaporites exposed on Ellesmere and Axel Heiberg Islands in the Sverdrup basin contain a succession of buildups characterized by a variety of reef-associated organisms. Most of the buildups are mechanically stabilized by pervasive submarine cements. The principal organic contributors to the buildups, from oldest to youngest, are:

1) Beresellid and donezellid tubular and branching algae that form an initially porous framework in Morrowan (lower Bashkirian) buildups enclosed by anhydrite in the Otto Fiord Formation;

2) Fenestellid bryozoans in Atokan (upper Bashkirian to lower Moscovian) reefs within the basal Hare Fiord Formation;

3) Phylloid algae in Desmoinesian to Wolfcampian (Moscovian to Sakmarian) buildups and banks; the most common genera are the codiacean algae *Eugonophyllum*, *Neoanchicodium*, and *Ivanovia*; the enigmatic organism *Palaeoaplysina* (possible hydrozoan, sponge, or algal affinities) is commonly associated with phylloid algae;

4) The encrusting organism *Tubiphytes* and ramose bryozoans in Lower Permian (Sakmarian) buildups on southwestern Ellesmere Island; and

5) Sponges and fenestellid bryozoans in uppermost Lower Permian (Artinskian) buildups on southwestern Ellesmere Island.

INTRODUCTION

Carbonate reefs and related submarine-cemented carbonate units have been found at more than five stratigraphic levels in the Upper Carboniferous (Pennsylvanian) and Lower Permian succession of the Sverdrup basin in the Canadian Arctic archipelago (Fig. 1; Davies and Nassichuk, 1986; Davies, Richards, Beauchamp, and Nassichuk, 1989). Each occurrence is characterized by the dominance of one or more biotic groups including fenestellid bryozoans, phylloid algae, and the enigmatic organism *Palaeoaplysina*. The paleogeographic and stratigraphic distribution of these organisms provides insights to evolutionary pathways and endemism.

No reefs are known to have been penetrated by the more than 35 exploratory wells that have been drilled into Carboniferous and Permian rocks in the Canadian Arctic archipelago. Upper Carboniferous and Lower Permian reefs in the Ural Mountains in the Soviet Union have a biotic composition closely comparable to those in Arctic Canada, and are important reservoirs. Moreover, recent discoveries of major oil and gas reserves in Upper Carboniferous and Lower Permian reef complexes in the North Caspian basin of the Soviet Union, between the Caspian Sea and the Ural Mountains (Shebaldin and others, 1988) have rekindled interest in their potential in the Canadian Arctic.

TECTONO-STRATIGRAPHIC FRAMEWORK OF THE SVERDRUP BASIN

The Sverdrup basin is a pericratonic successor basin underlying the northern islands of the Canadian Arctic archipelago. It formed by rifting and collapse of Precambrian to Devonian basement rocks contained within the Franklinian Mobile Belt, which experienced several episodes of Paleozoic tectonism, the latest being the Ellesmerian orogeny from Middle Devonian to Early Carboniferous (Mississippian) time (Thorsteinsson, 1974). A diverse succession of Lower Carboniferous to lower Tertiary siliciclastics, carbonates, and evaporites exceeding 14 km accumulated within the Sverdrup basin. The entire sequence was deformed by the Eurekan orogeny in Tertiary time.

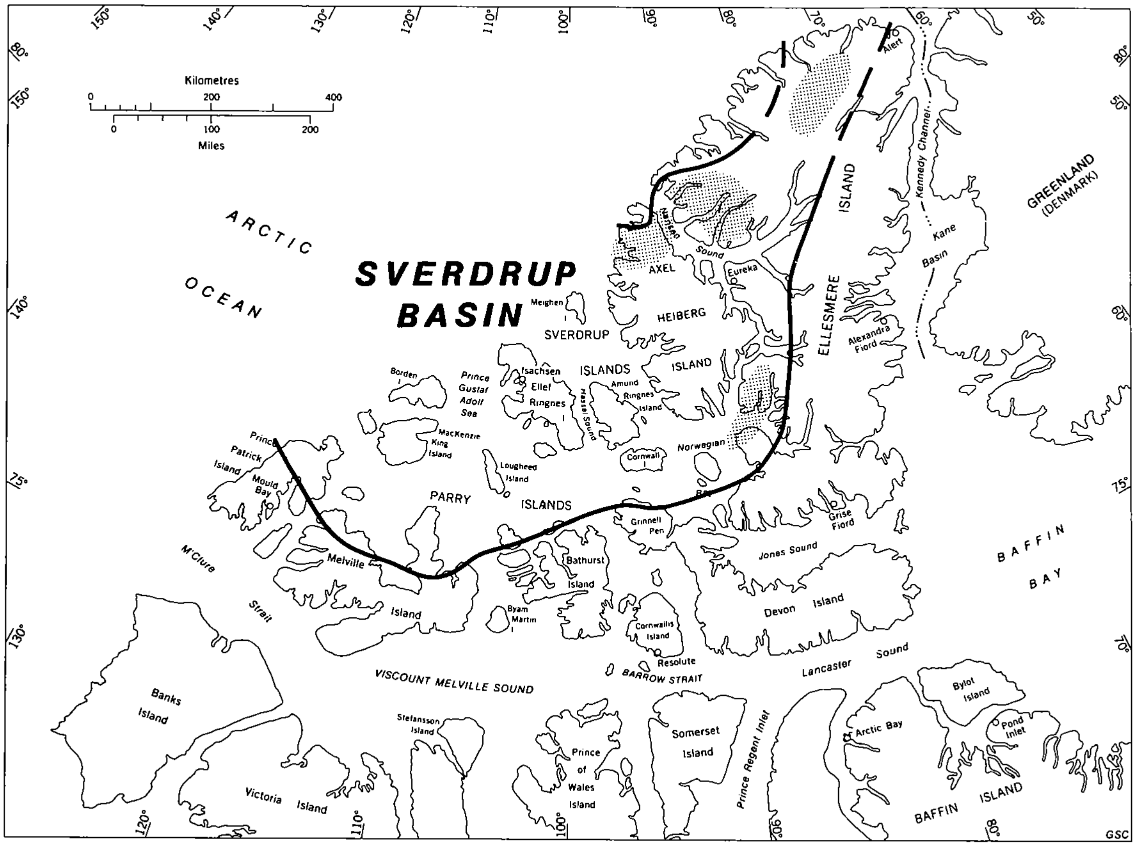

Figure 1. Location map of the Canadian Arctic archipelago showing outline of the Sverdrup basin and areas (shaded) of exposed Pennsylvanian and Lower Permian reefs.

