OKLAHOMA GEOLOGICAL SURVEY
Charles J. Mankin, *Director*

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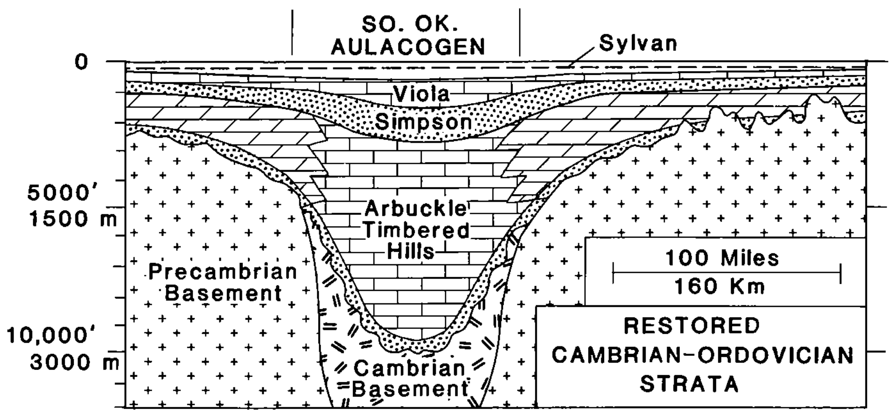
LATE CAMBRIAN–ORDOVICIAN GEOLOGY OF THE SOUTHERN MIDCONTINENT, 1989 SYMPOSIUM

KENNETH S. JOHNSON, *Editor*



Proceedings of a symposium held October 18–19, 1989, at Norman, Oklahoma.

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OKLAHOMA GEOLOGICAL SURVEY

CHARLES J. MANKIN, *Director*
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SURVEY STAFF

Title-Page Illustration

Schematic cross section showing restored thickness of Late Cambrian and Ordovician strata in Oklahoma.

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PREFACE

The Oklahoma Geological Survey (OGS) has initiated a program of yearly symposia dealing with a topic of major concern to geologists and others involved in exploration and resource development in Oklahoma. The first symposium, on the Anadarko basin, was held in 1988; the proceedings were published as OGS Circular 90. The second symposium, the current one, deals with all aspects of Late Cambrian and Ordovician geology of the southern Midcontinent.

Rocks of Late Cambrian and Ordovician age in most parts of the southern Midcontinent are a thick sequence of shallow-marine carbonates, interbedded with thinner marine sandstones and shales. They grade laterally into deep-water black shales, cherts, and sandstones in the Ouachita trough. These strata preserve a nearly continuous record of sedimentation in the Oklahoma basin and the Ouachita trough, and they are of considerable economic importance as sources of oil and gas, non-fuel minerals, and ground water, as well as disposal zones for liquid wastes.

To provide a forum for open discussion of research on Late Cambrian and Ordovician geology of the southern Midcontinent, the OGS sponsored this symposium October 18–19, 1989. The symposium was held at the Oklahoma Center for Continuing Education, The University of Oklahoma, in Norman, and this volume contains the proceedings of that conference. Research reported upon at the symposium includes petroleum exploration and development, structure, sedimentation, diagenesis, petrography, geochemistry, biostratigraphy, dolomitization, and paleokarst. We hope that the symposium and these proceedings will bring such research to the attention of the geoscience community, and will help foster exchange of information and increased research interest by industry, university, and government workers. Seventeen papers were presented orally at the symposium, and an additional 15 reports were given as posters. All 17 of the oral papers are presented here as full papers or abstracts, and 14 of the poster presentations are given as short reports or abstracts. About 300 persons attended the symposium.

Stratigraphic nomenclature and age determinations used by the various authors in this volume do not necessarily agree with those of the OGS.

Persons involved in the organizing, planning, and execution of this symposium include: Kenneth Johnson, Charles Mankin, Jock Campbell, Lloyd Gatewood, Suzanne Takken, Bob Allen, and Phil Chenoweth (Symposium Steering Committee); Michelle Summers and Tammie Creel (Registration Co-Chairs); Jock Campbell (Poster Session Chair); Connie Smith (Publicity Chair); and Helen Brown and Gwen Williamson (Exhibits Coordinators). Appreciation is expressed to each of them and to the many authors who worked toward a highly successful symposium.

KENNETH S. JOHNSON
General Chairman

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PART I

PAPERS PRESENTED AT THE SYMPOSIUM

GEOLOGIC OVERVIEW AND ECONOMIC IMPORTANCE OF LATE CAMBRIAN AND ORDOVICIAN ROCKS IN OKLAHOMA

Kenneth S. Johnson
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ABSTRACT.—Late Cambrian and Ordovician strata in most parts of Oklahoma are 1,000 to 10,000 ft of shallow-marine carbonates (with several sandstone and shale interbeds) deposited in the Oklahoma basin, a broad epicontinental sea that extended across almost all parts of the southern Midcontinent. Strata are remarkably widespread and laterally persistent in most of the Oklahoma basin, reflecting the stability of the region and the importance of epeirogenic movements during early Paleozoic time. Strata include the Timbered Hills, Arbuckle, Simpson, and Viola Groups; the Sylvan Shale; and the base of the Hunton Group (Keel oolite). The depocenter for these sediments was the southern Oklahoma aulacogen, which embraced the Anadarko, Ardmore, and Marietta protobasins, as well as the future sites of the Arbuckle, Wichita, and Criner uplifts. Equivalent strata that accumulated in the Ouachita trough of southeastern Oklahoma are more than 3,000 ft of black shale, chert, and sandstone.

Rocks of Late Cambrian and Ordovician age are of major economic significance in Oklahoma. They are important sources of oil and gas, nonfuel minerals, and ground water; they are also used as disposal zones for oil-field brines and hazardous wastes.

From 1983 to 1988, the average annual yield from Cambrian and Ordovician rocks in Oklahoma was ~17.7 million bbl of oil and 55.5 Bcf of natural gas: this represented about 11.8% and 2.8%, respectively, of total State oil and gas production during this time. Simpson strata yielded most of the oil and gas produced from Cambrian–Ordovician reservoirs, followed by Viola, Arbuckle, and Timbered Hills production. Recent discoveries in Arbuckle Group rocks of southern Oklahoma have yielded major production of oil from Cottonwood Creek field and natural gas from Wilburton field.

Principal nonfuel mineral resources include limestone, dolomite, silica sand, and rock asphalt; other resources are shale, hematite, limonite, manganese, zinc, lead, chert, and vein quartz. The major area for mining these resources is the Arbuckle Mountains; other outcrop areas include the Wichita and Ouachita Mountains, the Criner Hills, and parts of the Ozark uplift.

Significant ground-water resources are present in aquifers of the Arbuckle and Timbered Hills Groups in the Wichita Mountains–Lawton region, and in the Simpson and Arbuckle Groups in the Arbuckle Mountains. The Roubidoux, Gasconade, and Eminence Formations are an important aquifer in northeastern Oklahoma, and the Bigfork Chert is fractured and yields water in the Ouachita Mountains.

Porous and permeable zones of the Arbuckle and Simpson Groups deep below the land surface locally are suitable hosts for disposal of liquid wastes; they are used extensively for salt-water disposal in the State, and are also used for disposal of industrial wastes in 10 wells at 8 locations in the Tulsa, Pryor, Bartlesville, and Oklahoma City areas.

INTRODUCTION

This paper discusses the geology of Late Cambrian and Ordovician rocks in Oklahoma and presents information on the economic importance of these rocks. A geologic overview is given first, followed by discussions of the oil and gas, nonfuel, and ground-water resources; a discussion of the use of Late Cambrian and Ordovician strata as host rocks for disposal of liquid wastes is given in conclusion. This paper is also intended to help set the stage for the other reports that follow in this symposium volume.

Appreciation is expressed to Rodger E. Denison and R. Nowell Donovan for reviewing this manuscript.

GEOLOGIC OVERVIEW

Oklahoma is a geologically complex region with a number of major depositional and structural basins, separated by orogenic uplifts and mountain ranges created mainly during Pennsylvanian time (Fig. 1). However, the early Paleozoic history of the State is tectonically and sedimentologically much simpler. Late Cambrian and Ordovician strata in most parts of Oklahoma and the southern Midcontinent are characterized by 1,000–10,000 ft of shallow-marine carbonates (limestone and dolomite), interbedded with several sandstone and shale units. These strata were deposited in a broad epicontinental sea, the Oklahoma basin (Fig. 2), which extended across all parts of Okla-

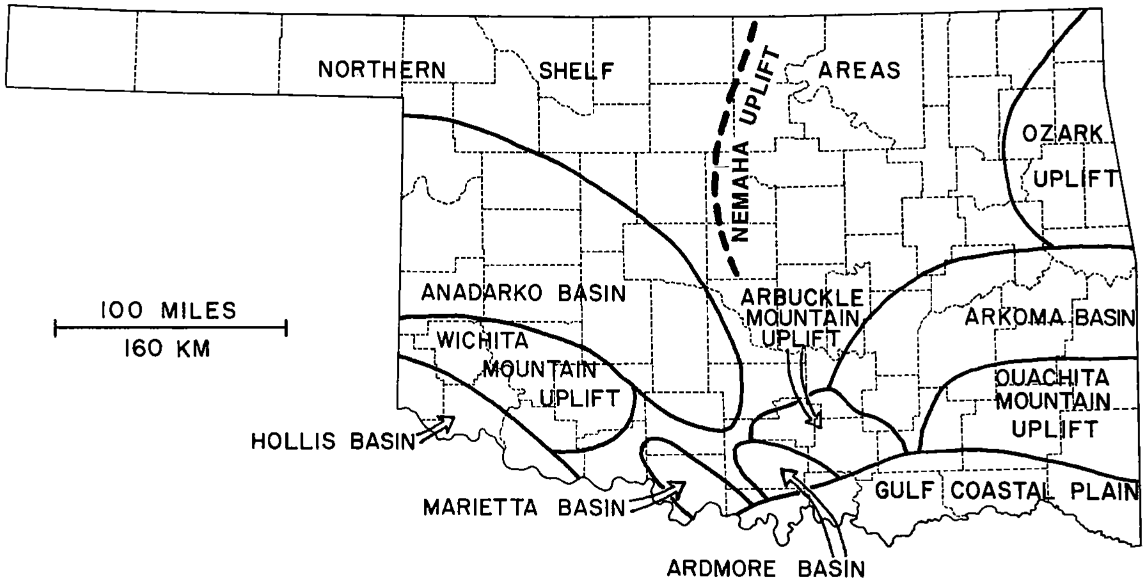

Figure 1. Major geologic provinces of Oklahoma (Johnson and others, 1980).

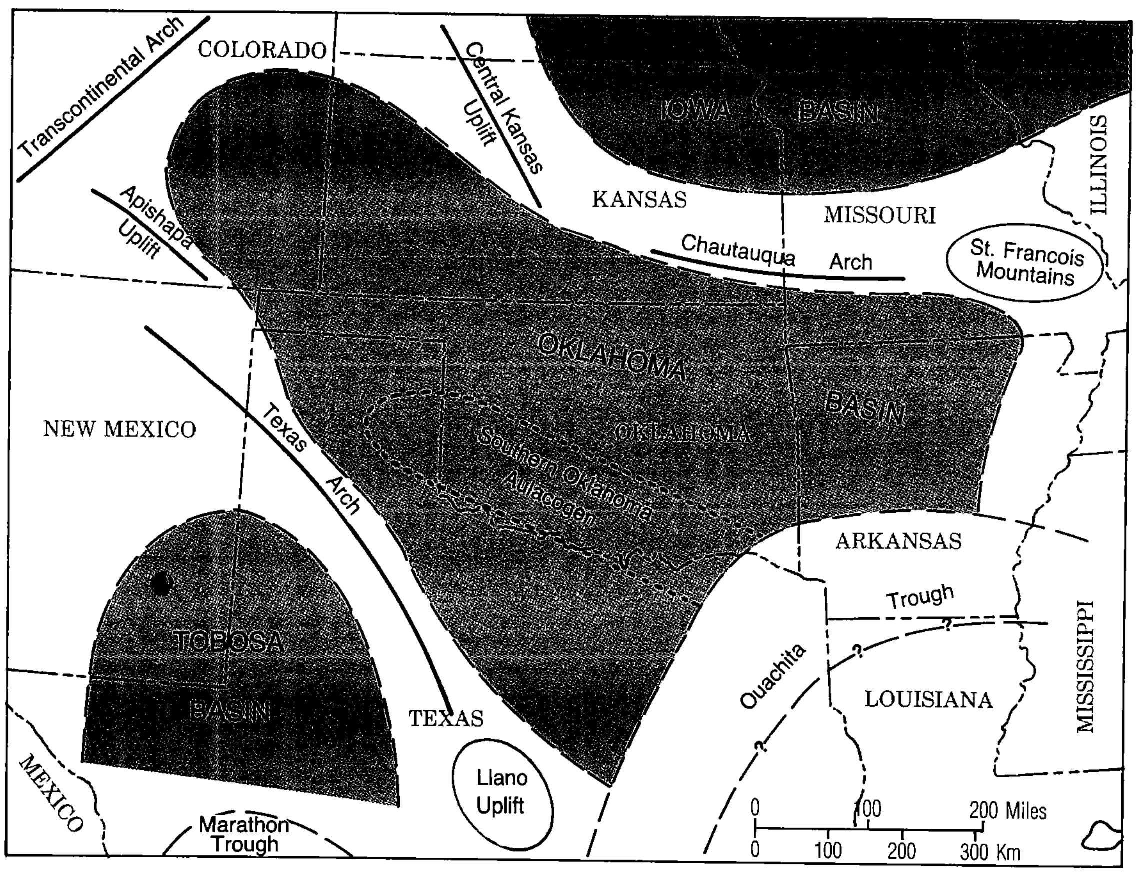

Figure 2. Map of southwestern United States, showing approximate outline of the Oklahoma basin and other major features that existed in early and middle Paleozoic time (Johnson and others, 1988).

homa (except the Ouachita trough in the southeast) during Late Cambrian through Mississippian time (Johnson and others, 1988). Stratigraphic units deposited in most parts of the Oklahoma basin are remarkably widespread and laterally persistent, reflecting the stability of this part of the craton and the importance of epeirogenic (rather than orogenic) movements during early Paleozoic time.

The depocenter for the Oklahoma basin was the southern Oklahoma aulacogen (SOA) (Fig. 2), a WNW-trending trough where subsidence and sediment accumulation was two to three times greater than on nearby shelf areas. The SOA comprised a region that now includes the Anadarko, Ardmore, and Marietta basins, along with the Arbuckle anticline, the Wichita Mountains uplift, and the Criner Hills.

Precambrian through Middle Cambrian basement rocks that underlie the Oklahoma basin consist mainly of igneous rocks, but some low-rank metasedimentary rocks are also present (Fig. 3) (Ham and others, 1964; Denison, 1981; Denison and others, 1984; Johnson and others, 1988). Precambrian rocks beneath most of Oklahoma are epizonal and mesozonal granites with comagmatic rhyolites. Early and Middle Cambrian basement rocks, limited to the aulacogen (Fig. 4), consist of granite, rhyolite, gabbro, and basalt. Basement rocks beneath the Ouachita trough have not been drilled and are unknown.

Late Cambrian seas transgressed to the north and west across Oklahoma, covering a basement-rock surface that was of modest to locally rugged relief. A time-transgressive basal sand, the Reagan Sandstone, was deposited across most of the State; it grades upward into the succession of shallow-water carbonates and thin clastic units deposited more or less continuously until the end of Ordovician time. The term "Arbuckle facies" is used commonly for the sequence of thick carbonates and accompanying shales and sandstones, extending from the base of the Reagan Sandstone to the top of the Late Devonian–Early Mississippian Woodford Shale (Ham, 1959), and they are named for the excellent exposures in the Arbuckle Mountains: these strata were deposited mainly in shallow-marine environments ranging from the intertidal zone to water depths of several hundred feet. In contrast, the "Ouachita facies" consists mainly of black shales, cherts, and sandstones of equivalent age, deposited in several thousand feet of water in the Ouachita trough off the southern margin of the North American craton. Reports

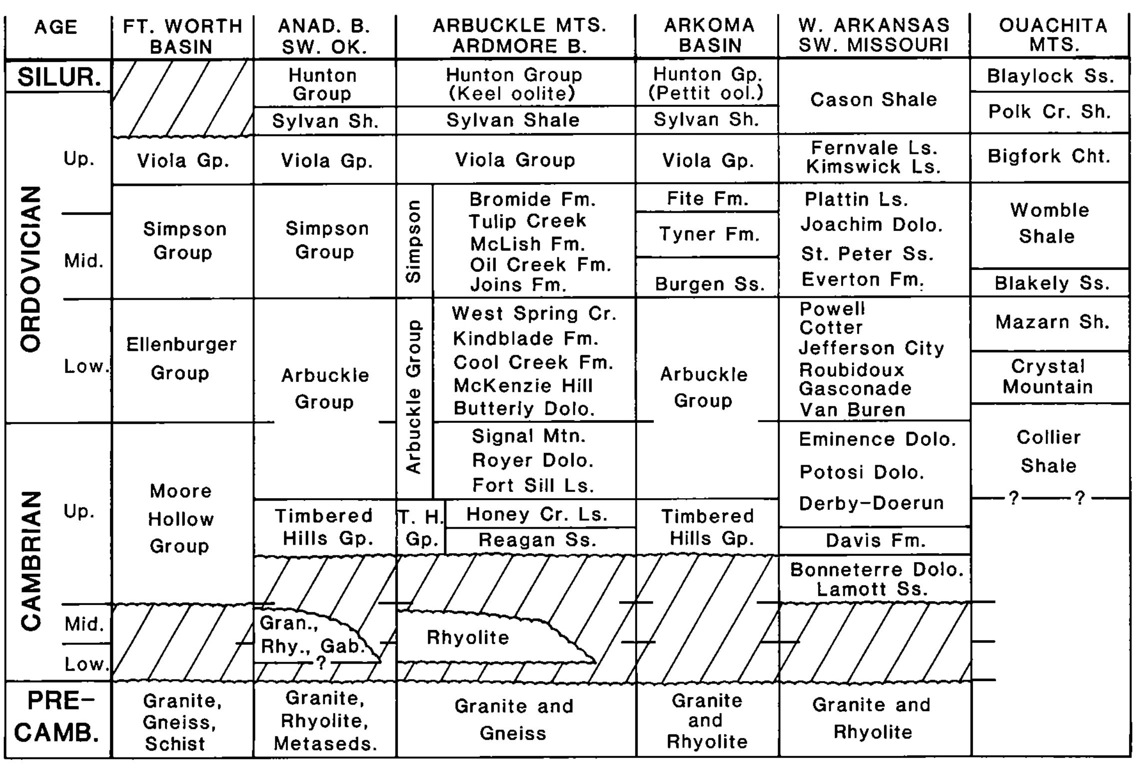

Figure 3. Correlation chart for Cambrian and Ordovician rocks of Oklahoma and adjacent areas, based on COSUNA charts (Hills and Kottlowski, 1983; Mankin, 1987).

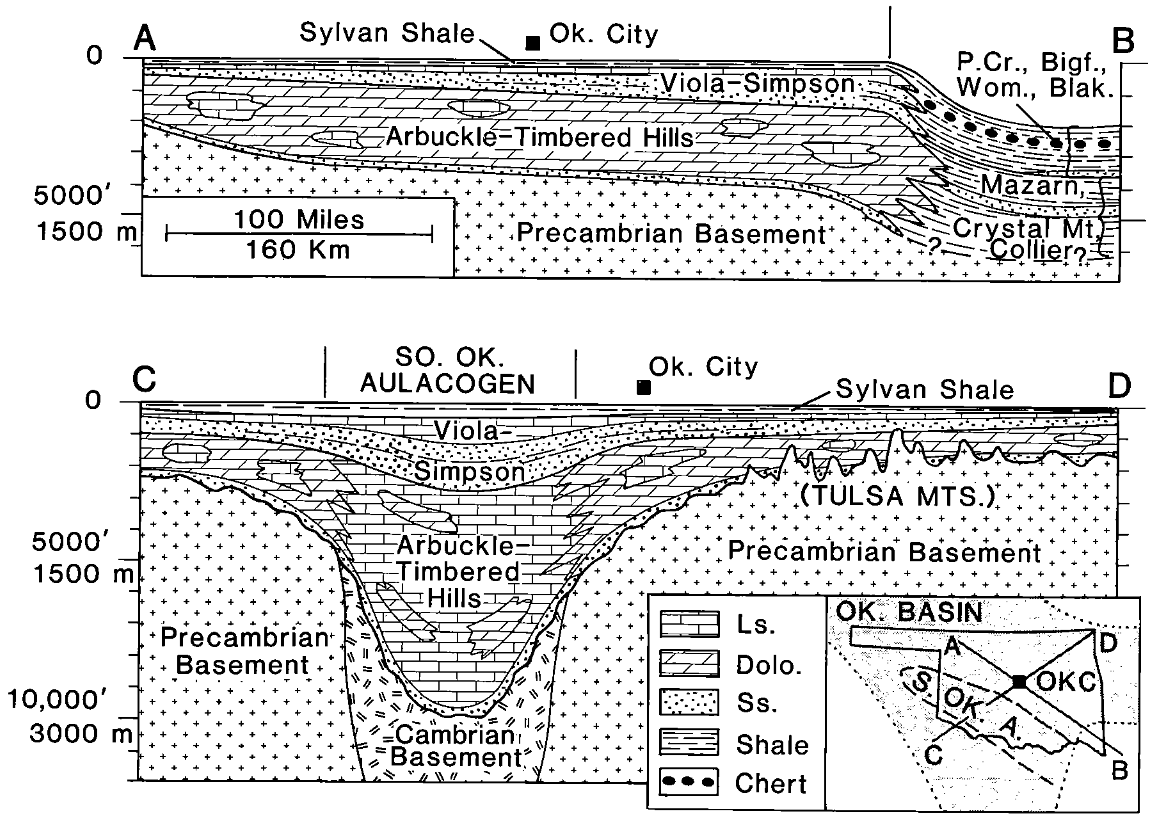

Figure 4. Schematic cross sections showing restored thickness of Late Cambrian and Ordovician strata in Oklahoma (based on data in Johnson and others, 1988).

dealing broadly with these Late Cambrian and Ordovician strata in Oklahoma have been published in recent years by Ham (1959,1969); Schramm (1964,1966); McHugh (1964); Herndon (1965); Ham and Wilson (1967); Donovan (1986); Johnson and others (1988); and Fay (1989); the current report draws upon each of these earlier reports.

The Reagan sandstone and overlying Honey Creek Limestone constitute the Timbered Hills Group of Late Cambrian age (Fig. 3). The Reagan was deposited throughout the Oklahoma basin, except on topographic highs such as the "Tulsa Mountains" (basement highs in northeastern Oklahoma, as shown on Fig. 4) or the basement-rock hills buried in the Arbuckle Mountains (Ham, 1969) and the Wichita Mountains (Donovan, 1986). The Reagan is a feldspathic and glauconitic sandstone, consisting mainly of reworked detritus weathered from the basement complex. It commonly is 50–200 ft thick, but reaches 450 ft in the Arbuckle Mountains part of the SOA. The Honey Creek is a fossiliferous limestone in the aulacogen that grades into sandy dolomite on the shelf: the unit typically is 100–200 ft thick. Rocks equivalent to the Timbered Hills Group in the Ouachita province are neither exposed or encountered in drilling.

The Arbuckle Group, which overlies the Timbered Hills Group, is the thickest sequence of lower Paleozoic strata in Oklahoma; it is as much as 6,700 ft of limestone in the aulacogen, as exposed in the Arbuckle anticline (Fay, 1989), and it thins to ~4,000 ft of dolomite in the eastern Arbuckle Mountains, and about 1,000–4,000 ft of dolomite in most other shelf areas of the Oklahoma basin (Johnson and others, 1988). The transition from mainly limestone to mainly dolomite occurs near the boundary of the SOA (Fig. 4), the dolomite probably resulting from deposition of carbonates in an alternating shallow-marine and supratidal environment, as well as their early exposure to meteoric waters. Paleokarst features, such as dissolution cavities, collapse breccias, fractures enlarged by dissolution, and locally extensive vuggy porosity, probably result in part from intermittent exposure and subsidence of the vast carbonate shelf during Arbuckle deposition. Anhydrite has been encountered in Arbuckle Group boreholes in parts of south-central Oklahoma, but outcrop evidence is limited to anhydrite relics in chert, salt pseudomorphs, and karstic collapse features.

The Late Cambrian–Early Ordovician Arbuckle Group is divided into eight formations (six limestones and two secondary dolomites), based upon excellent exposures in the Arbuckle and Wichita Mountains (Fig. 3). The limestone units are readily correlated from the Arbuckle Mountains to outcrops in the Slick Hills, on the northeast side of the Wichita Mountains (Donovan, 1986), although zones of secondary dolomite here are present at stratigraphic levels other than those in the Arbuckle Mountains. Limestones of these two major outcrop areas attest to shallow-water marine deposition. The strata consist mainly of cyclic arrangements of interbedded carbonate mudstones, intraclast calcarenites, oolitic calcarenites, stromatolites, and laminated dolomites or dolomitic limestones (Ham, 1969). They are highly fossiliferous, containing trilobites, brachiopods, mollusks, pelmatozoans, sponges, and stromatolites. In the shelf areas, equivalent strata consist of similar limestones that have been wholly or partly dolomitized.

Arbuckle Group shallow-water carbonates give way to the Ouachita facies of black shales, sandstones, and fine-grained dark limestones in the Ouachita trough (Fig. 4). The Collier, Crystal Mountain, and Mazarn Formations are sparingly fossiliferous, but graptolites enable reasonable correlation of these units with the Arbuckle facies. The base of the Collier Shale is not known in the Oklahoma Ouachitas, so the full thickness of these strata is uncertain. Estimated thicknesses in McCurtain County are 600 ft of Mazarn, 100 ft of Crystal Mountain, and at least 300 ft of Collier (Ham, 1959); the Viersen and Cochran No. 25-1 Weyerhaeuser well, drilled in the core of the Broken Bow uplift, penetrated ~10,000 ft of complexly folded and faulted black phyllites, quartzites, and dolomitic marble without reaching basement (Goldstein, 1975).

Rocks of the Simpson Group represent a profound change of depositional environment in the Oklahoma basin. Thick beds of cleanly washed quartzose sandstone are interbedded with thick, shallow-water marine limestones and thin to moderately thick, greenish-gray shales. Total thickness of the Simpson ranges from 100 to 300 ft in the northern part of Oklahoma (it has been eroded in the northeastern counties), to as much as 2,300 ft in the Arbuckle anticline, along the aulacogen depocenter (Statler, 1965). In Arbuckle Mountains outcrops, each formation except the lowermost Joins is typified by a basal quartzose sand overlain by limestone with varying amounts of shale. The sands, equivalent to the St. Peter Sandstone farther north, are very fine- to fine-grained, well sorted, in some cases uncemented, and (where uncemented) have high porosity (30%) and high permeability (1–3 darcies). The thick basal sands of the Simpson formations are regionally extensive, sheet-like sand bodies, generally 50–200 ft thick. In the subsurface of central Oklahoma some of these sands are referred to as "Wilcox sands" ("First Wilcox" or "Second Wilcox"). Other sands in the Simpson are highly variable in thickness and less continuous, and cannot be traced far laterally (Statler, 1965).

Simpson deposition began after sea level was lowered, resulting in the draining and exposure of most of the vast carbonate shelf built up during Arbuckle sedimentation (McPherson and others, 1988). Marine carbonate deposition continued in the aulacogen during Joins time, while a blanket of sand, probably windblown from the north, mantled the emergent carbonate shelf. As sea level rose again, the sand was reworked into an extensive and sheet-like transgressive deposit, which is overlain by marine shales and carbonates. Successive fluctuations of sea level during the remainder of Simpson time produced a series of sands, shales, and limestones.

In the Ouachita trough, strata equivalent to the Simpson Group are the Blakely Sandstone and Womble Shale. In Oklahoma, the Blakely is 10–15 ft of gray to brown, fine- to coarse-grained, micaceous, quartzose sandstone. The Womble is ~600 ft of gray to black shale interbedded with minor amounts of siltstone and limestone.

The Viola Group is a marine-limestone sequence that is widespread in most parts of the Oklahoma basin. It contains nodular chert at several stratigraphic levels and commonly is highly fossiliferous. The Viola Group embraces clean-washed skeletal limestones (Welling Formation = "Fernvale" Formation) in the upper part. Viola strata are generally 600–900 ft thick in the aulacogen, and they thin away from the depocenter to 50–500 ft in most other parts of the State (they are eroded from the Ozark uplift in the northeast).

The Bigfork Chert, the Ouachita-facies equivalent of the Viola Group, consists of dark-gray to black, bedded cherts, interbedded with graptolitic black shales and gray to black, cherty limestones. The thickness of the Bigfork Chert is 600–800 ft. Equivalence of the Viola and Bigfork is well documented (Ham, 1959; Finney, 1988), and it demonstrates clearly the transition from Arbuckle facies to Ouachita facies in a single rock unit: shallow-water, light-gray limestone containing disseminated silica and nodules of chert of the Oklahoma basin (Arbuckle facies) grades laterally into black shale, bedded chert, and cherty limestone in Black Knob Ridge and the core area of the Ouachita Mountains.

The Sylvan Shale is widespread and greenish gray. It is noncalcareous in its lower part in eastern Oklahoma, and becomes increasingly calcareous throughout its thickness in western Oklahoma; locally it grades into argillaceous, skeletal carbonate in the northwest. The thickness of the Sylvan ranges from 30 to 200 ft in most shelf areas, and is 300–400 ft in the aulacogen. Strata equivalent to the Sylvan in the Ouachita trough are the Polk Creek Shale, which consists of 75–150 ft of dark-gray to black shale with some chert nodules.

The youngest Ordovician unit in the Oklahoma basin is the Keel oolite (Pettit oolite in the east), at the base of the Hunton Group. The Keel is part of an oolitic facies distributed over a wide area of eastern North America, and results from eustatic shoaling of sea water with no significant terrigenous sedimentation. In Oklahoma the Keel has a patchy distribution; where present, it is 10–20 ft thick. The Keel is probably equivalent to the upper part of the Polk Creek Shale in the Ouachita trough.

OIL AND GAS RESOURCES

Cambrian and Ordovician strata have been, and are, important sources of oil and natural gas in Oklahoma. Production data for the years 1983–88 have been computerized by the Oklahoma Geological Survey's Natural Resources Information System (NRIS). During these six years, Cambrian and Ordovician rocks in Oklahoma yielded an annual average of ~17.7 MM bbl (million barrels) of oil and 55.5 Bcf (billion cubic ft) of natural gas (Tables 1,2). This represented about 11.8% and 2.8%, respectively, of the total State oil and gas production during this time.

Simpson production was greater during 1983–88 than that of any other Cambrian–Ordovician unit (Tables 1,2). Simpson oil production averaged ~12.1 MM bbl during those six years, and accounted for 69% of all Cambrian–Ordovician oil production in the State. Also, gas production from the Simpson averaged ~30.7 Bcf and was 55% of all Cambrian–Ordovician production in the State. The remaining Cambrian–Ordovician units, in decreasing order of production for both oil and gas, are the Viola, Arbuckle, and Timbered Hills Groups (Tables 1,2). Comingled Cambrian and Ordovician production (i.e., Viola and Simpson) of oil and gas is fairly small; it is not possible to determine the amount of Cambrian–Ordovician production where it is comingled with non-Cambrian–Ordovician production (i.e., Simpson comingled with Pennsylvanian).

Many oil and gas fields in Oklahoma produce from Cambrian and Ordovician rocks. A total of 2,158 fields throughout the State produced oil during 1983–88, and 582 of them yielded some or all of their oil from Cambrian–Ordovician reservoirs (data from NRIS files). Likewise, of the 1,590 gas fields that operated during the same period of

TABLE 1. — CAMBRIAN–ORDOVICIAN OIL PRODUCTION, BY GROUP, 1983–88

Group	Annual Avg., 1983–88 (MM bbl)	Percent of Camb.–Ord. (%)
Viola	4.1	23
Simpson	12.1	69
Arbuckle	0.9	5
Timbered Hills	0.003	0.02
Comingled Camb.–Ord.*	0.6	3
Total Camb.–Ord. oil	17.7	100

Total of all oil for State = 149.7 MM bbl

$$\frac{\text{Total Camb.–Ord. oil}}{\text{Total of all oil for State}} = \frac{17.7\ \text{MM bbl}}{149.7\ \text{MM bbl}} = 11.8\%$$

*Does not include Cambrian–Ordovician oil comingled with non-Cambrian–Ordovician oil (i.e., Simpson oil comingled with Pennsylvanian oil).
Source: NRIS.

TABLE 2. — CAMBRIAN–ORDOVICIAN GAS PRODUCTION, BY GROUP, 1983–88

Group	Annual Avg., 1983–88 (Bcf)	Percent of Camb.–Ord. (%)
Viola	15.6	28
Simpson	30.7	55
Arbuckle	6.6	12
Timbered Hills	0.01	0.02
Comingled Camb.–Ord.*	2.6	5
Total Camb.–Ord. gas	55.5	100

Total of all gas for State = 1,963 Bcf

$$\frac{\text{Total Camb.–Ord. gas}}{\text{Total of all gas for State}} = \frac{55.5\ \text{Bcf}}{1{,}963\ \text{Bcf}} = 2.8\%$$

*Does not include Cambrian–Ordovician gas comingled with non-Cambrian–Ordovician gas (i.e., Simpson gas comingled with Pennsylvanian gas).
Source: NRIS.

time, 284 produced partially or totally from Cambrian–Ordovician rocks. Major yields of oil from Cambrian–Ordovician reservoirs in recent years came from such fields as St. Louis, Golden Trend, Payne, and Eola-Robberson (Table 3); major Cambrian–Ordovician gas fields include Golden Trend, Custer City North, Wilburton, and Eola-Robberson (Table 4).

Recent discoveries of oil and gas in Arbuckle strata in the Cottonwood Creek and Wilburton fields of southern Oklahoma have set off a new wave of interest in prospecting these lower Paleozoic reservoirs. Cottonwood Creek field, discovered by CNG Producing Co. in November 1987, is producing from the Arbuckle Group "Brown zone" at depths of 8,500–9,500 ft along the Ardmore basin's Hewitt field trend in Carter County. Initial flow rates in the 14-well field approached 4,000 bbl of oil and 3 MMcf of gas per day. Production during 1987–88 was 493,000 bbl of oil (Table 3), and through October 1989 these wells had produced more than 1 million bbl of oil and 500 MMcf of casinghead gas (Petroleum Information, Mid-continent Regional Newsletter, November 6, 1989). The deep Wilburton field, opened by ARCO Oil & Gas Co. in late 1987, is producing gas from the Arbuckle Group at depths of 12,800–14,500 ft. Open-hole potential for some of the wells has been >100 MMcf per day, although sustained production generally has been 15–35 MMcf per day per well. Production during 1987–88 was 17.932 Bcf of gas (Table 4), and through May 1989 seven of the field's nine Arbuckle wells had produced 43.9 Bcf (Petroleum Information, Mid-continent Regional Newsletter, August 31, 1989).

TABLE 3. — TOP 20 OIL FIELDS PRODUCING FROM CAMBRIAN–ORDOVICIAN ROCKS, 1983–88

Oil field	Annual average, 1983–88 (M bbl)
St. Louis	862
Golden Trend	738
Payne	665
Eola–Robberson	564
Davis NE	435
Davis SW	326
Earlsboro	303
Oklahoma City	301
Apache	282
Davis	247
Whitebead West	233
Bebee–Konawa SW	217
Butterly NE	203
Seminole	179
Healdton	164
Washington District	162
Cumberland	161
Bowlegs	158
Lucien	142
Cushing	142

Note: Production from the Cottonwood Creek field was limited to 1987 (79 M bbl) and 1988 (414 M bbl).

Source: NRIS.