The Lower Carboniferous to Permian succession in the Sverdrup basin reflects sedimentation on a basin-rimming, shallow platform surrounding several deeper subbasins, which are recorded by marked facies changes across the depositional axis of the basin (Fig. 2). The geometry of the facies belts varied through time, mainly as a result of broad eustatic fluctuations, progradation of carbonate platforms, and synsedimentary block-faulting indicative of contemporaneous rifting. Three major depositional sequences are recognized (Beauchamp, Harrison, and Henderson, 1989):

1) Lower Serpukhovian to upper Sakmarian,
2) Upper Sakmarian to upper Artinskian, and
3) Upper Artinskian to Wordian.

The oldest of these major sequences (1) is characterized by basal and basin-margin transgressive-regressive redbed clastics of the Borup Fiord and Canyon Fiord Formations, basin-rimming platform and reefal carbonates of the Nansen and Belcher Channel Formations, rift-controlled basin-center evaporites and carbonates of the Otto Fiord Formation, and slope and basinal shales and carbonates (including basal transgressive reefs) of the Hare Fiord Formation (Fig. 2). It is in this first sequence that four of the main groups of reefs and reef-associated biotic assemblages occur (Fig. 3). The fifth assemblage occurs in the second depositional sequence, from Sakmarian to late Artinskian time, and is characterized by sponges and bryozoans (Beauchamp, 1989a,b).

REEFS AND BIOTIC ASSEMBLAGES

The Pennsylvanian and Lower Permian buildups in the Sverdrup basin preserve evidence for a diverse marine biota ranging from foraminifers, pelmatozoans, and ammonoids to a wide variety of algae (Fig. 3). However, each reef type is characterized by one or two biotic groups that played a major role in sediment trapping, and variable roles in sediment binding, skeletal contribution, and fabric control (Fig. 4). In nearly all the reefs, syndepositional mechanical (physical) stabilization was achieved by pervasive submarine cementation (Davies, 1977a; Davies and Nassichuk, 1990). The four main biotic associations and carbonate buildups that occur in the lowest or first depositional

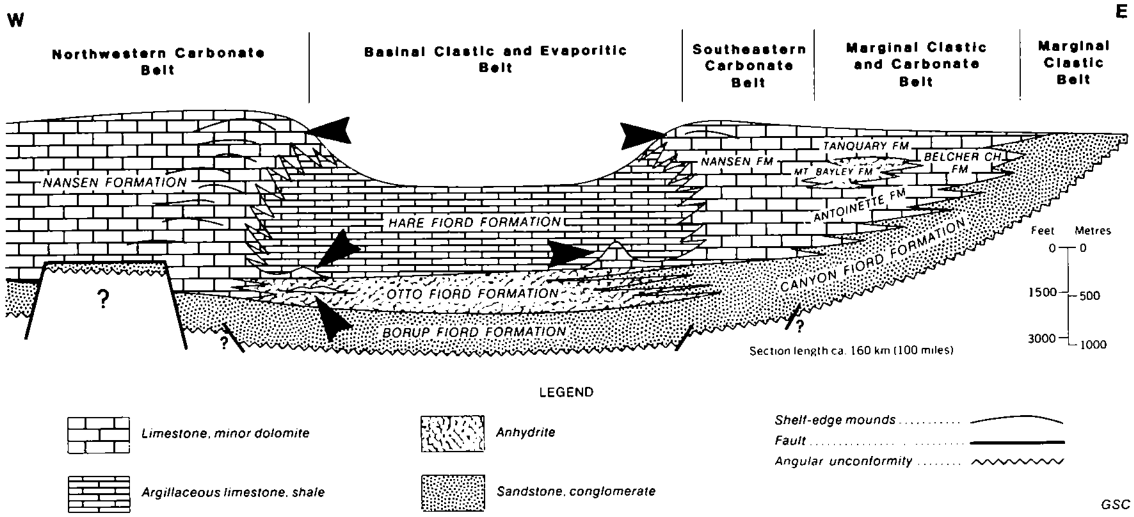

Figure 2. Schematic west-east cross-section across the northern Sverdrup basin showing inferred relationships of Mississippian–Pennsylvanian and Lower Permian lithofacies. Reef locations are indicated by arrows.

sequence, and a fifth that occurs in the middle or second sequence, are described in the following sections.

Lower Pennsylvanian Algal Reefs

Stratigraphic Setting

Lower Pennsylvanian algal reefs (Fig. 5) are exposed on northwestern Ellesmere Island within the Otto Fiord Formation, which contains interbedded anhydrite, limestone, and shale in interior regions of the Sverdrup basin (Nassichuk and Davies, 1980; Davies and Nassichuk, 1989). The ammonoids *Branneroceras branneri* and *Gastrioceras* sp., both indicative of an Early Pennsylvanian (Morrowan) age, were recovered both from below and above algal reef horizons in the type area of the Otto Fiord Formation (Nassichuk, 1975; Nassichuk and Davies, 1980). Abundant foraminifer, algal, and conodont faunas in the same interval also indicate a Morrowan age (see reports by B. L. Mamet and P. K. Bender, *in* Nassichuk, 1975).