TABLE 4. — TOP 20 GAS FIELDS PRODUCING FROM CAMBRIAN–ORDOVICIAN ROCKS, 1983–88

Gas field	Annual average, 1983–88 (Bcf)
Golden Trend	7.040
Custer City North	3.361
Wilburton*	2.989
Eola–Robberson	2.465
Mayfield West	2.334
Enville SW	2.327
Payne	1.952
Knox	1.347
Sooner Trend	1.132
Cumberland	1.012
Powell South	1.010
Union City	0.935
Madill	0.871
Crescent–Lovell	0.727
Aledo**	0.618
Oklahoma City	0.603
Apache Townsite	0.445
Enville West	0.412
Sho-Vel-Tum	0.400
Aylesworth District SE	0.358

*Production from the Wilburton field was limited to 1987 (0.197 Bcf) and 1988 (17.735 Bcf).

**Production from the Aledo field was limited to 1987 (1.111 Bcf) and 1988 (2.594 Bcf).

Source: NRIS

NONFUEL MINERAL RESOURCES

Principal nonfuel mineral resources in Late Cambrian and Ordovician rocks of Oklahoma include limestone, dolomite, silica sand, and rock asphalt. Other resources include shale, hematite, limonite, manganese, zinc, lead, chert, and vein quartz. The greatest area of outcrops of these mineral deposits is in the Arbuckle Mountains, although important outcrops are present in the Wichita and Ouachita Mountains and the Criner Hills, and small exposures are scattered in the Ozark uplift (Fig. 5). A map showing nonfuel mineral resources of Oklahoma was prepared by Johnson (1969).

Limestone reserves are enormous and are mainly in the Arbuckle and Viola Groups. These units are quarried at many places in the Arbuckle and Wichita Mountains and the Criner Hills for aggregate (crushed stone), which is used in building roads and in other construction (Fig. 5). Quarries producing millions of tons of Arbuckle and Viola stone each year are the main source of aggregate being supplied to the Oklahoma City metropolitan area, as well as to southern and western parts of the State. Stone from some of these quarries is also being shipped out of the State, mainly to Texas. The great thickness of the Arbuckle and Viola limestones makes them especially attractive for quarrying. A general survey of Oklahoma's carbonate deposits (including Cambrian–Ordovician carbonates) was prepared by Rowland (1972a); Rowland (1972b) also prepared information on chemical and physical properties of stone from selected limestone and dolomite quarries.

Thick dolomite units are present in the Arbuckle Group in the Arbuckle and Wichita Mountains, and these also can be sources of aggregate similar to limestone. In addition, however, abundant reserves of high-purity dolomite are present in the Royer and Butterly Dolomites in the Arbuckle Mountains. These two units are being quarried at three localities for fluxing stone, glass manufacture, refractories, animal feeds, and conventional aggregate uses. The high-purity Royer is mostly 97–99% dolomite and has been described in its major prospect area, the Mill Creek–Ravia area (Ham, 1949).

Large reserves of high-purity silica sand in the Simpson Group are mined hydraulically by two companies in the Arbuckle Mountains (Fig. 5). Crude sand consisting of 98% silica is upgraded to 99.8% silica (and normally only 0.01–0.03% iron oxide), and the product is marketed for glass-making, foundry sands, ceramics, and the manufacture of sodium silicate. Specific sands being worked now are the Oil Creek sand and the McLish sand.

Additional small resources are present in the Burgen Sandstone in northeast Oklahoma, but these outcrops are in the valley of the scenic Illinois River, and probably are unavailable for mining. Ham (1945) prepared a comprehensive study of the glass-sand resources in the central Arbuckle Mountains.

Large reserves of natural rock asphalt are present in many Arbuckle Mountains outcrops where Simpson sands and Viola limestones locally are impregnated with heavy oil. Open-pit mining of these deposits between 1891 and 1960 provided

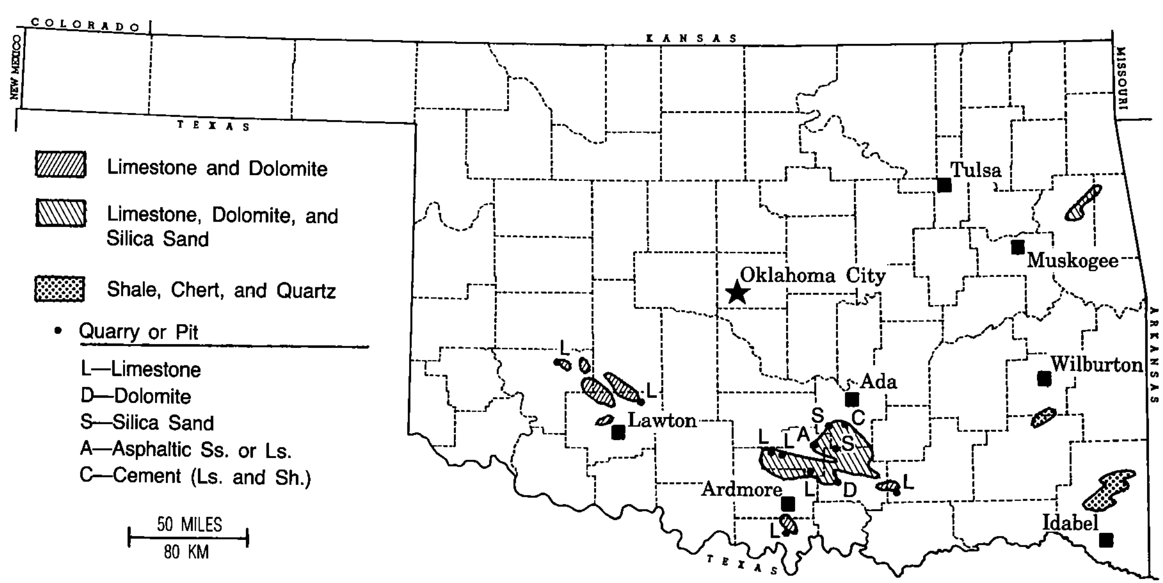

Figure 5. Map showing nonfuel mineral resources in Late Cambrian and Ordovician rocks of Oklahoma (after Johnson, 1969).

a major part of the local road-surfacing materials, but all natural-rock-asphalt pits are now inactive. Processing of the asphaltic Simpson sands now might enable recovery of the heavy oil along with byproduct recovery of high-purity silica sand.

Cement is being manufactured from Sylvan Shale and the Viola and basal Hunton (Keel oolite) limestones south of Ada in the Arbuckle Mountains. In the Wichita Mountains, small deposits of low-grade hematite in the Reagan Formation have been used as a paint pigment, whereas in the Arbuckle Mountains a number of small deposits of limonite (derived by intense weathering of iron-bearing parts of the Arbuckle Group) have been mined as a source of high-grade iron ore. Arbuckle strata in the Arbuckle Mountains host small deposits of manganese oxide associated with a fault zone, as well as many low-grade vein deposits of zinc and lead occurring as sphalerite and galena. Veins and crystals of colorless and milky quartz are present at scattered locations in Cambrian and Ordovician strata of the Ouachita Mountains, but these have not been developed commercially (as they have been in Arkansas). The Bigfork Chert in the Ouachita Mountains may locally be suitable for road material, railroad ballast, construction material, or abrasives. References for most of these mineral deposits are given by Johnson (1969).

GROUND-WATER RESOURCES

Late Cambrian and Ordovician carbonates, sands, and fractured cherts are major sources of fresh ground water in various parts of Oklahoma (Fig. 6): these aquifers include the Arbuckle and Timbered Hills Groups in the Wichita Mountains–Lawton area; the Simpson and Arbuckle Groups in the Arbuckle Mountains; the Roubidoux, Gasconade, and Eminence Formations in the Ozarks region; and the Bigfork Chert in the Ouachita Mountains. The distribution and general character of these aquifers, along with a bibliography of hydrologic reports, are described by Johnson (1983).

The Arbuckle and Timbered Hills Groups aquifer consists of limestone, dolomite, and sandstone in the Wichita Mountains and Lawton area of southwestern Oklahoma (Havens, 1983). Wells commonly yield 25–600 gpm (gallons per minute) of water that is of good to fair quality (generally 300–1,200 mg/L dissolved solids). Most wells are 200–1,500 ft deep; the water is generally under artesian conditions, and it flows at the surface from some wells.

The Simpson–Arbuckle aquifer in the Arbuckle Mountains area consists of limestone and dolomite with beds of uncemented quartzose sand (Fairchild and others, 1990). Wells typically yield 25–600 gpm of water that is of good quality (generally <500 mg/L dissolved solids). Wells commonly are 150–1,500 ft deep.

Roubidoux, Gasconade, and Eminence strata in northeastern Oklahoma comprise an aquifer generally 200–500 ft thick. The aquifer consists mainly of cherty dolomite, with some interbedded sandstone. Wells commonly yield 50–250 gpm, and locally yield as much as 1,000 gpm. Water is of good to fair quality (generally 150–1,500 mg/L dissolved solids). The depth to the aquifer commonly is 500–1,500 ft.

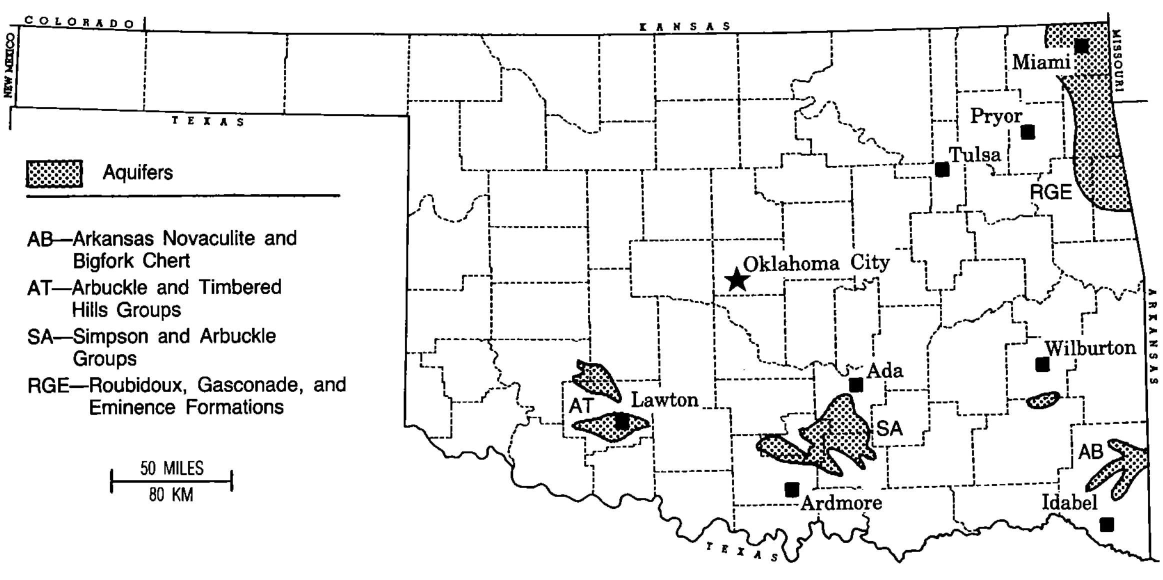

Figure 6. Map showing Late Cambrian and Ordovician aquifers in Oklahoma (after Johnson, 1983).

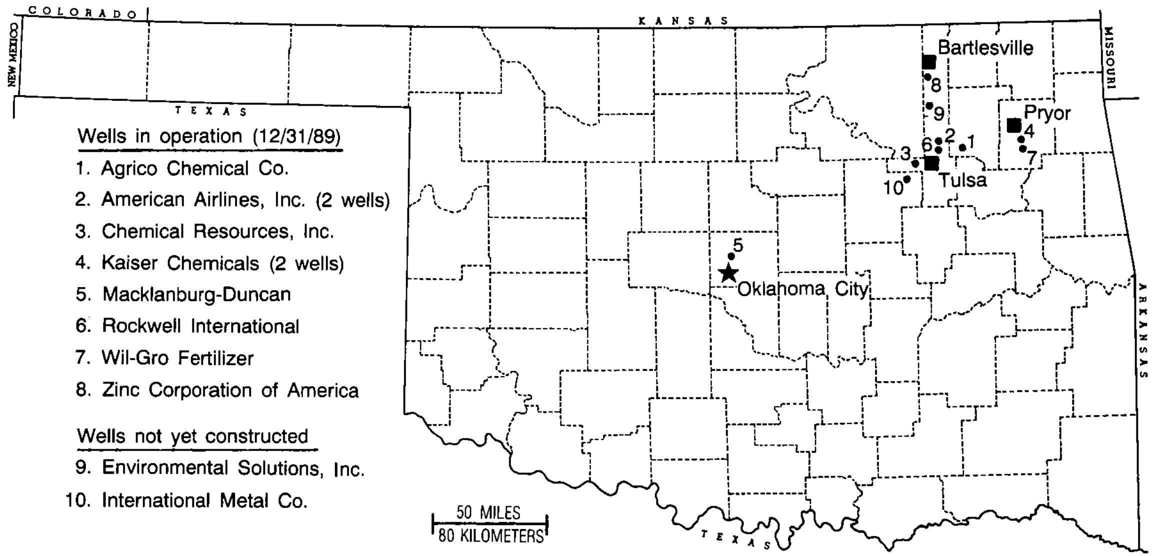

Figure 7. Map showing location of facilities for disposal of liquid industrial wastes in Oklahoma, as of December 31, 1989. Data from Oklahoma State Department of Health, Industrial Waste Division.

The Bigfork Chert (along with the Devonian Arkansas Novaculite) is an aquifer consisting of highly fractured chert, with some interbedded shale and sandstone in the Ouachita Mountains. The combined Arkansas Novaculite–Bigfork Chert aquifer is 850–1,200 ft thick. Wells commonly yield 10–50 gpm of water that is of good quality (generally <500 mg/L dissolved solids).

WASTE DISPOSAL

Liquid-waste materials, such as oil-field brines and industrial wastes, are being disposed of by deep-well injection into porous and permeable zones of the Arbuckle and Simpson Groups at many places in Oklahoma. These strata are used in more than 1,000 salt-water-disposal (SWD) wells handling oil-field brines produced from the State's oil and gas wells, and in 10 wells currently being used to dispose of liquid industrial wastes (Fig. 7). A properly sited and constructed disposal well requires that shale or other confining layers overlie the disposal zone (Fig. 8), in order to prevent waste from migrating to fresh-water zones or to the biosphere. A study of the disposal of industrial wastes in Oklahoma (Johnson and others, 1980) contains data on the potential use of Arbuckle and Simpson strata as host rocks in various parts of the State.

The 1,000–6,700 ft sequence of Arbuckle Group carbonates has intergranular and vuggy porosity, combined with extensive fractures and paleokarst zones, to form reservoirs with porosities of 5–20% and permeabilities ranging from 100 to several thousand md (milliarcies). The depth to the top of the Arbuckle Group is <1,000 ft in the Ozark region of northeast Oklahoma, and it increases to >15,000 ft in the Arkoma basin, and to >30,000 ft in the Anadarko basin (Fig. 9).

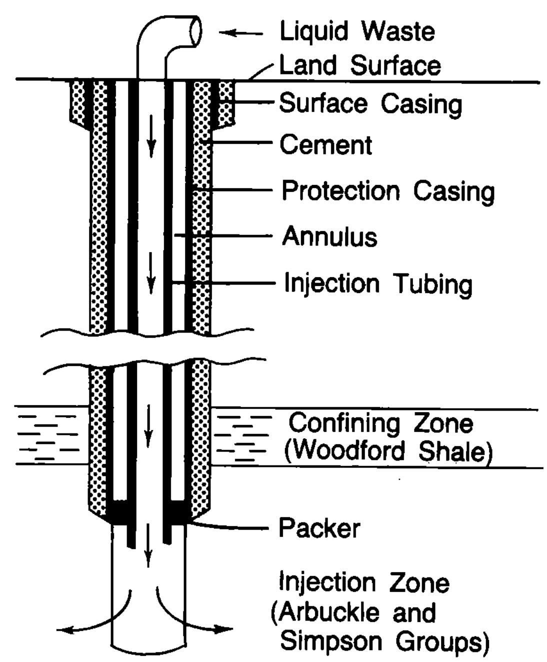

Figure 8. Schematic diagram showing major features of industrial-waste-disposal wells in Oklahoma. Injection zone is 400–6,800 ft below the land surface at various sites.

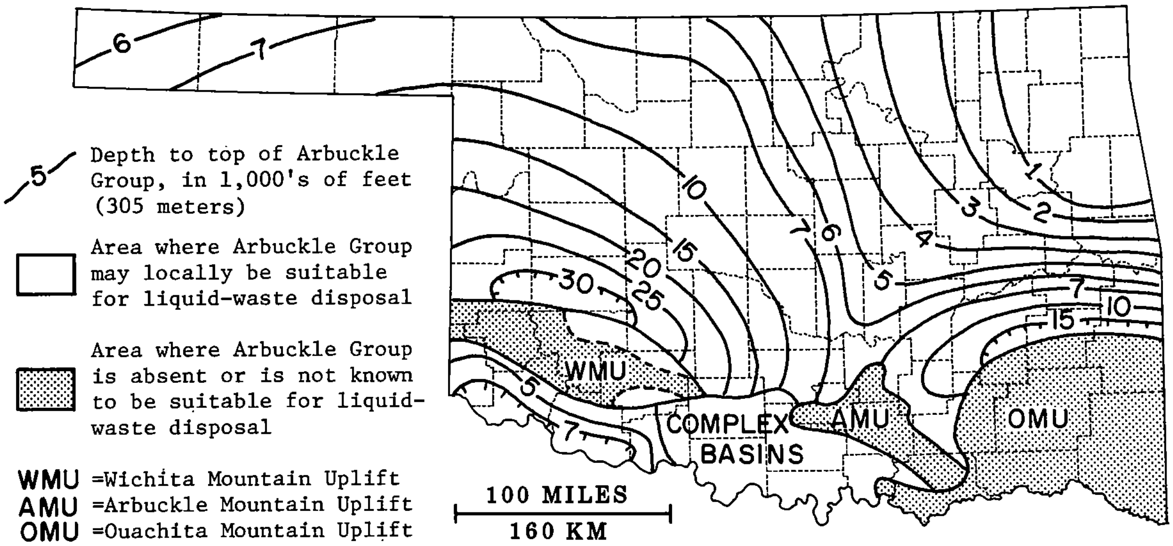

Figure 9. Generalized map showing depth to top of the Arbuckle Group in Oklahoma (from Johnson and others, 1980).