Biotic Assemblage

Beresellid, donezellid, and other tubular and branching algae with possible dasycladacean affinities (Mamet and Rudloff, 1972) are the principal organic components of the Otto Fiord algal reefs. These algae formed three-dimensional, initially highly porous meshworks in the core stages of the Pennsylvanian reefs. The meshworks were stabilized by isopachous rims of fibrous submarine cements, originally magnesium calcite, then infilled and partly replaced by radial arrays of aragonite, now replaced by calcite spar (Fig. 6).

The Otto Fiord reefs are underlain by open marine limestones in the same formation containing a diverse faunal assemblage dominated by crinoids, but intermixed with brachiopods, ammonoids, bryozoans, gastropods, conodonts, ostracodes, and foraminifers.

Geometry and Size

Dimensions of the Lower Pennsylvanian algal reefs in the Otto Fiord Formation (Fig. 5) range from 15 to 30 m in outcrop length and 3 m in height, to 350 m in outcrop length and 35 m in height. The larger reefs in the type area of the Otto Fiord Formation occur at a single stratigraphic horizon within a rhythmically repeated succession of marine limestone and subaqueous anhydrite evaporite (Davies and Nassichuk, 1975; Nassichuk and Davies, 1980). The reefs are internally complex, with the anhydrite units off the flanks of the reefs pinching out by apparent onlap onto the flanks of the buildups between successive carbonate core-flank growth stages (Davies, 1977b; Davies and Nassichuk, 1989, fig. 7). They provide a useful surface analog for a number of evaporite-enclosed reefs in the subsurface.

Middle Pennsylvanian Bryozoan Reefs

Stratigraphic Setting

Middle Pennsylvanian reefs characterized by fenestellid bryozoans, mud-rich internal units, and massive submarine cement are exposed on northwestern Ellesmere Island and northern Axel Heiberg Islands (Davies, Nassichuk, and Beauchamp, 1989). They are buried in siltstones and argillaceous limestones near the base of the Hare Fiord and Nansen Formations (Fig. 7).

Davies and Nassichuk (1990) compared the Middle Pennsylvanian fenestellid bryozoan buildups in the Arctic with the older Waulsortian reef facies of Europe and mainland North America. Gross similarities include the major role of bryozoans in sediment trapping, the pervasive occurrence of and role played by submarine cement, the high volumes of carbonate mud in internal units, and the presence of "stromatactis"-like fabrics. However,

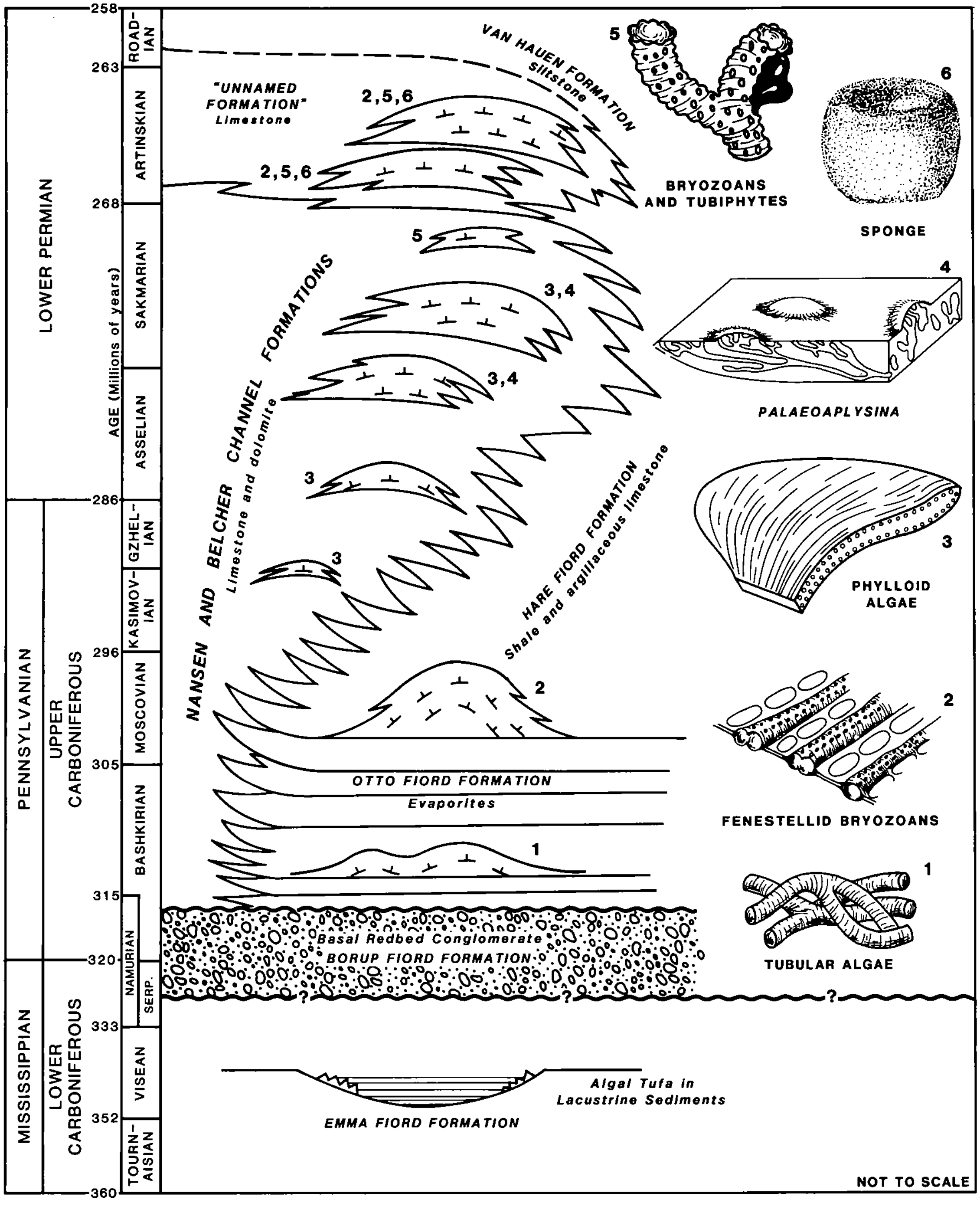

Figure 3. Schematic summary of the six organic contributors to five stages of carbonate buildup in the Pennsylvanian to Lower Permian succession in the Sverdrup basin.

because of current concerns over the use of the term "Waulsortian" (Lees, 1988), direct comparison is avoided in this paper.

Ammonoids are abundant and diverse in reefs exposed in the lower part of the Hare Fiord Formation in the vicinity of Hare Fiord and in the Blue Mountains of northern Ellesmere Island. Nassichuk (1975) described 15 species from a biocoenose on the flank of a small reef on the north side of Hare Fiord at Stepanow Creek, and 13 species from a large reef in the southern Blue Mountains. Many species from those reefs, including representatives of *Winslowoceras, Pseudopronorites, Syngastrioceras,* and *Diaboloceras,* are very similar to species from Atokan (upper Bashkirian to lower Moscovian) strata in Arkansas, Oklahoma, and Texas. A more complete discussion of Atokan ammonoids and biostratigraphy in the Canadian Arctic archipelago is contained in Nassichuk (1984).

Biotic Assemblage

Fenestellid bryozoans are texturally the most significant contributors to the Middle Pennsylvanian Arctic reefs, although volumetrically they are far outnumbered by the diverse assemblage of other skeletal components, by subpeloidal matrix, and by submarine cements. The bryozoans are interpreted to have been upright filter feeders that acted as sediment baffles. They also formed initially very porous "platestone" fabrics (Davies and Nassichuk, 1990) with ponded sediment and pervasive submarine cements (Fig. 8).

Other components of the reef biota include brachiopods, gastropods, foraminifers including fusulinaceans, and a wide variety of algae.

Geometry and Size

The Atokan fenestellid bryozoan reefs in the lower part of the basin-filling Hare Fiord Formation are the largest of the Arctic upper Paleozoic reefs, ranging from 50 to 550 m thick. These form either as individual, isolated reefs that are <90 m in diameter, such as the buildup at Stepanow Creek on Ellesmere Island, or form as large complexes, such as those in the Blue Mountains of the same island, that are at least 10 km long. The flanks of the buildups commonly are very steep (exceeding 30°, which is a result of stabilization by submarine cements. Crinoid-dominant carbonate turbidites occur on the flanks of the reefs (Davies, 1977c).

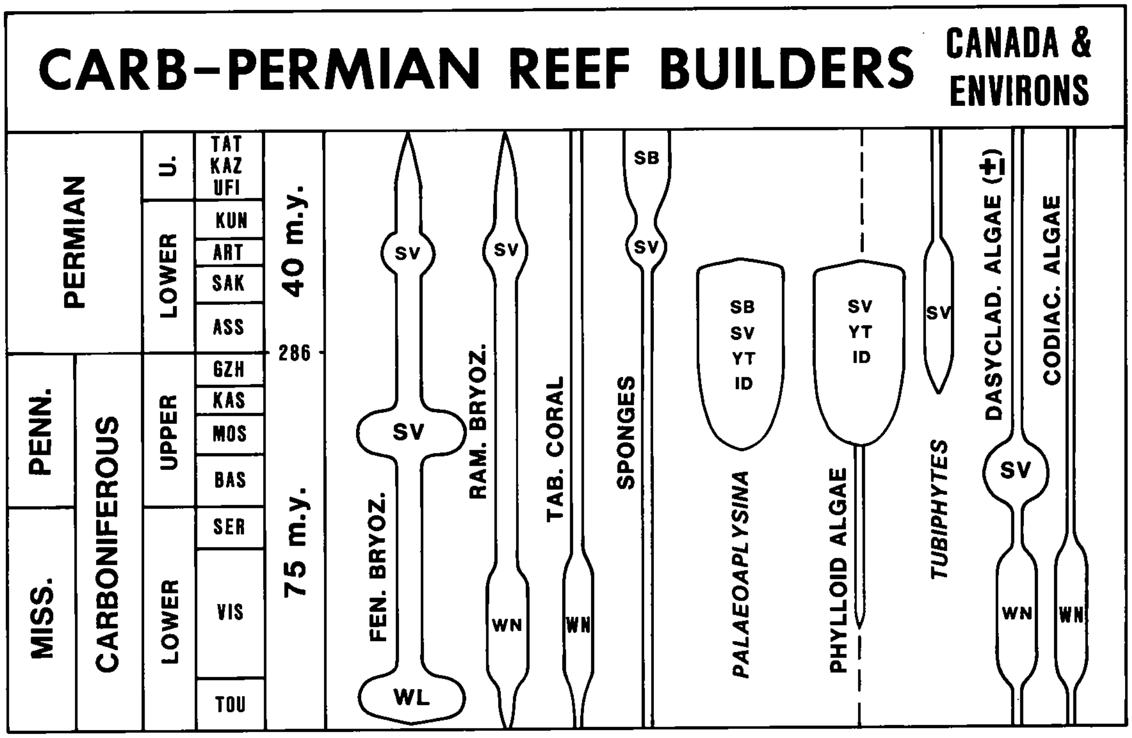

Figure 4. Temporal distribution of upper Paleozoic reef-building or reef-associated organisms in Canada and environs (from Davies, Richards, Beauchamp, and Nassichuk, 1989). Abbreviation "SV" within the columns indicates occurrences in the Sverdrup basin. Other abbreviations: WN—Windsor basin; ID—Idaho; WL—Williston (and Alberta) basin; YT—Yukon Territory; SB—Svalbard (Barents Sea basin).