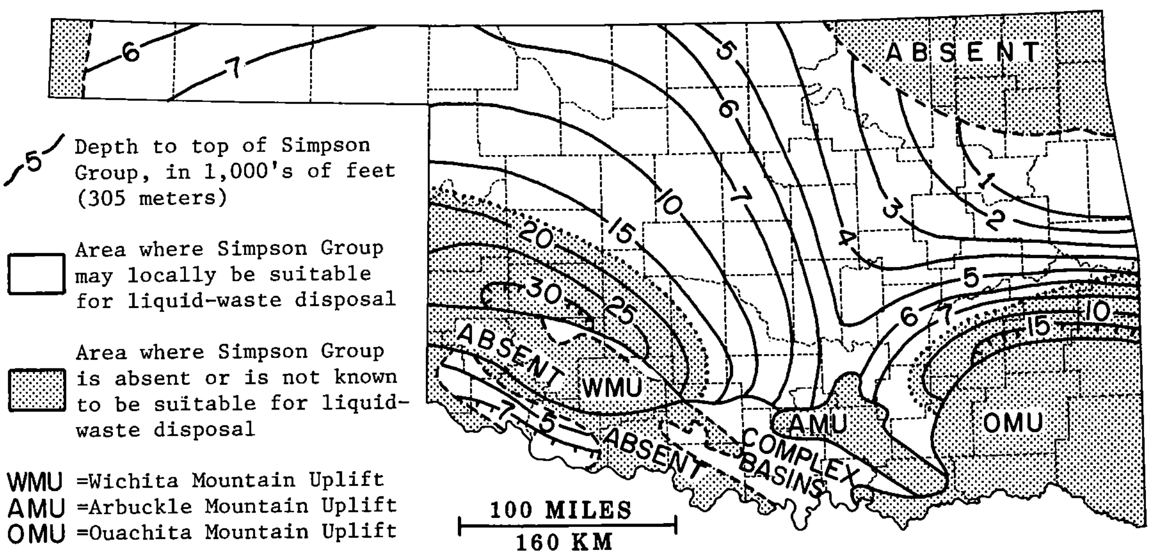

Figure 10. Generalized map showing depth to top of the Simpson Group in Oklahoma (from Johnson and others, 1980).

The Simpson Group is 100–2,300 ft thick and contains several widespread, sheet-like sands and sandstones that consist of fine-grained, well-rounded, clean quartz sand. Individual sands generally are 50–200 ft thick. Intergranular porosity in most areas is 5–20%, and the permeability is several hundred to several thousand md. The top of Simpson strata is <1,000 ft deep in the northeast, is 4,000–7,000 ft deep elsewhere in the north, and plunges to >15,000 and >30,000 ft deep in the Arkoma and Anadarko basins, respectively (Fig. 10). In the deeper parts of the major basins (below 10,000–15,000 ft), compaction and recrystallization have reduced the porosity and permeability significantly.

In addition to the large number of wells used for SWD in the Arbuckle and Simpson Groups, all 10 of the State's hazardous-waste-disposal wells (at 8 different locations) are injecting into Arbuckle and/or Simpson strata. The Woodford Shale is the confining zone that rests upon the Arbuckle or Simpson–Arbuckle waste-disposal zones in the Tulsa, Pryor, and Bartlesville areas of northeast Oklahoma (Figs. 7,8). In the Oklahoma

City area, liquid wastes are injected into Simpson sands ("Second Wilcox") truncated by Middle Pennsylvanian shales. Disposal horizons are 400–900 ft deep in the Pryor area, 1,400–3,300 ft deep in the Tulsa and Bartlesville areas, and 6,700–6,800 ft deep in the Oklahoma City area.

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Oklahoma Geological Survey Circular 92, 1991

BIOSTRATIGRAPHY OF THE TIMBERED HILLS, ARBUCKLE, AND SIMPSON GROUPS, CAMBRIAN AND ORDOVICIAN, OKLAHOMA: A REVIEW OF CORRELATION TOOLS AND TECHNIQUES AVAILABLE TO THE EXPLORATIONIST

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ABSTRACT.—The Timbered Hills, Arbuckle, and Simpson Groups as exposed in southern Oklahoma comprise one of the most complete and best-exposed sequences of Late Cambrian through Middle Ordovician strata in North America. Despite some gaps in knowledge and the need for additional study of some parts of the section, the entire sequence is now well known in terms of biostratigraphic control utilizing diverse fossil groups.

In the 20 years since Stitt and Derby published brief notes in Oklahoma Geological Survey Guidebook 17 revising the position of the Cambrian/Ordovician and Lower/Middle Ordovician boundaries, respectively, biostratigraphic knowledge of the Oklahoma Cambrian–Ordovician section has increased tremendously. Almost without exception, all formations can be subdivided into biostratigraphic zones ranging in thickness from many hundreds to a few tens of feet and accurately correlated with equivalent strata elsewhere. We now clearly recognize that the sequence began in the Franconian (not the Dresbachian), that the currently recognized Cambrian/Ordovician boundary falls within the Signal Mountain Limestone, and that the Lower/Middle Ordovician boundary falls within the West Spring Creek Formation of the Arbuckle Group. Major unconformities separate the lower Simpson Joins and Oil Creek Formations from the overlying McLish Formation, and the uppermost Simpson Group Bromide Formation from the overlying Viola Springs Formation. Numerous minor unconformities fall within the Arbuckle Group. Recognition of the major unconformities suggests that the top of the Sauk sequence is at the top of the Oil Creek Formation of the Simpson Group, not at the top of the Arbuckle Group.

Acid-resistant phosphatic microfossils (conodonts) can be recovered from well cuttings throughout the entire sequence; palynomorphs (acritarchs and chitinozoans) and ostracodes are useful in the Simpson Group. Macrofossils, especially trilobites and brachiopods, generally require well-core or outcrop samples for recovery, but they provide needed control for intervals in which microfossils have long ranges or remain poorly known.

INTRODUCTION

This paper reviews progress in biostratigraphic knowledge of the Timbered Hills, Arbuckle, and Simpson Groups of Oklahoma accomplished during the last two or three decades. Much of this data has not been published formally, or it has been published in journals or monographs not readily available to petroleum and mineral explorationists, and it is thus not obvious to compilers of stratigraphic information. Consequently, many recent summaries of the physical stratigraphy of this interval suffer from failure to include more recent information on age, correlation, and paleoecology. It is hoped that this summary will make the new paleontologic information more accessible. This paper emphasizes information from recent biostratigraphic studies of faunas that are useful for time-stratigraphic correlation. We do not summarize all the literature dealing with taxonomy, paleoecology, and paleogeography (e.g., Lewis, 1982; Ross, 1976; Sprinkle, 1982); readers are urged to search the bibliographies of cited papers to obtain a more complete review of the available literature. Viola Group and younger Ordovician strata are not reviewed; for information on the Late Ordovician, readers are referred to the recent reviews by Amsden (1989), by Amsden and Sweet (1983), and by Finney (1986).

CHRONOSTRATIGRAPHIC CLASSIFICATION

The summary stratigraphic chart (Fig. 1) shows the latest chronostratigraphic classification and the intervals studied by current workers. This classification is that of Ross and others (1982), modified slightly by more recent work; it differs significantly from previous ones in the following:

1) Ibexian is used here in the sense of Ross and others (1982) and replaces Canadian as the series name for Lower Ordovician strata, although the included strata remain largely the same;

2) Whiterockian replaces and includes the "Chazyan"; and

3) "Trenton" is no longer used as a time or series term. Some strata previously considered Trentonian are correlated with Cincinnatian (Upper Ordovician) strata (Sweet and Bergström, 1971).

CORRELATION

Correlation of the Upper Cambrian through Ordovician units in the southern North American Midcontinent is shown in Figure 2. This diagram was compiled by J. E. Repetski and J. R. Derby from numerous sources and with the advice of co-authors. Not all co-authors fully agree with all details of correlation; inevitably, additional work will resolve these few differences. The northeast Oklahoma column was compiled from data in Bauer (1989), Yochelson and Bridge (1958), and Kurtz and others (1975). Included in the diagram is the stratigraphic column for the Ibex area of western Utah. This relatively complete and well-exposed sequence is well known through the many works of L. F. Hintze (see Hintze, 1973,1979) and others (see Miller and others, 1982; Ethington and Clark, 1981). The Ibex area is becoming increasingly important in North America as a reference section for Lower and Middle Ordovician biostratigraphic correlation.

BIOSTRATIGRAPHIC METHODS

Modern biostratigraphy rests on initial rigorous and objective collection of data. Well-exposed rock sequences are located and then systematically collected. Closely spaced samples of bulk rock are taken for later processing for microfossils; macrofossils are found by careful examination of the weathered surfaces and by rock-breaking.

Some macrofossils are phosphatic or commonly silicified and can be extracted by dissolving the rock in acid. In the Arbuckle Group, articulate brachiopods and mollusks are commonly silicified and, therefore, are relatively easily extracted using hydrochloric or other acids. Inarticulate brachiopods are phosphatic and can be extracted with organic acids (acetic or formic acid). Unfortunately, trilobites, one of the most useful groups for biostratigraphy, are rarely silicified and must be extracted by use of rock splitters, grinding wheels, and small percussion tools such as vibro-engravers. Because of their size, macrofossils are difficult to recover from drill cuttings, although cuttings of the Simpson Group biosparites commonly yield identifiable trilobites, ostracodes, and brachiopods. Ostracodes are traditionally considered microfossils, but many Ordovician species are truly macrofossils, for they reach lengths >1 cm. Ostracodes are recoverable from limestones by mechanical means, and they also can be washed from shales.

Acid-resistant microfossils are readily recoverable from drill cuttings, as well as from core and outcrop. True conodonts are phosphatic microfossils; they are found abundantly in marine rocks of Late Cambrian through Triassic age. Conodonts are relatively easily removed from carbonates by organic acids. They are large enough (most are about 0.1–1 mm) to be studied under a low-power binocular microscope; they have proven extremely useful for correlation in the uppermost Cambrian and throughout the Ordovician. Acid-resistant, organic-walled microfossils (acritarchs and chitinozoans) are considerably smaller than conodonts and require palynological techniques for extraction and study. Acritarchs occur commonly in late Precambrian through Devonian strata, but are less common in younger geological systems. Chitinozoans appeared in the Early Ordovician and became extinct in the Late Devonian. They also appear to have been largely destroyed by diagenesis in the carbonate rocks of the upper part of the Arbuckle Group but are recovered in good numbers from shaly rocks of the Simpson Group.

After extracting and identifying fossils, the modern biostratigrapher plots fossil occurrence relative to the rock sequence. Several examples of such "range charts" are shown in the figures. If the fauna is abundant and/or well understood, simple

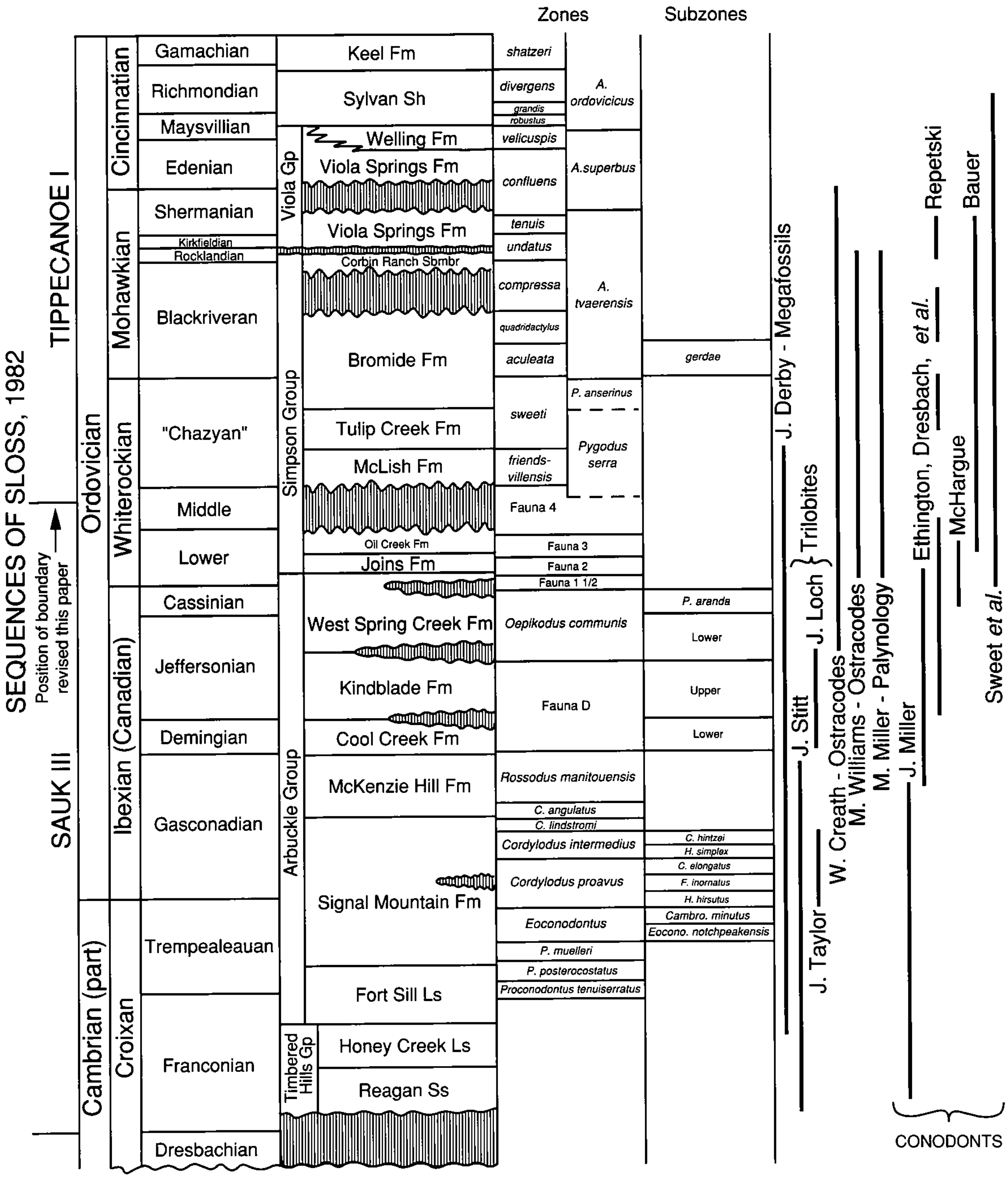


Figure 1. Chronostratigraphic classification of Late Cambrian through Ordovician strata of the Arbuckle and Wichita Mountains, Oklahoma, with conodont zonation and intervals studied by the authors. The Royer and Butterly Dolomites are intentionally omitted, see Stitt (1977,1983) and Ham (1955,1973) for their position. Compiled by J. E. Repetski and J. R. Derby for this paper.

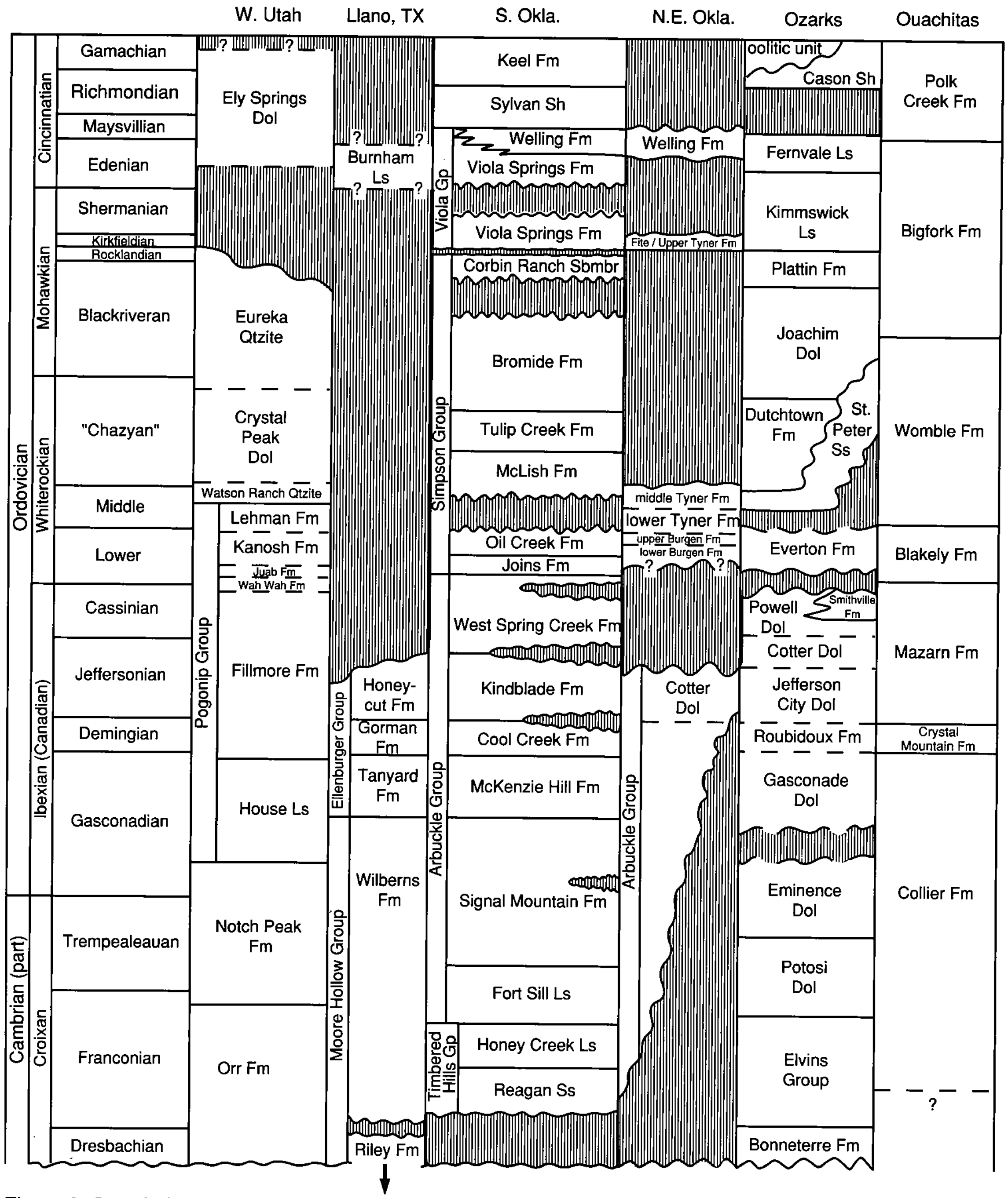


Figure 2. Correlation of the southern Oklahoma Arbuckle and Wichita Mountains section with sections in the Ibex area of western Utah, the Llano uplift of Texas, northeastern Oklahoma outcrop and subsurface, the Ozark region of Missouri and Arkansas, and the Ouachita Mountains of Oklahoma and Arkansas. Compiled for this paper by J. E. Repetski and J. R. Derby, with the advice of co-authors.

comparison of plotted ranges in a well-known section compared to the "new" section reveals the correlation (Fig. 3). A more sophisticated technique is the graphic correlation or "cross-plotting" technique of Shaw (1964; see detailed discussion by Sweet, in Amsden and Sweet, 1983, p. 26–32), in which faunal tops and bases in one section are plotted against those in another section on a conventional x–y graph. Figure 4 (from Taylor, 1984) shows how an unconformity is readily recognized on these cross-plots. The graphic-correlation technique provides a powerful and objective tool for evaluating unknown sequences and for measuring the magnitude of a hiatus or the throw of a fault. Use of the graphic technique enables compilation of a composite reference, or standard, section in which the combined ranges of all correlated species are compiled.