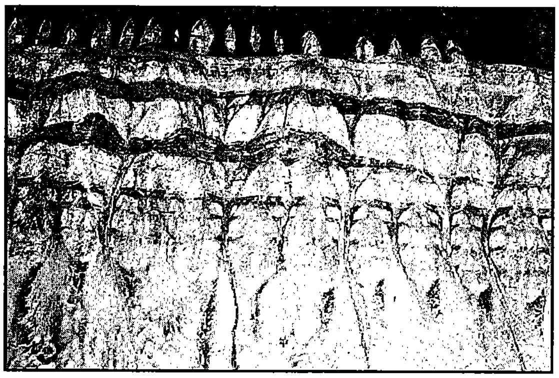

Figure 5. Algal boundstone reef of Morrowan age enclosed in anhydrite of the Otto Fiord Formation near van Hauen Pass on Ellesmere Island. The largest reefs are 115 ft thick and 1,150 ft long.

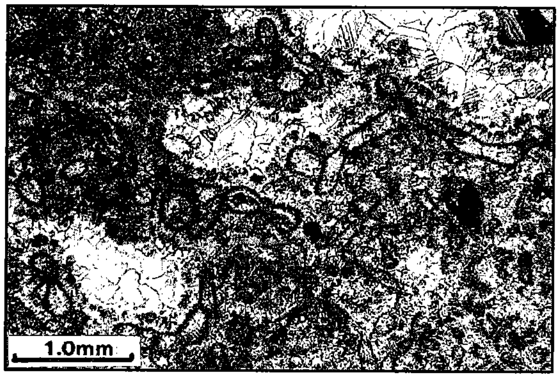

Figure 6. Photomicrograph of tubular algae in a boundstone fabric stabilized by submarine cements, characteristic of the core of Morrowan algal reefs in the Otto Fiord Formation.

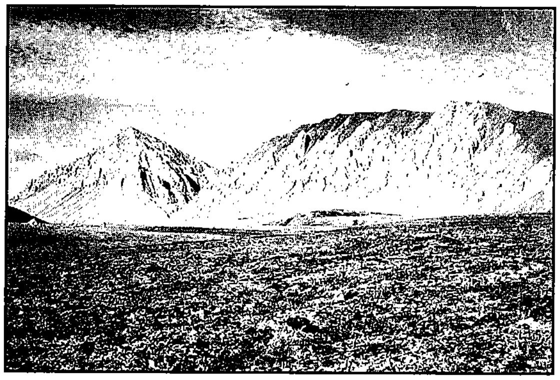

Figure 7. Exposure of Atokan fenestellid bryozoan reefs in the Blue Mountains of Ellesmere Island; the reefs are buried by siltstones and argillaceous limestones in the Hare Fiord Formation.

Middle Pennsylvanian and Permian Phylloid Algal and *Palaeoaplysina* Reefs

Stratigraphic Setting

Buildups characterized by phylloid algae and the enigmatic organism *Palaeoaplysina* occur as reef and bank morphologies (Beauchamp, Davies, and Nassichuk, 1989) in shelf, shelf margin, and upper slope settings in the Nansen and Belcher Channel Formations on Ellesmere and Axel Heiberg Islands (Fig. 9). In the thick (2,006 m) type section of the Nansen Formation on northern Ellesmere Island, phylloid algae occur throughout the uppermost member, which comprises nearly 1,215 m of massive shelf carbonate ranging from upper Middle Pennsylvanian (Desmoinesian) to Lower Permian (Sakmarian). In the lower (Desmoinesian) part of this member, phylloid algae form minor buildups in association with a variety of sponges. In the uppermost Carboniferous interval, where representatives of *Pseudofusulina* are abundant, phylloid algae are the dominant organism and *Palaeoaplysina* plays a minor role. Higher in the section, in the Lower Permian (Asselian to Sakmarian) part of the member, phylloid algae and *Palaeoaplysina* are equally abundant.

Palaeoaplysina is restricted to Upper Carboniferous and Lower Permian strata in the northern hemisphere. A number of occurrences are known from the Yukon and British Columbia (Beauchamp, Davies, and Nassichuk, 1989). Outside Canada, it is known as an important reef builder in Spitzbergen (Lonøy, 1988), in the Ural Mountains (Chuvashov, 1983), in Idaho (Breuninger, 1976), and in California (Watkins and Wilson, 1989).

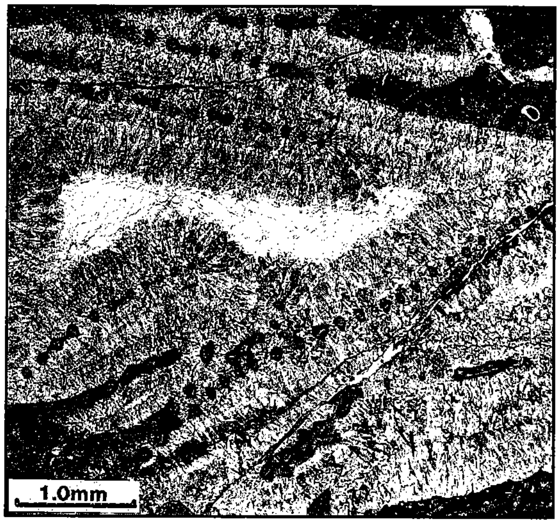

Figure 8. Photomicrograph of fenestellid bryozoan "platestone" fabric with pervasive submarine cements from Atokan reef on Ellesmere Island.

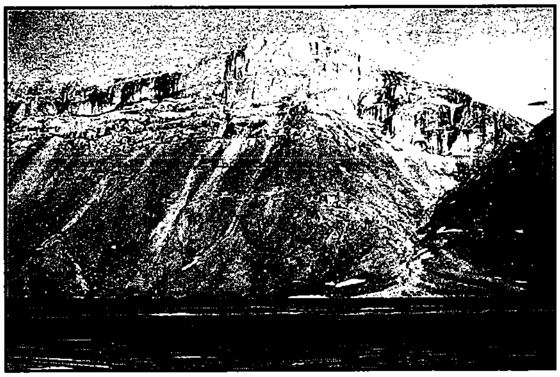

Figure 9. Lower Permian *Palaeoaplysina* and phylloid algal reef in a shelf-edge setting on the west side of Blind Fiord, Ellesmere Island (from Beauchamp, Davies, and Nassichuk, 1989).

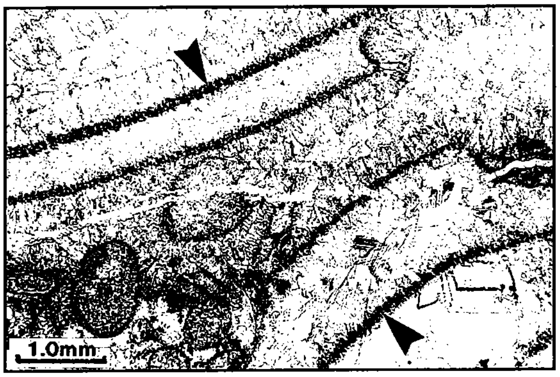

Figure 10. Photomicrograph of fragments of phylloid algal plates from the Nansen Formation on Ellesmere Island, illustrating the characteristic micrite-filled utricles (arrows).

Biotic Assemblage

The Middle Pennsylvanian and Lower Permian reefs in this category contain varying proportions of the skeletal plates of phylloid algae and *Palaeoaplysina*. "Phylloid algae" is a descriptive term for the skeletal remains of large leaf-like or blade-like, partly calcified algae (Fig. 10) without definitive taxonomic significance, even though the most common forms described in the upper Paleozoic are codiaceans (Wray, 1977). In the Sverdrup basin, the most common genus of phylloid algae is *Eugonophyllum*, followed by *Neoanchicodium* and *Ivanovia* (Mamet and others, 1979,1987; Beauchamp, Davies, and Nassichuk, 1989). The aragonitic algal plates probably grew as multisegmented erect forms with a rhizome-like substrate attachment. While the phylloid algae may have played a sediment binding and possibly a sediment trapping role, their principal contribution was as skeletal sediment. In the Sverdrup basin, phylloid algae occur in close association with *Palaeoaplysina* in mid-Pennsylvanian to Lower Permian buildups.

Palaeoaplysina formed large calcified plates (Figs. 11,12) usually <5 mm in thickness that may exceed 1 m in length. The plates have a complex internal structure of interconnecting canals, a cellular texture between the canals (Fig. 13), and surface mamelon-like protuberances (Figs. 11,14) on one side (upper?) of the plate (Davies and Nassichuk, 1973; Beauchamp, Davies, and Nassichuk, 1989).

The taxonomic affinities of *Palaeoaplysina* remain uncertain; it may be a hydrozoan, but it also has characteristics reminiscent of sponges and, less likely, of algae (Davies and Nassichuk, 1973). It may have grown as subhorizontal plates on the sea floor, or as erect fronds. In the Sverdrup basin, *Palaeoaplysina* and phylloid algae often are closely related and intermixed in rock fabrics, although the plates of the larger and more wave-resistant *Palaeoaplysina* generally are more frequent toward the margin (flank and crest) of buildups (Beauchamp, Davies, and Nassichuk, 1989).

Geometry and Size

The *Palaeoaplysina*-phylloid algal buildups occur in several morphological types (Beauchamp, Davies, and Nassichuk, 1989). Patch reefs have a maximum height of 15 m and a maximum width of 50 m. Banks have a maximum thickness of 18 m and a maximum width of 6.5 km. Reef mounds may be as much as 100 m thick and 500 m wide (Beauchamp, Davies, and Nassichuk, 1989). Like the other reefs, steep flank geometry commonly is the result of pervasive and massive submarine cementation.

Lower Permian *Tubiphytes*-Fenestellid Bryozoan Buildups

Stratigraphic Setting

A single buildup characterized by the enigmatic encrusting organism *Tubiphytes* and fenestellid bryozoans occurs in Lower Permian (Sakmarian) strata in the upper part of the Nansen Formation in the vicinity of Blind Fiord, southwestern Ellesmere Island (Beauchamp, 1989a,b). Similar buildups have not been discovered elsewhere in the Sverdrup basin, but *Tubiphytes* associated with fenestellid bryozoans is known from talus in the type area of the Nansen Formation on northern Ellesmere Island.

Biotic Assemblage

These buildups are composed of a wide variety of organisms, but the encrusting *Tubiphytes* (Fig. 15) and fragments of fenestellid bryozoans are the texturally significant components. *Tubiphytes* appeared during Pennsylvanian time, proliferated during the Permian, and continued to occupy reefal environments until Late Jurassic time (Fig. 4). *Tubi-*