REVIEW OF BIOSTRATIGRAPHIC STUDIES

Studies by the authors began in 1964 with the work of Stitt on the Cambrian and with the work of Derby and Creath on the Signal Mountain Formation through Simpson Group. Stitt has continued detailed studies of the trilobites from the lower part of the sequence; Derby and Creath emphasized the upper part of the Arbuckle Group and lower part of the Simpson but ended their major research efforts in the early 1970s. Early results (Stitt, 1969; Derby, 1969) and recognition of the biostratigraphic utility of conodonts stimulated additional studies on the well-exposed, thick, largely carbonate sequences of the Timbered Hills through Simpson Groups. By the 1970s, three principal centers of conodont studies had emerged: J. F. Miller focused on the Cambrian–earliest Ordovician interval, Ethington and his students on the Early and earliest Middle Ordovician, and Sweet and Bergström and their students on the Middle and Late Ordovician. Macrofossil studies of the 1960s and early 1970s were reviewed by Derby and others (1977); the detailed range charts from that review are published here for the first time. More recently, M. A. Miller of Amoco Research has been studying marine palynomorphs (acritarchs and chitinozoans) of the Simpson Group, and M. Williams is restudying the ostracodes of the Simpson Group. Other current studies include those of Dresbach and Ethington

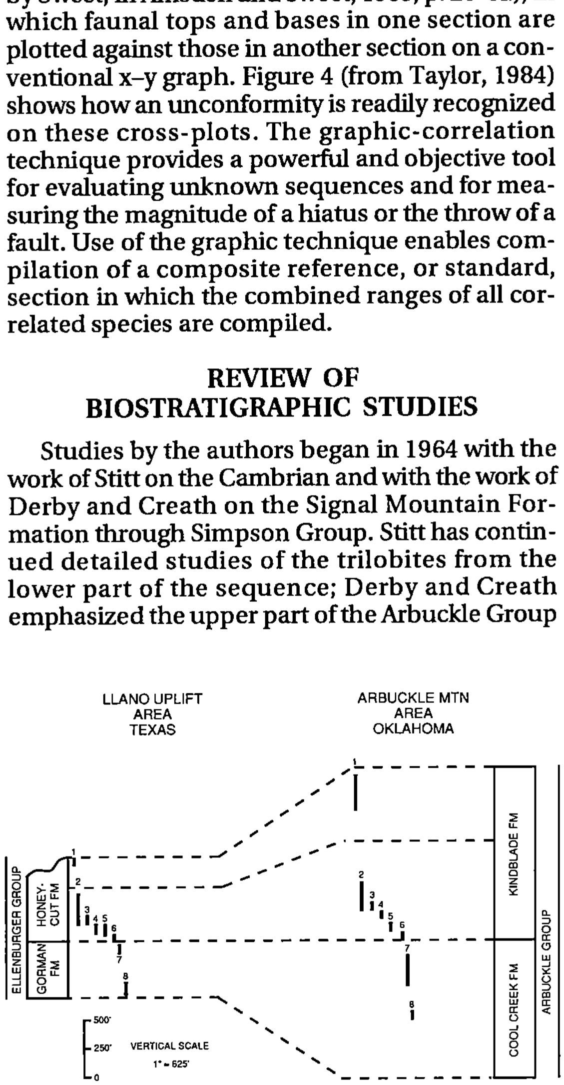


Figure 3. An example of correlations based on plotted ranges of species in common to two sections, showing correlation of the Gorman and Honeycut Formations of the Llano uplift of Texas with the Cool Creek and Kindblade Formations of the Arbuckle Mountains of Oklahoma. Compiled by J. R. Derby from data in this paper, from Bridge and Cloud (1947), Cloud and Barnes (1948), Cloud (1948), Yochelson and Bridge (1958), and Derby (in Toomey, 1964). Species are (1) *Ceratopea tennesseensis*, (2) *Ischyrotoma abruptus*, (3) *Jeffersonia producta*, (4) *Ceratopea capuliformis*, (5) *Ceratopea corniformis*, (6) *Xenelasma syntrophoides*, (7) *Diaphelasma oklahomense*, (8) *Rhombella umbilicata*.

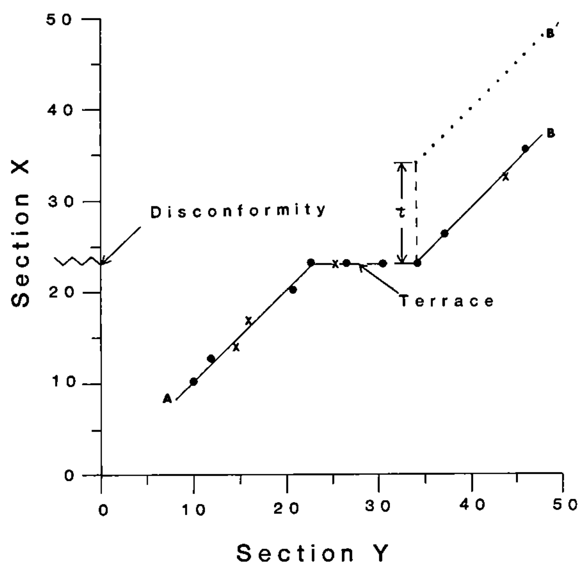


Figure 4. An idealized "cross-plot" or bivariate graph in which a section containing a disconformity (section X) is plotted against a section with no significant breaks (section Y). Solid dots represent the lowest occurrences of species, x's represent the highest occurrence of species. A "best-fit" line through the tops and bases is called a "line of correlation." The section missing at the disconformity causes a terrace in the line of correlation. The line of correlation below the disconformity (A) and the line of correlation above the disconformity (B) have the same slope; consequently, the thickness of section (t) missing from section X owing to erosion and/or nondeposition can be determined by raising line B to the position shown as B'. Both scales are in feet above base of section. From Taylor (1984, fig. 26).

on Early Ordovician conodonts, Loch on Kindblade Formation trilobites, Repetski on conodonts from the subsurface, and Bauer on Middle Ordovician subsurface and outcrop conodonts.

REVIEW OF BIOSTRATIGRAPHIC INFORMATION AND STRATIGRAPHIC CONCLUSIONS

Cambrian and Lowest Ordovician

Stitt (1971,1977,1983) described and illustrated trilobites from the basal Timbered Hills Group through the McKenzie Hill Limestone (Fig. 5; only the most abundant are shown). His evidence indicates that deposition of the Reagan Sandstone did not begin in Oklahoma until Franconian time (equivalent to the Wilberns Formation of Texas), and he clarified the relationship between the Royer and Butterly dolomite units and equivalent limestone formations (see Stitt, 1977, fig. 2, p. 6; 1983, fig. 2, p. 6; note the change in placement of the McKenzie Hill/Signal Mountain formational boundary). Derby and others (1972), utilizing data from the Arbuckle Mountains in making comparisons with other sections, demonstrated for the first time that conodont, brachiopod, and trilobite ranges can be used for precise correlation in the Cambrian/Ordovician boundary interval across western North America. Subsequently Derby, Stitt, and J. F. Miller (see Derby, 1986) joined efforts, and their combined data on the boundary interval led to additional studies and refinements in correlation, including recognition of unsuspected faults and consequent errors in correlation. Taylor (1984) studied the macrofauna and detailed lithostratigraphy of the boundary interval in both Oklahoma and Texas.

J. F. Miller (1988; Miller and others, 1982) has continued to study the conodonts of the boundary interval in the North American Midcontinent and worldwide, and has demonstrated that in this interval the North American Midcontinent conodont zonation can be recognized on several continents. Brachiopod ranges by Stitt and conodont ranges by Miller, shown in Figure 6, are based on collections from the Wichita Mountains. Similar detailed range charts for both conodonts and trilobites have been published for central Texas and the Ibex area of Utah (Miller and others,1982). Kurtz (1981) demonstrated by use of conodont ranges that the uppermost Eminence Dolomite of Missouri is Early Ordovician, not Cambrian. Recently, J. F. Miller and Stitt (in the informally published Circular 26 of the I.U.G.S. International Working Group on the Cambrian–Ordovician Boundary, 1990) reported recognition of a 63-ft omission of strata across an offset within a key reference section in the Llano uplift of Texas; these authors also corrected the consequent errors in intercontinental correlation. The error was made in the 1940s and was recognized because the missing interval represents three conodont zonal units. The ability to correlate such thin intervals of strata demonstrates the power of the correlation tools now in use. Because of the recent intense study of the Cambrian/Ordovician boundary interval, many stratigraphic sequences can now be correlated with great confidence to intervals as thin as a few feet.

Lower Ordovician

Conodonts

Lower Ordovician conodont faunas have been studied by Ethington and colleagues for some years (Ethington and Repetski, 1984; Ethington and others, 1987). The Lower Ordovician conodonts of the Arbuckle Mountains are currently being restudied by Dresbach and Ethington (this volume; Ethington and Dresbach, 1990). With the exception of the dominantly shoal-water Cool Creek Formation, conodonts are relatively abundant throughout the Arbuckle Group. Although most of the Arbuckle Group has not received the intense study that the Cambrian/Ordovician and Lower/Middle Ordovician boundary intervals have, those detailed studies suggest that the rest of the sequence will be correlated with a high degree of precision when the faunas have been fully documented. McHargue's (1981) detailed study of conodonts and lithofacies at the top of the Arbuckle will be discussed below.

Macrofossils

Stitt's range chart (Fig. 5) includes the most common trilobites known from the McKenzie Hill Limestone; not presented here are ranges of less-abundant trilobites and six brachiopod species reported by Derby (in Derby and others, 1977) and by Cooper (1952). Both brachiopods and trilobites indicate general correlations with the Tanyard Formation of Texas and the Gasconade Dolomite of the Ozarks, but lack of detailed studies in the latter areas precludes detailed correlations at this time.

The Cool Creek Formation (Fig. 7) is sparsely fossiliferous and was collected only in a reconnaissance manner by Derby in the 1960s and by the late W. E. Ham during the course of his geologic mapping. Six brachiopod species, one trilobite, and one gastropod are reported. The latter, *Rhombella umbilicata,* is well known in the Roubidoux Formation of the Ozarks (Heller, 1954) and the Gorman Formation of central Texas (Bridge and Cloud, 1947). *R. umbilicata* and the brachiopod *Diaphelasma oklahomense* suggest correlation of the upper two-thirds of the Cool

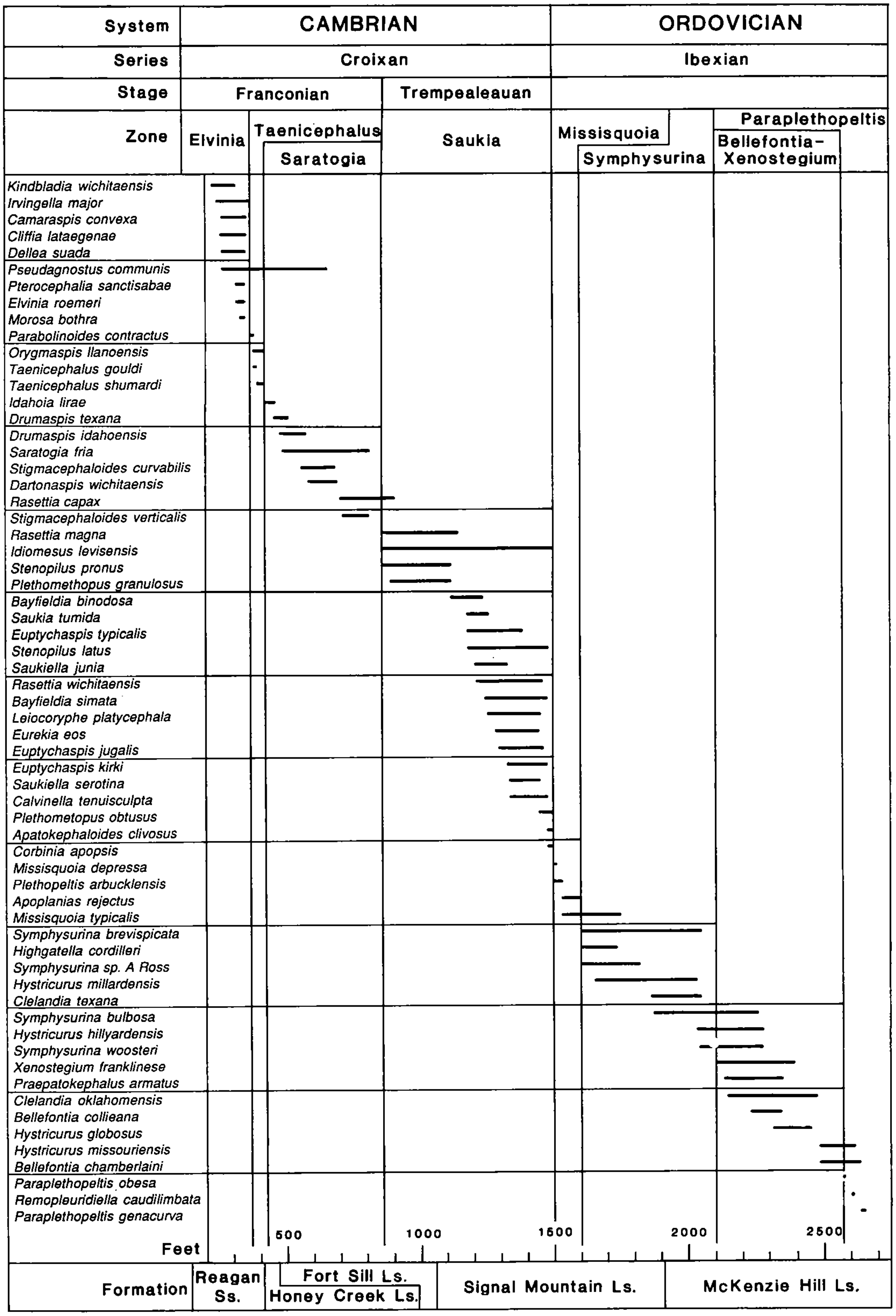


Figure 5. Composite ranges of the most common trilobites from the Reagan through McKenzie Hill Formations, Arbuckle and Wichita Mountains, Oklahoma. Plotted to footages in the Chandler Creek section of Stitt (1977,1983), Wichita Mountains. Compiled by J. H. Stitt for this paper.

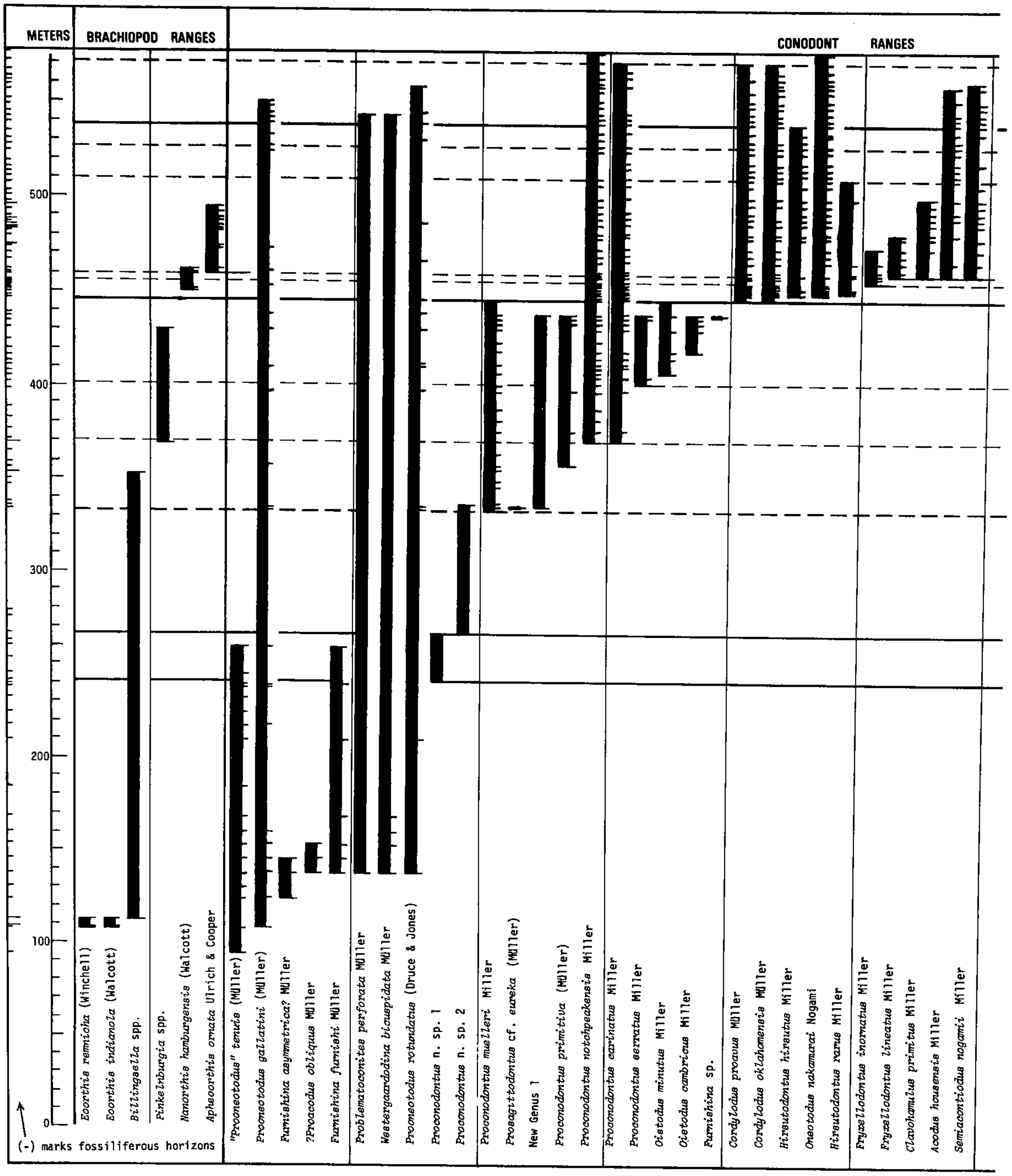
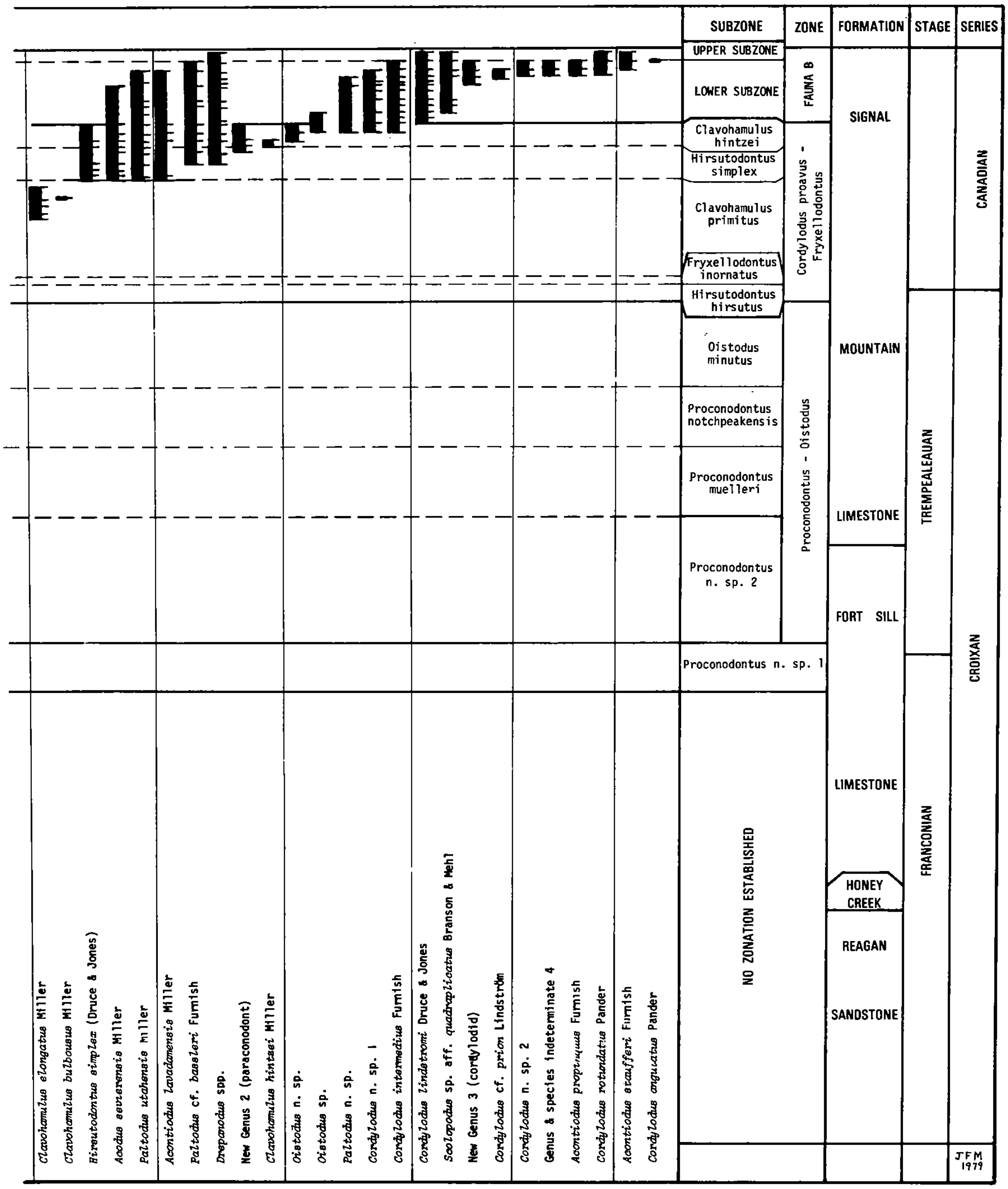


Figure 6. Conodont and brachiopod ranges in the Reagan through Signal Mountain Formations, Chandler Creek section, Wichita Mountains, Oklahoma. Conodont ranges based on 13,467 specimens identified by



J. F. Miller from 91 productive samples. Brachiopods identified by J. H. Stitt. Tick marks on solid bars indicate horizons of samples containing the identified species. From J. F. Miller and others (1982).

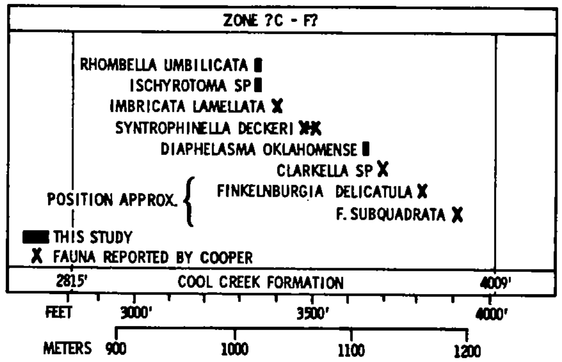


Figure 7. Gastropod, trilobite, and brachiopod fauna of the Cool Creek Formation, Highway 77 section, Arbuckle Mountains, Oklahoma (see Fig. 12), from Derby and others (1977). The datum for footages on this and following diagrams by these authors is the base of the Reagan Sandstone near the Highway 77 section of Stitt (1971), which is nearly the same as the Chapman Ranch section of Derby and others (1972). The higher beds were measured immediately west of the old highway. Alphabetic zonal nomenclature refers to the trilobite zones of Hintze (1953) and Ross (1949,1951).

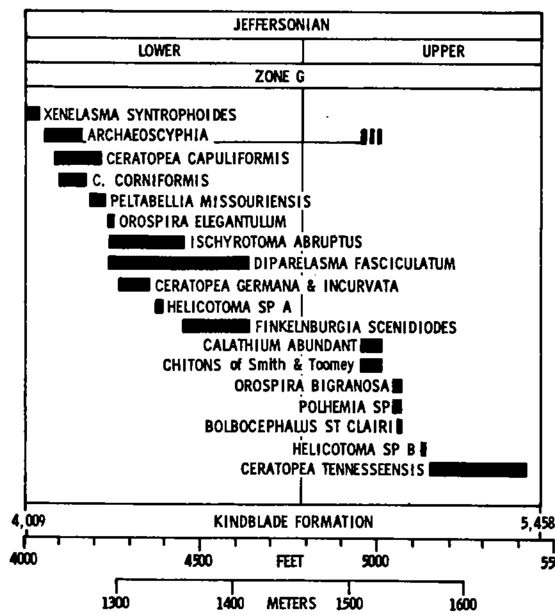


Figure 8. Brachiopods, sponges, mollusks, and trilobites from the Kindblade Formation, Highway 77 section, Arbuckle Mountains, Oklahoma (see Fig. 12). *Archaeoscyphia* from Toomey (1964); *Ceratopea* in part from Yochelson (1973); chitons from Smith and Toomey (1964); the remainder is original work by Derby. From Derby and others (1977). See Figure 7 for explanation of footage and zones.

Creek Formation with the Gorman Formation of Texas (Fig. 3).

The Kindblade Formation has a diverse fauna of brachiopods, gastropods, chitons, trilobites, and sponges. Figure 8 shows the species known as of 1977. J. D. Loch had identified an additional 14 trilobite species by 1989 (Figs. 9,10), and currently reports 30 species, of which one-third are new. The trilobite and brachiopod fauna is similar to that identified by Derby from the sponge beds in the Honeycut Formation of central Texas (Toomey, 1964, p. 102), and the overall fauna is very similar to that of the Jefferson City Formation of the Ozarks (Cullison, 1944). The large gastropod *Ceratopea*, whose large, often silicified operculum is readily identifiable in the field, makes its first appearance in the Kindblade and ranges up through the Lower Ordovician part of the West Spring Creek Formation (Yochelson and Bridge, 1958; Yochelson, 1973). *Ceratopea* is common in the shoal-water dolomites of the Ozarks and facilitates correlations with those facies.

In 1977, Derby (Derby and others, 1977) correlated the lower Kindblade fauna with those of the lower part of the Jefferson City Dolomite of the Ozarks and the lower part of the Honeycut Formation of Texas (Figs. 2,3). The upper Kindblade fauna indicates correlation with the upper part of the Honeycut and with the uppermost part of the Jefferson City and at least the basal part of the Cotter Dolomite of Missouri.

The West Spring Creek Formation, which, except for its uppermost ~100 ft in the Arbuckle Mountains, is clearly of Early Ordovician age, and has a diverse fauna of brachiopods, large ostracodes, trilobites, and gastropods (Fig. 11). Most of the brachiopods had been earlier reported by Ulrich and Cooper (1938); knowledge of the trilobites and ostracodes resulted from the work of Derby and Creath. Derby (1973b) subdivided this Early Ordovician fauna into two informal units and commented in detail upon their correlation; little new can be added. The lower fauna is poorly known in the Midcontinent, but the upper fauna is well represented in the Smithville Formation and its Black Rock Limestone Member in northeastern Arkansas. This uppermost Lower Ordovician fauna, characterized by *Ceratopea unguis* and large species of *Syntrophopsis*, is widely known in North America, but only in settings marginal to the craton where the section is more complete. As shown in Figure 11, nearly all of this macrofauna is abruptly terminated and is replaced by a fauna of Middle Ordovician (Whiterockian) age; the replacement is within the thick-bedded limestones that clearly belong to the Arbuckle Group. The key element of this fauna, the brachiopod *Desmorthis nevadensis* Cooper, had never been found below the (Whiterockian) *Anomalorthis* Zone of the western United States. This observation prompted

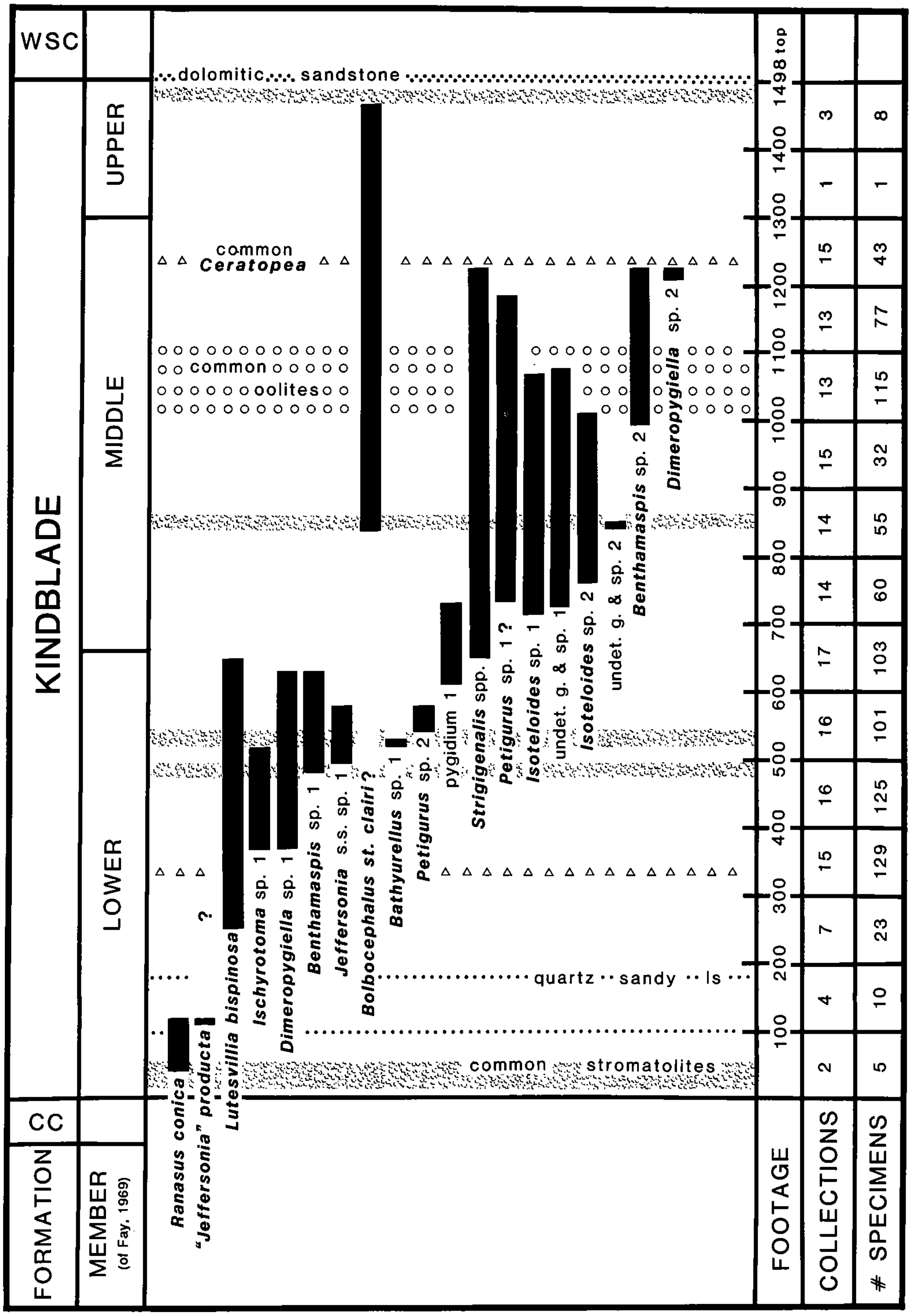


Figure 9. Preliminary trilobite ranges for the Kindblade Formation, Arbuckle Mountains, from a current study by J. D. Loch. Measured section east of northbound lane of Interstate 35, beginning at and continuing south from scenic turnout on the south flank (see Fig. 12). Two informal trilobite faunas are recognized, based on large trilobites easily identified in the field. The *Lutesvillia bispinosa* fauna, in the lower member, is easily recognized by the two-spined pygidium of that species. The *Petigurus–Bolbocephalus* fauna is easily recognized by the large inflated cranidia of the two genera (see Fig. 10). Compiled for this paper by J. D. Loch.

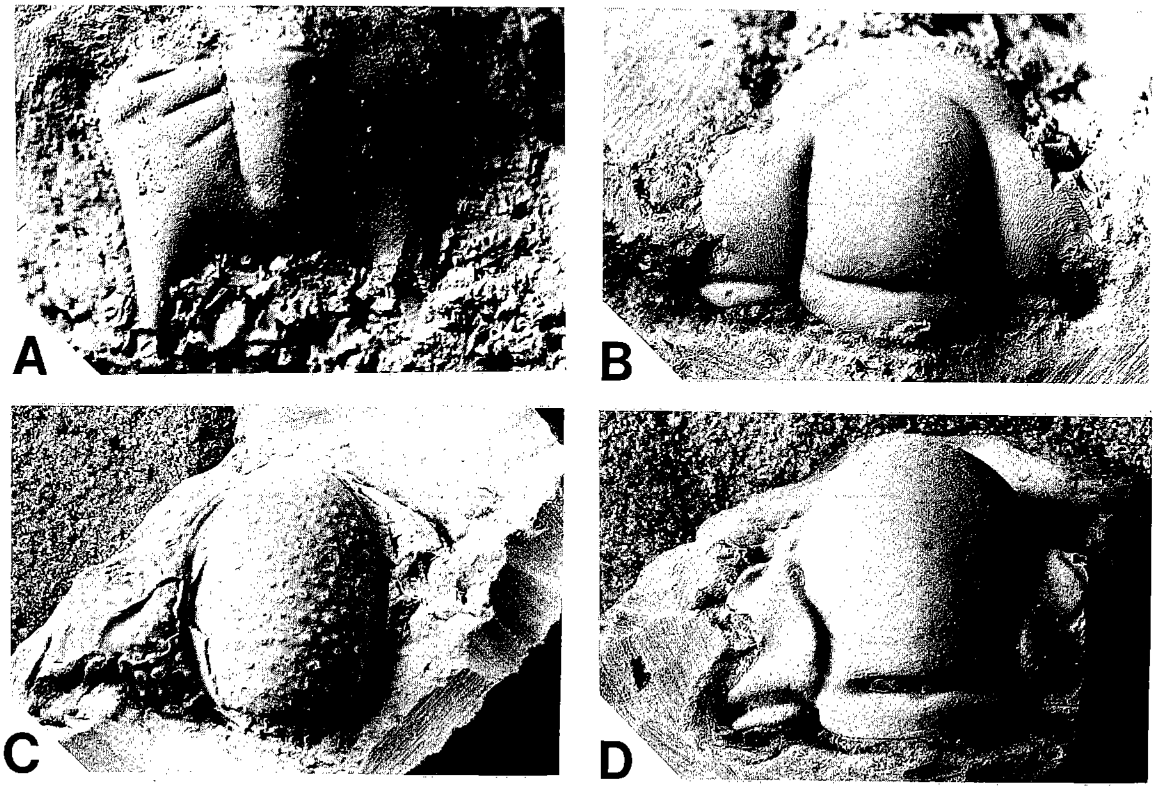


Figure 10. Trilobites of the Kindblade Formation. *A—Lutesvillia bispinosa* Cullison, 1944; pygidium from 556 ft above the base of the formation; ×5.2. *B—Benthamaspis* sp. 1; cranidium from 499 ft; ×4.7. *C—Petigurus* sp. 1; cranidium from 1,127 ft; note coarse pustules on cranidium; ×2.3. *D—Bolbocephalus stclairi* Cullison, 1944; cranidium from 1,135 ft; ×2.1. Prepared for this paper by J. D. Loch.