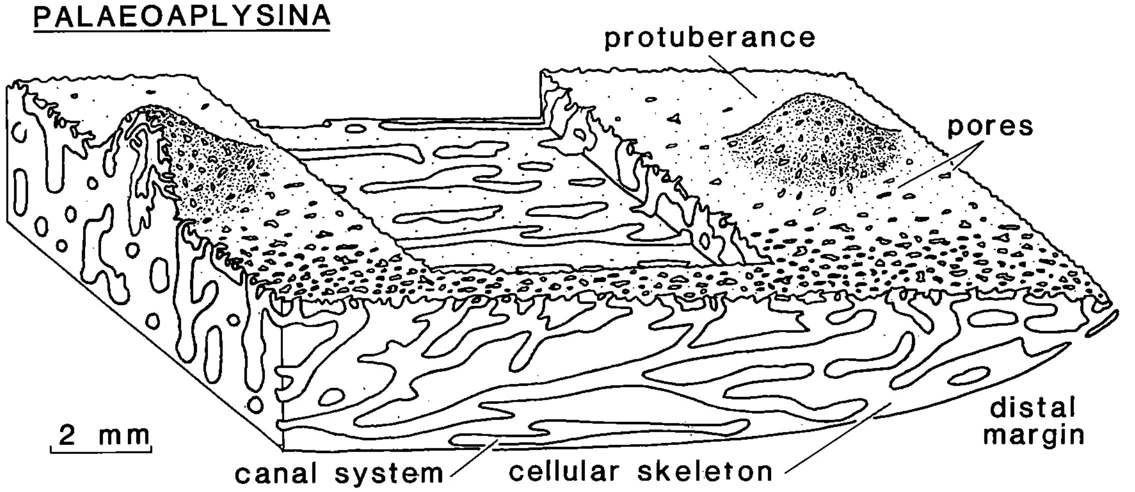

Figure 11. Diagrammatic sketch of the fabric of *Palaeoaplysina* showing internal canal system and surface mamelon-like structures (from Beauchamp, Davies, and Nassichuk, 1989).

Figure 12. *Palaeoaplysina* floatstone fabric in Lower Permian Belcher Channel Formation, southwestern Ellesmere Island.

Figure 13. Internal cellular fabric of *Palaeoaplysina* plate; cellular fabric is rarely preserved because of inferred original aragonitic composition of plate.

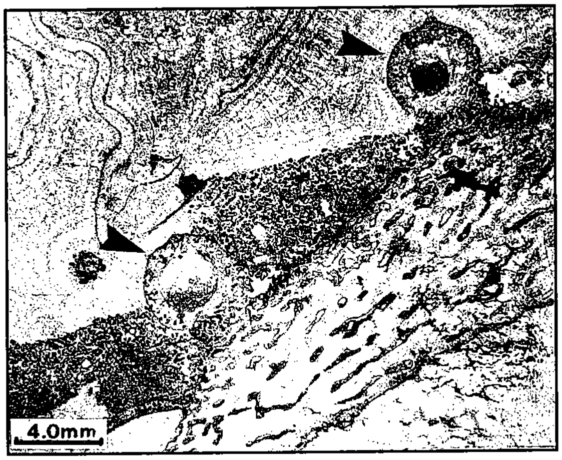

Figure 14. Photomicrograph of *Palaeoaplysina* plate from the Nansen Formation showing conceptable-like protrusions (arrows) on the upper surface of the plate.

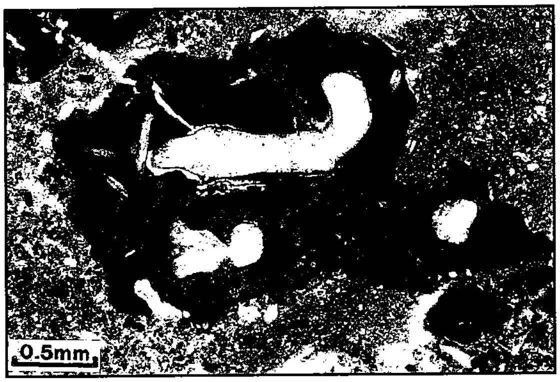

Figure 15. Photomicrograph of the enigmatic organism *Tubiphytes* from the Nansen Formation, Ellesmere Island.

phytes is characterized by a dense microcrystalline wall structure, small rounded encrusting habit with internal cavities, and spicule-like inclusions in or projecting from the wall. Associated biota, in addition to bryozoans, are brachiopods, a wide variety of foraminifers (including fusulinaceans), and calcareous algae.

Geometry and Size

The *Tubiphytes*-bryozoan buildup is <12 m thick and has a diameter of ~37 m. It occurs in a shelf-margin setting in the Nansen Formation facing the deeper trough of the Sverdrup basin.

Lower Permian Sponge-Bryozoan Buildups

Stratigraphic Setting

Buildups constructed in part by fenestellid bryozoans and sponges occur in a moderately deep, mixed siliciclastic-carbonate ramp or slope setting in an unnamed Lower Permian (Sakmarian to Artinskian) formation on southwestern Ellesmere Island (Beauchamp, 1989a). As the sponge-bryozoan buildups occur in the upper part of the formation, they probably are of Artinskian age.

Biotic Assemblage

The principal biotic constituents in the Artinskian buildups are fenestellid bryozoans and sponges, but a wide range of other organisms, including crinoids, ramose, branching and encrusting bryozoans, gastropods, brachiopods, trilobites, ostracodes, *Tubiphytes*, and foraminifers are also associated. Sponges are preserved mainly as spicules and stromatactis-like cemented molds, probably after the sponge *Raanespongeia* Rigby.

Geometry and Size

The Artinskian buildups reach a maximum height of 120 m, a width of 275 m, and length of nearly 610 m. These buildups, like all of the other upper Paleozoic reefs on Ellesmere Island, are stabilized pervasively by submarine cements.

CONCLUSIONS

Pennsylvanian and Lower Permian reefs in the Sverdrup basin of the Canadian Arctic archipelago contain diverse assemblages of organisms that played varying roles in sediment accumulation and that were in turn stabilized by pervasive submarine cements. Principal reef-builders or organic contributors were tubular and branching beresellid and donezellid algae, fenestellid bryozoans, phylloid algae, filamentous algae, *Palaeoaplysina*, *Tubiphytes*, and sponges. Associated biota include ammonoids, brachiopods, fusulinaceans and other foraminifers, conodonts, and a wide variety of algae. The total biotic assemblages in specific reef types show strong paleogeographic affinity to Carboniferous and Lower Permian sequences elsewhere in western North America as well as the Soviet Union, particularly Siberia and the Ural Mountains. Ammonoids associated with Lower Pennsylvanian algal buildups and Middle Pennsylvanian bryozoan buildups on Ellesmere Island closely resemble Morrowan and Atokan species, respectively, in midcontinental North America.

REFERENCES CITED

Beauchamp, B., 1989a, Lower Permian (Artinskian) sponge-bryozoan buildups, southwestern Ellesmere Island, Canadian Arctic archipelago, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 575–584.

______1989b, Lower Permian (Sakmarian) *Tubiphytes*-bryozoan buildup, southwestern Ellesmere Island, Canadian Arctic archipelago, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 585–589.

Beauchamp, B.; Davies, G. R.; and Nassichuk, W. W., 1989, Upper Carboniferous to Lower Permian *Palaeoaplysina*-phylloid algal buildups, Canadian Arctic archipelago, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 590–599.

Beauchamp, B.; Harrison, J. C.; and Henderson, C. M., 1989, Transgressive-regressive sequences, *part 2 of* Upper Paleozoic stratigraphy and basin analysis of the Sverdrup basin, Canadian Arctic archipelago, *in* Current research, Part G: Geological Survey of Canada Paper 89-1G, p. 115–124.

Breuninger, R. H., 1976, *Palaeoaplysina* (Hydrozoan?) carbonate buildups from upper Paleozoic of Idaho: American Association of Petroleum Geologists Bulletin, v. 60, p. 584–607.

Chuvashov, B. I., 1983, Permian reefs of the Urals: Facies, v. 8, p.191–212.

Davies, G. R., 1977a, Former magnesian calcite and aragonite submarine cements in upper Paleozoic reefs of the Canadian Arctic: a summary: Geology, v. 5, p. 11–15.

______1977b, Carbonate-anhydrite facies relationships in the Otto Fiord Formation (Mississippian–Pennsylvanian), Canadian Arctic archipelago: American Association of Petroleum Geologists, Special Volume on Reefs and Evaporites, v. 61, no.11, part 1, p. 1929–1949.

______1977c, Turbidites, debris sheets and truncation structures in upper Paleozoic deepwater carbonates of the Sverdrup basin, Arctic archipelago: Society of Economic Paleontologists and Mineralogists Special Publication 25, p. 221–247.

Davies, G. R.; and Nassichuk, W. W., 1973, The hydrozoan (?)*Palaeoaplysina* from the upper Paleozoic of Ellesmere Island, Arctic Canada: Journal of Paleontology, v. 47, p. 251–265.

______1975, Subaqueous evaporites of the Carboniferous Otto Fiord Formation, Canadian Arctic archipelago: a summary: Geology, v. 3, p. 273–278.

______1986, Ancient reefs in the high Arctic: GEOS, v. 15, p. 1–5.

______1989, Upper Carboniferous tubular algal boundstone reefs in the Otto Fiord Formation, Canadian Arctic archipelago, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 649–657.

______1990, Submarine cements and fabrics in Carboniferous to Lower Permian, reefal, shelf margin and slope carbonates, northwestern Ellesmere Island, Canadian Arctic archipelago: Geological Survey of Canada Bulletin 399, p. 1–77.

Davies, G. R.; Nassichuk, W. W.; and Beauchamp, B., 1989, Upper Carboniferous "Waulsortian" reefs, Canadian Arctic archipelago, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E. (eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 658–666.

Davies, G. R.; Richards, B. C.; Beauchamp, B.; and Nassichuk, W. W., 1989, Carboniferous and Permian reefs in Canada and adjacent areas, *in* Geldsetzer, H. H. J.; James, N. P.; and Tebbutt, G. E.(eds.), Reefs, Canada and adjacent area: Canadian Society of Petroleum Geologists Memoir 13, p. 565–574.

Lees, A., 1988, Waulsortian reefs: the history of a concept: Memoires de l'Institut geologique de l'Universite de Louvain, v. 34, p. 43–55.

Lonøy, A., 1988, Environmental setting and diagenesis of Lower Permian palaeoaplysinid build-ups and associated sediments from Bjørnøya: implications for the explanation of the Barents Sea: Journal of Petroleum Geology, v. 11, p. 141–156.

Mamet, B. L.; Nassichuk, W. W.; and Roux, A., 1979, Algues et stratigraphie du Paléozoique Supérieur de l'Arctique Canadien: Bulletin Centre Recherche et Exploration-Production Elf-Aquitaine, v. 3, no. 2, p. 669–683.

______1987, Algues et stratigraphie du Paléozoic Supérieur de l'Arctique Canadien: Geological Survey of Canada Bulletin 348, p. 1–143.

Mamet, B. L.; and Rudloff, F., 1972, Algues Carbonifères de la Partiè Septentriale de l'Amèrique du Nord: Revue Micropaleontologie, v. 15, p. 75–114.

Nassichuk, W. W., 1975, Carboniferous ammonoids and stratigraphy in the Canadian Arctic archipelago: Geological Survey of Canada Bulletin 237, p. 1–240.

______1984, Atokan stratigraphy in the Sverdrup basin, Canadian Arctic archipelago, *in* Sutherland, P. K.; and Manger, W. L. (eds.), The Atokan Series (Pennsylvanian) and its boundaries—a symposium: Oklahoma Geological Survey Bulletin 136, p. 157–167.

Nassichuk, W. W.; and Davies, G. R., 1980, Stratigraphy and sedimentation of the Otto Fiord Formation—a major Mississippian–Pennsylvanian evaporite of subaqueous origin in the Canadian Arctic archipelago: Geological Survey of Canada Bulletin 286, p. 1–87.

Shebaldin, V. P.; Selenkov, V. N.; and Akimova, A. B., 1988, Geology of the Tengiz oil field from geophysical data: International Geology Review, v. 30, p. 1052–1056.

Thorsteinsson, R., 1974, Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic archipelago: Geological Survey of Canada Bulletin 224.

Watkins, R.; and Wilson, E. C., 1989, Paleoecologic and biogeographic significance of the biostromal organism *Palaeoaplysina* in the Lower Permian McCloud Limestone, eastern Klamath Mountains, California: PALAIOS, v. 4, p. 181–192.

Wray, L. L., 1977, Calcareous algae: Elsevier, New York, 185 p.