Derby (1973b) to postulate an unconformity within the West Spring Creek Formation to account for the missing (lower Whiterockian) *Orthidiella* Zone. Ethington and Dresbach (1990, p. 37) reported a conodont change that occurs 80 ft below the top of the West Spring Creek on Interstate 35 (Fig. 12). Consequently the top of the Lower Ordovician Series appears to be within the Arbuckle Group, not at its top.

Lower/Middle Ordovician Boundary Study of McHargue

McHargue (1981) performed a detailed, integrated stratigraphic analysis of three sections in the Arbuckle anticline (Fig. 12), utilizing conodonts for biostratigraphic control. The western section is along West Spring Creek, the center section is along Highway 77 (a short distance east of Interstate 35), and the eastern section is ~1 mi west of Sycamore Creek. McHargue recognized an unconformity within the Arbuckle Group, but not at the Arbuckle–Simpson (West Spring Creek–Joins) contact, based on lithostratigraphic and diagenetic evidence (Fig. 13). Detailed analysis of the conodont faunas and cross-plotting the local first occurrences of the more significant species (Fig. 14) verified the presence of a minor unconformity and showed that it increases in magnitude westward and separates the previously reported Lower and Middle Ordovician faunas; a minor local unconformity within the Middle Ordovician portion was also postulated (Fig. 15). McHargue questioned whether the supposedly missing *Orthidiella* Zone might be present in the more complete eastern section, or whether the apparently missing zone is due to incomplete knowledge of the faunas. At present, that question awaits further study of both conodont and macrofossil ranges in Oklahoma and elsewhere; however, the apparent hiatus appears relatively minor.

Middle Ordovician

Middle Ordovician faunas are more diverse than older ones, and the lithologies yield their fossils more easily; consequently, many workers have described Simpson faunas. The stage for

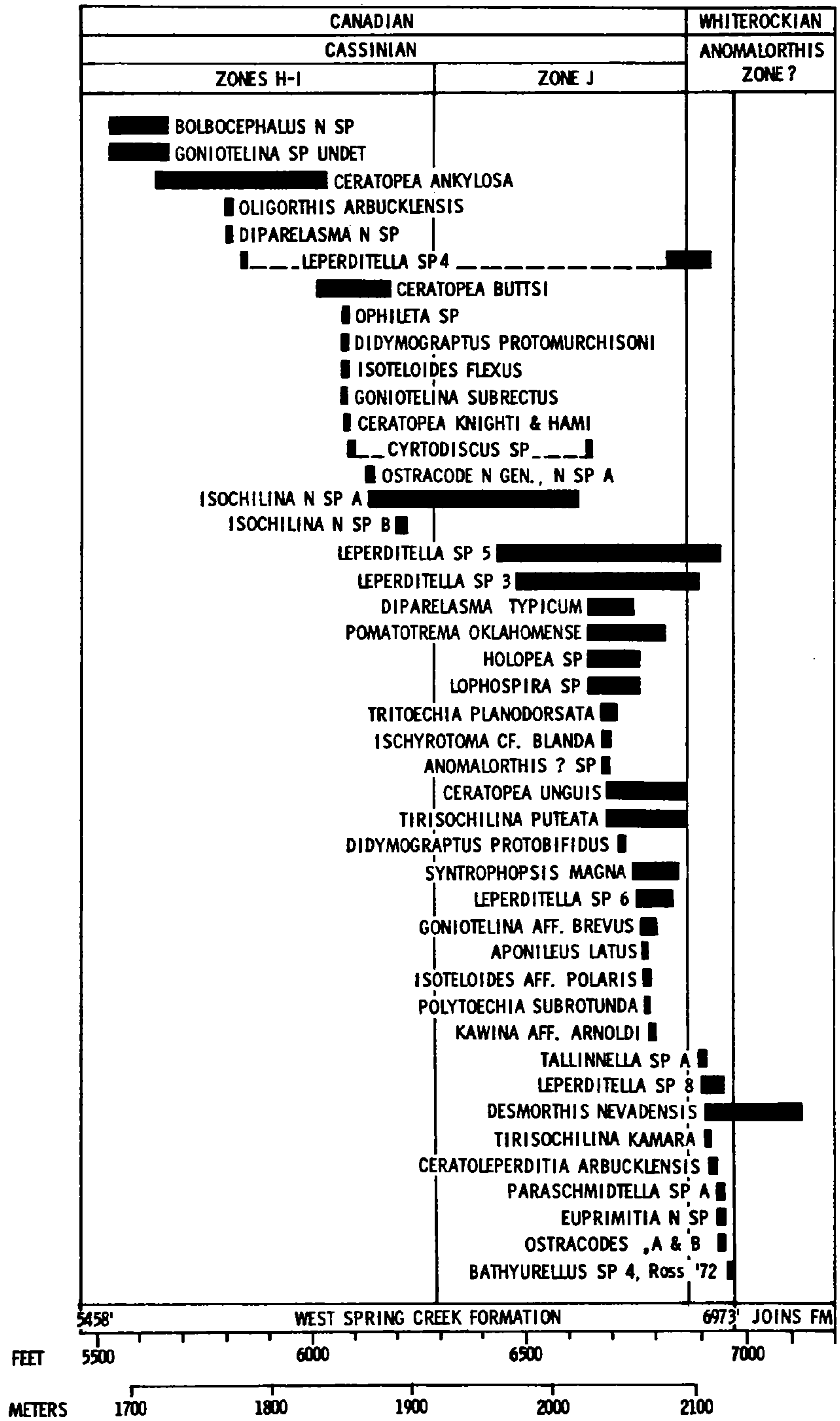


Figure 11. Trilobites, graptolites, brachiopods, and ostracodes of the West Spring Creek Formation: composite ranges for the Highway 77 and West Spring Creek sections, western Arbuckle Mountains, Oklahoma (see Fig. 12). From Derby and others (1977); all data from original collections by J. R. Derby and W. B. Creath. See Figure 7 for explanation of footage and zones. *Anomalorthis* Zone is that of Cooper (1956).

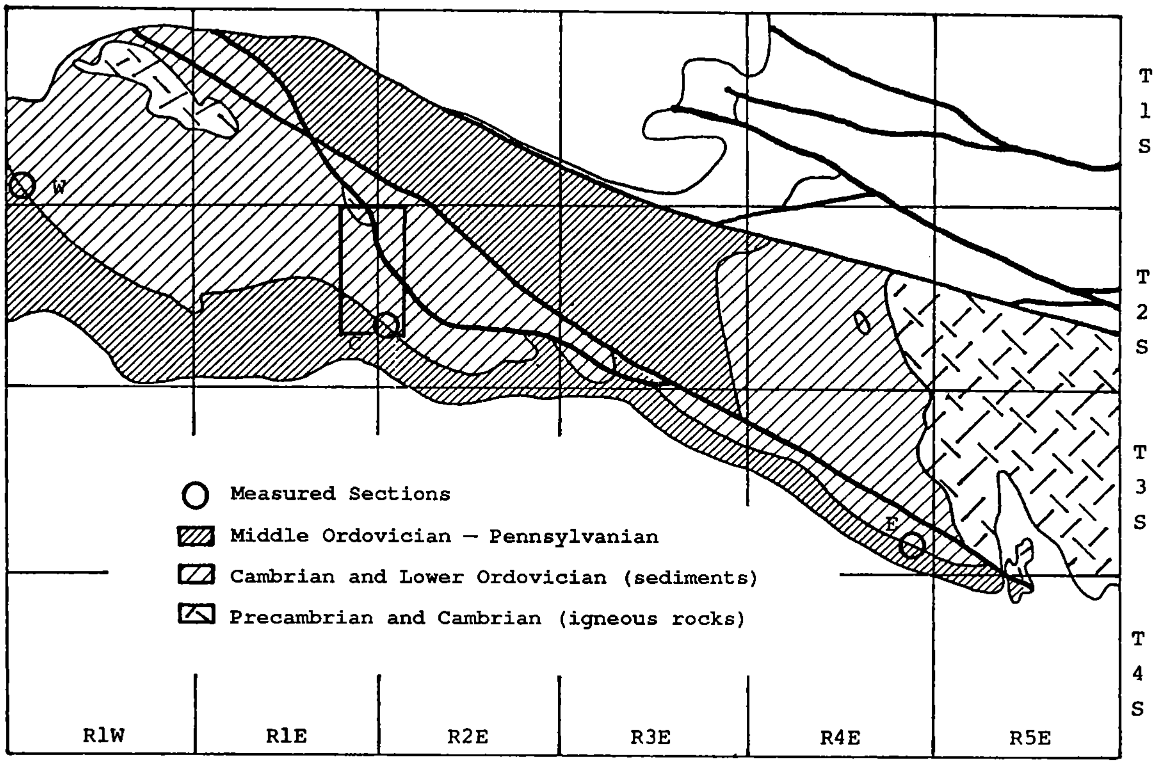


Figure 12. Generalized geologic map of the westernmost Arbuckle Mountains, Oklahoma, showing measured sections of uppermost West Spring Creek and basal Joins Formations of McHargue (1981). The western (W) and central (C) sections are also the West Spring Creek and Highway 77 sections, respectively, of Derby and Creath. The eastern (E) section is 1 mi west of Sycamore Creek. Old Highway 77 essentially follows the range line between R. 1 E. and R. 2 E.; the newer Interstate 35 lies a few hundred feet to a few hundred yards to the west. The south-flank sections studied by Ethington and Dresbach (1990), Grahn and Miller (1986), Loch (this paper), Stitt (1971,1983), and Sweet (1984) are all located within the rectangle. From McHargue (1981).

modern studies was set by Decker and Merritt (1931), who described the entire sequence. Cooper (1956) described brachiopods, and Harris (1957) described the ostracode fauna largely from material washed from shales. More recently, Sprinkle, his students, and others (Sprinkle, 1982) have contributed greatly to our knowledge of echinoderms, lithofacies, and paleoecology of the Simpson Group.

As mentioned above and shown in Figures 11 and 18, the top 90 ft of the West Spring Creek at both Highway 77 and West Spring Creek (Fig. 12) contains the Whiterockian brachiopod *Desmorthis nevadensis*, which ranges upward into the Joins Formation, and an associated ostracode fauna, most species of which are restricted to this thin interval. Ethington and Dresbach (1990) reported that hyaline conodont elements which are the dominant elements of the Joins fauna appear 80 ft below the top of the West Spring Creek in their Interstate 35 section. This conodont evidence supports the macrofossil evidence of a Whiterockian age.

Shaw (1974) described Simpson trilobites from his own collections, and from material collected by Creath and prepared, photographed, and informally described by Derby (Figs. 16,17). These trilobites, along with brachiopods, confirmed correlation of the uppermost West Spring Creek, Joins, and Oil Creek Formations with the strata formerly classified as Whiterockian. These strata are now considered to represent the lower and middle Whiterockian (Fig. 1). No species of trilobites ranges from the Oil Creek (Fig. 16) into the overlying McLish (Fig. 17); this suggests a hiatus. McLish trilobites are similar to those found in upper Whiterockian ("Chazyan") strata elsewhere. Other macrofossils from the Simpson Group, especially brachiopods, need systematic collection and restudy to establish their biostratigraphic value.

Creath studied the Simpson ostracodes (Fig.

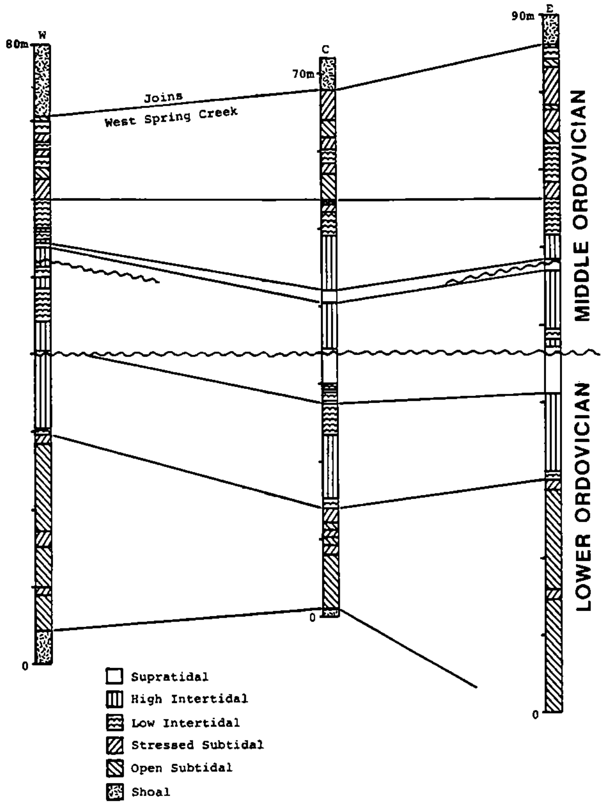


Figure 13. Correlation of depositional environments in the uppermost West Spring Creek and basal Joins Formations by McHargue (1981) showing lithostratigraphic confirmation of an unconformity ~30 m below the top of the West Spring Creek Formation (see Fig. 12 for locations). Lower unconformity used as a datum. Ages as determined by macrofossils added by Derby. From McHargue (1981).

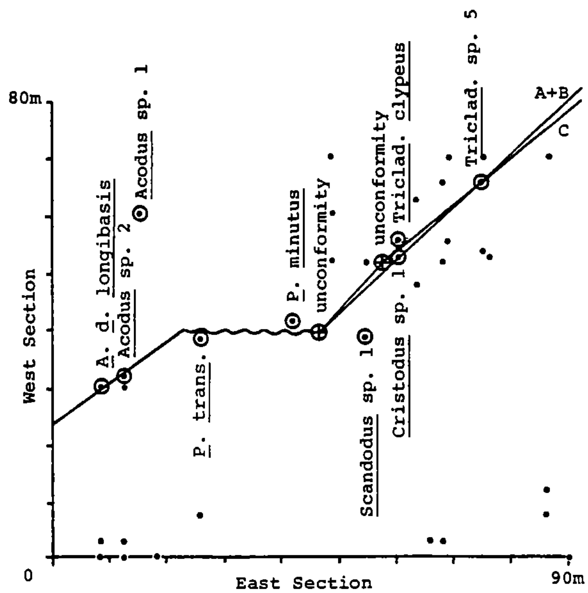


Figure 14. Graphic-correlation cross-plot of first occurrences of conodont species in the uppermost West Spring Creek and basal Joins Formations in the western (vertical axis) and eastern (horizontal axis) sections of McHargue (Fig. 12). Three optional interpretations of the upper part were presented by McHargue (1981). Option A assumes that the first occurrence of *Tricladiodus* sp. 5 is unreliable in the central section (not shown here) and that no upper unconformity is present in either section. Option B uses the first occurrence of *Tricladiodus* sp. 5 but also assumes no upper unconformity (B and C create same line of correlation). Option C (preferred by McHargue; see Fig. 15) uses the first occurrence of *Tricladiodus* sp. 5, but also recognizes an unconformity in the western section. From McHargue (1981).

18) and informally described many new West Spring Creek ostracodes from material mechanically prepared from limestones. He applied this knowledge to many subsurface studies for Amoco, but published only two short papers (Creath, 1966; Creath and Shaw, 1966). His material was made available to Berdan for her study of the leperditicopids of the Ibex area, Utah (Berdan, 1976). Williams is currently redescribing Simpson ostracodes from washed material and had identified ~140 species from the Simpson Group by mid-1989; of these, 25 species are restricted to the Joins–Oil Creek interval, and 32 to the McLish–Bromide interval in the western Arbuckle Mountains. He reports marked differences between the near-shore prolific, but low-diversity, leperditiid-dominated ostracode faunas, and more offshore, high-diversity palaeocopid-rich ostracode faunas. None of the palaeocopid species in the Joins–Oil Creek interval ranges into the overlying McLish Formation.

Derby (1973a,b) suggested that conodont, ostracode, and macrofossil evidence indicates a major hiatus between the Oil Creek Formation and the overlying McLish. In 1973, the state of knowledge was inadequate to document the hiatus quantitatively. Years of conodont studies across the Midcontinent by numerous workers have created the data base for Sweet (1984) to assemble a composite standard reference section and graphically correlate with nearly any North American conodont-bearing section (Figs. 19–21). Figure 19 includes Sweet's graphic correlation of the Interstate 35, Arbuckle Mountains, section. He shows an unconformity with a value of ~130 composite standard section units between the top

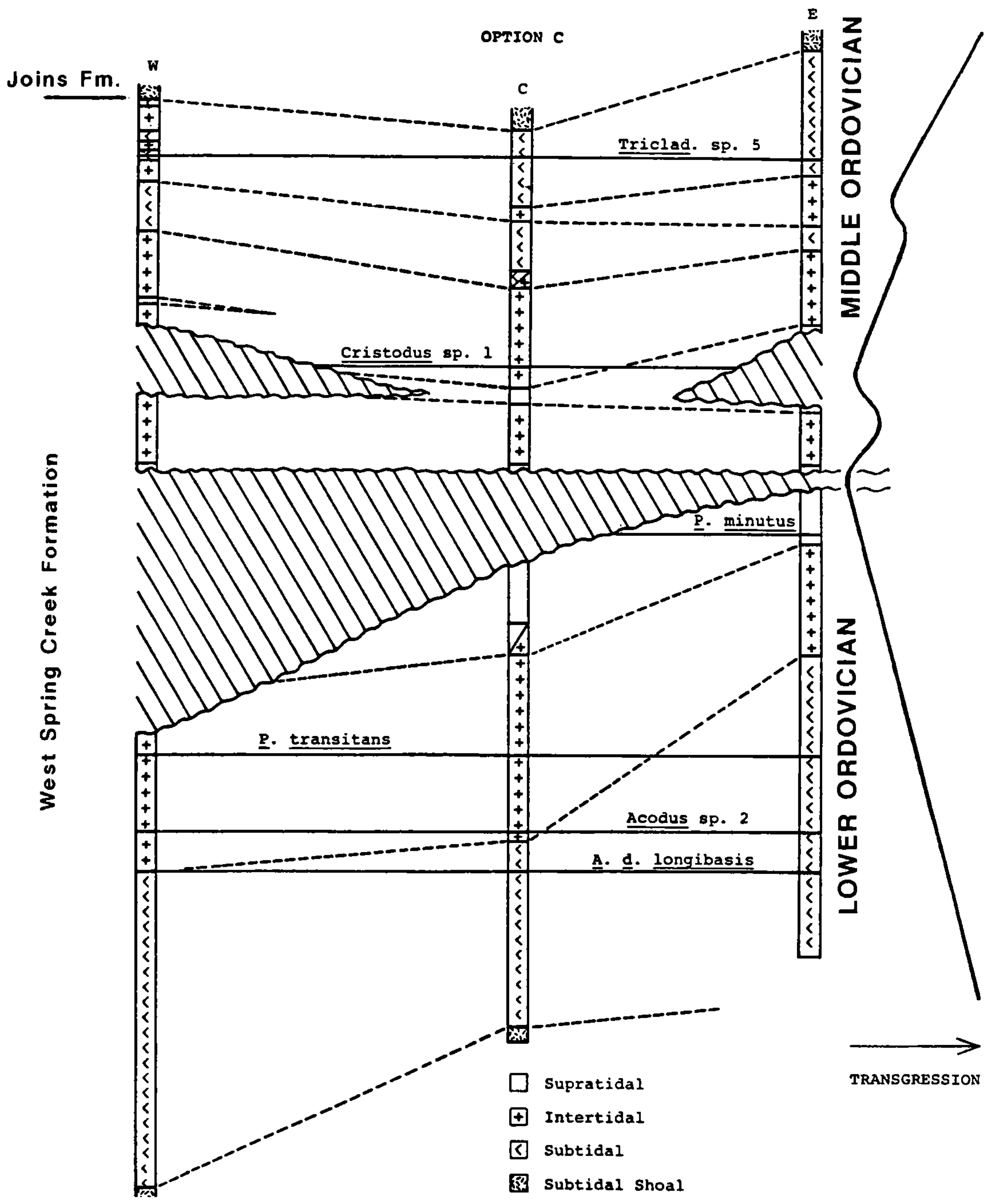


Figure 15. Time-distance cross section of the uppermost West Spring Creek and basal Joins Formations in sections studied by McHargue (1981) in the western Arbuckle Mountains, Oklahoma (Fig. 12). Option C of Figure 14 is shown. The horizontal scale is proportional to distance between the sections, and the vertical scale is proportional to relative time; thus, a horizontal line between the three sections is interpreted to be isochronous. Missing intervals, due either to erosion or nondeposition, are indicated by the diagonally ruled areas. General environmental (lithologic) correlations are indicated by the dashed lines. Modified from McHargue (1981), with addition of formation names and ages as determined by macrofossils by Derby.

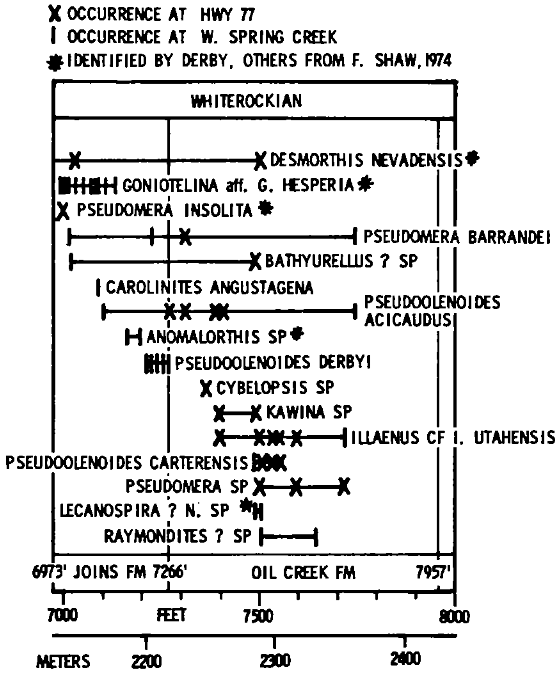


Figure 16. Early Whiterockian brachiopods and trilobites from the Joins and Oil Creek Formations; shown are composite ranges for the Highway 77 and West Spring Creek sections, western Arbuckle Mountains, Oklahoma (see Fig. 12). Brachiopods, gastropod, and the trilobite *Goniotelina* aff. *G. hesperia* identified by Derby; all others plotted from data in Shaw (1974). From Derby and others (1977). See Figure 7 for explanation of footages.

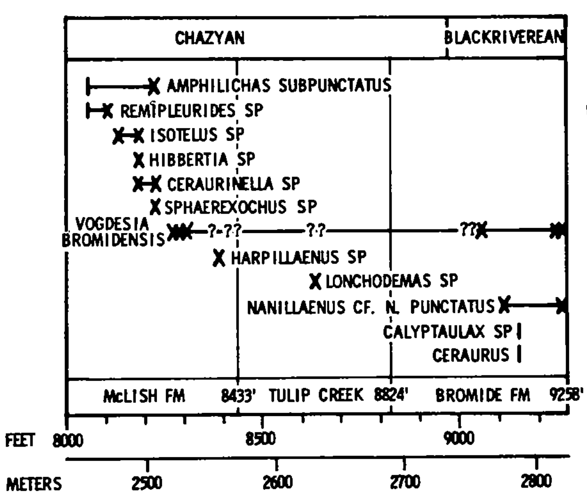


Figure 17. Late Whiterockian ("Chazyan") and Blackriveran trilobites of the McLish, Tulip Creek, and Bromide Formations; shown are composite ranges for the Highway 77 and West Spring Creek sections, western Arbuckle Mountains, Oklahoma (see Fig. 12). Plotted from data in Shaw (1974); from Derby and others (1977). See Figure 7 for explanation of footages.

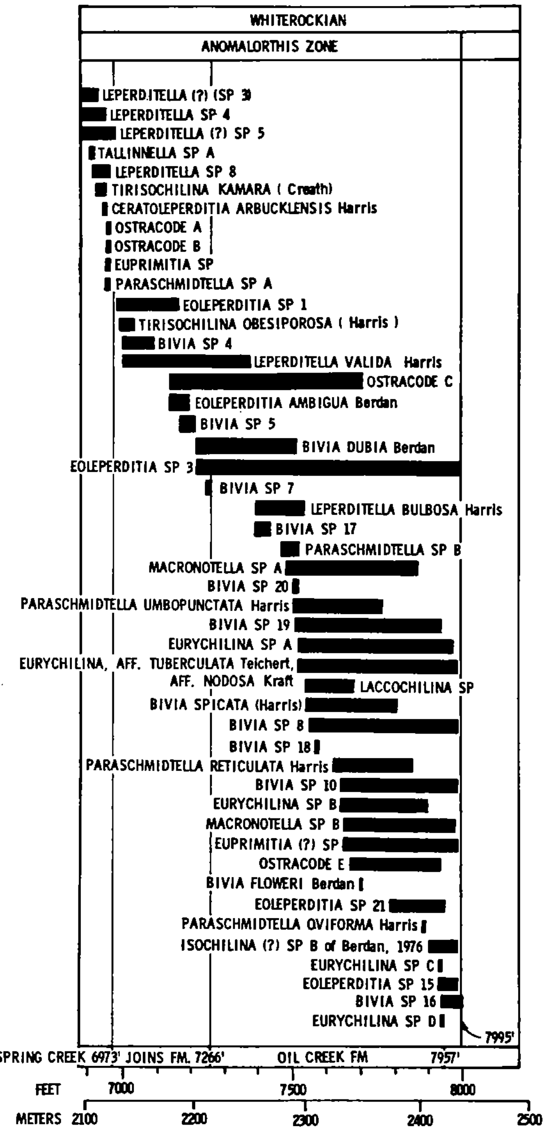


Figure 18. Early Whiterockian ostracodes from the uppermost West Spring Creek, Joins, and Oil Creek Formations; shown are composite ranges for the Highway 77 and West Spring Creek sections, western Arbuckle Mountains, Oklahoma (see Fig. 12). Note similarity in ranges to species described by Berdan (1976, fig. 2) from the Ibex area of Utah. From Derby and others (1977); all identifications by W. B. Creath. See Figure 7 for explanation of footage; *Anomalorthis* Zone is that of Cooper (1956).

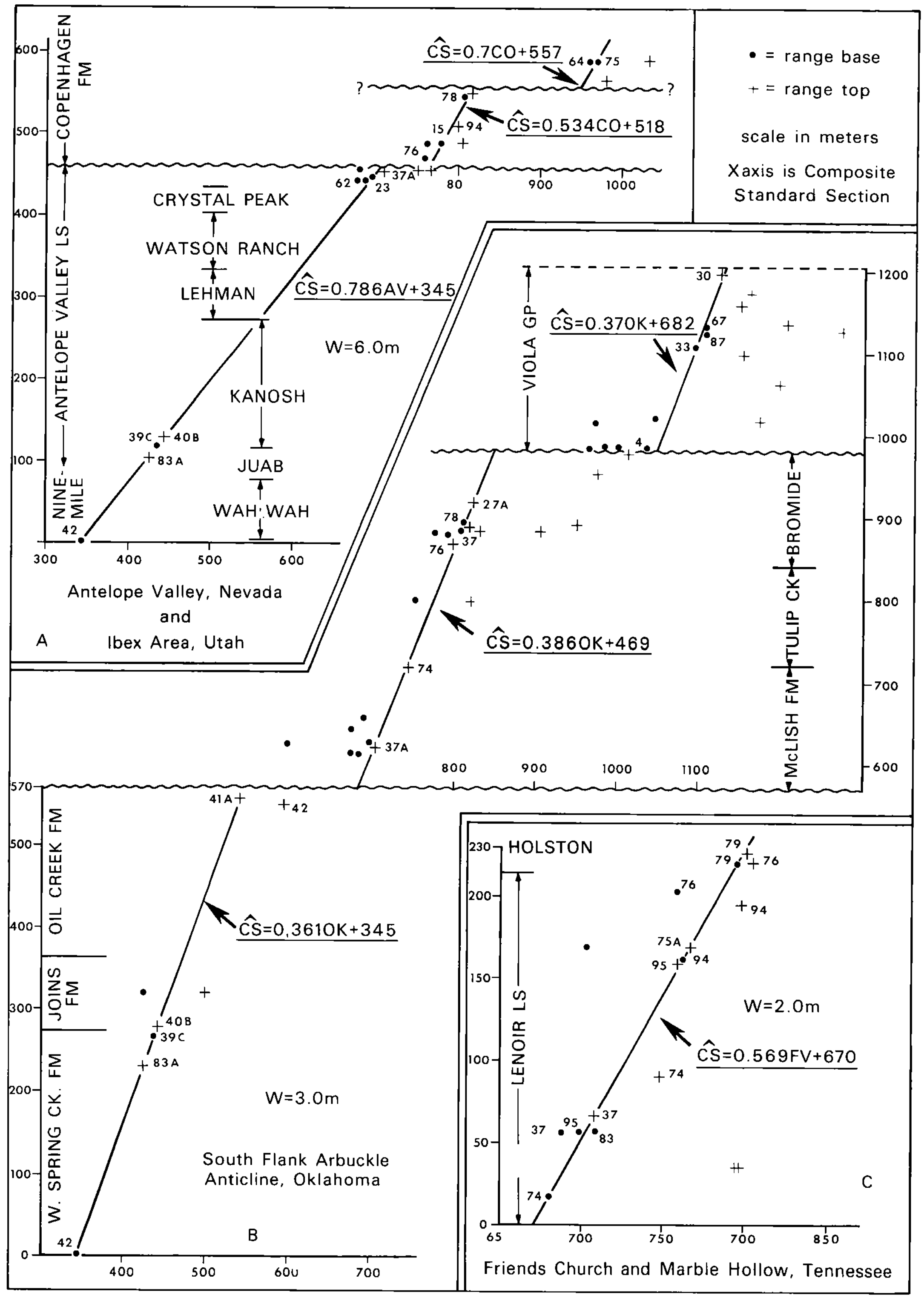


Figure 19. Graphic correlation by Sweet (1984) of three sections with his composite standard section (CSS), showing the magnitude of unconformities in the Middle and Late Ordovician of Oklahoma and Utah: *A*—Composite section of Antelope Valley, Nevada, and Ibex area, Utah. *B*—Upper West Spring Creek through Viola Springs Formations exposed along Interstate 35, Arbuckle Mountains,

of the Oil Creek Formation and the base of the McLish. In his correlation diagram (Fig. 20), Sweet suggested that this hiatus represents the time-equivalence of the Lehman, Watson Ranch, and Crystal Peak Formations of the Ibex area of Utah. The presence of this hiatus is suggested by other faunal groups as well; however, ostracode and macrofossil evidence suggests that the Lehman, at least in part, is equivalent to the Oil Creek (see Berdan, 1976; 1988, fig. 11).

Bauer (1987) documented the conodont biostratigraphy of the McLish and Tulip Creek Formations in an Oklahoma Geological Survey publication. More recently, Bauer (1989) studied the Burgen, Tyner, and Fite Formations of northeastern Oklahoma (Fig. 22). He concluded that the upper part of the Burgen through middle part of the Tyner contains conodonts that correlate with those elsewhere in lower and middle Whiterockian (pre- to lowest "Chazyan") (Figs. 2,23). The overlying upper part of the Tyner and the Fite Formations contain faunas identical to those identified by Sweet (in Amsden and Sweet, 1983) from outcrops farther south in eastern Oklahoma as being not older than Kirkfieldian (Corbin Ranch Formation or lower part of the Viola Springs Formation equivalent). Therefore, most of the upper Whiterockian ("Chazyan") and all of the Blackriveran are missing on the unconformity within the Tyner. The missing interval equates with the middle and upper parts of the McLish, the Tulip Creek, and the Bromide Formations. Derby (1973a) earlier had correlated the lower and middle Tyner with the Oil Creek; Bauer's data indicate that it is slightly younger. Bauer is continuing to study the Whiterockian strata of Oklahoma (Bauer, 1990; Ritter and Bauer, 1990) and will undoubtedly refine correlations in this interval.

Middle Ordovician (Bromide Formation) acritarchs (organic-walled microphytoplankton) were described in a series of papers by Loeblich and Tappan (see Loeblich and Tappan, 1978). Subsequent work on the microplankton from the remaining formations of the Simpson Group by M. A. Miller has resulted in recovery of distinctive acritarch assemblages from the Oil Creek, McLish, Tulip Creek, and Bromide Formations. Acritarchs are present in the Joins Formation, but have been damaged by diagenesis.

Chitinozoans were described from the uppermost Tulip Creek Formation by Wilson and Dolly (1964). Grahn and Miller (1986) described the chitinozoans of the Bromide Formation (Fig. 24). A distinctive chitinozoan, *Hercochitina? bromidensis*, may prove useful for local correlations. Chitinozoans are more common and assemblages more diverse in "deeper-water" settings than in "shallower-water" settings, which indicates that they have paleoenvironmental applications.

STRATIGRAPHIC CONCLUSIONS

The Timbered Hills, Arbuckle, and Simpson Groups as exposed in southern Oklahoma comprise one of the most complete and best-exposed sequences of Upper Cambrian through Middle Ordovician strata in North America. The proximity to economically significant hydrocarbon- and mineral-producing areas makes this sequence a vital stratigraphic reference section. Notwithstanding some gaps in knowledge and the need for additional study of some parts of the section, overall, the entire sequence is now well known in terms of biostratigraphic control by diverse fossil groups.

Our data indicate that the sequence is relatively complete, from the onset of sedimentation in Franconian time up to the middle Whiterockian hiatus between the Oil Creek and McLish Formations. Hiatuses below that level appear to be relatively minor and possibly of local extent. It seems reasonable that boundaries between major sedi-

Oklahoma (see Fig. 12). *C*—Friends Church–Marble Hollow section near Friendsville, Tennessee. Data for *A* largely from Harris and others (1979) and Ethington and Clark (1981); regional standard reference section (SRS) is Antelope Valley–Copenhagen Limestone section described by Ross (1970) and Harris and others (1979). Data for *B* (Arbuckle Mountains section) largely from undescribed Ohio State University collections, parts of which were studied by Dresbach in an unpublished Ohio State University M.S. thesis. Data for *C* (Tennessee section) from Bergström (1973) and Bergström and Carnes (1976); lower 90 m of Lenoir strata is type section of Cooper's (1956) Marmorian Stage. Individual species are indicated by two-digit numbers, some with alphabetic suffix (e.g., 37, 40B); names of species are listed in appendix of Sweet (1984). The upper part of the Arbuckle Mountains cross-plot (*B*) is well controlled in the CSS, as indicated by the pattern of tops (+) falling to the right of the line of correlation (LOC), and bases to the left of the LOC; most of Sweet's work had been in correlating and compositing the upper part of the section (Mohawkian and Cincinnatian). The lower part of the Arbuckle Mountains cross-plot (*B*) is less well controlled, and therefore subject to more potential revision as new data are obtained; however, notice the excellent LOC defined by species 42, 83A, 39C, and 40B in both the Arbuckle Mountains cross-plot (*B*) and the Nevada–Utah cross-plot (*A*). The current studies of Ethington and Dresbach on the Arbuckle, and of Bauer on the Joins–Oil Creek, will add new data to this part of the section. From Sweet (1984).

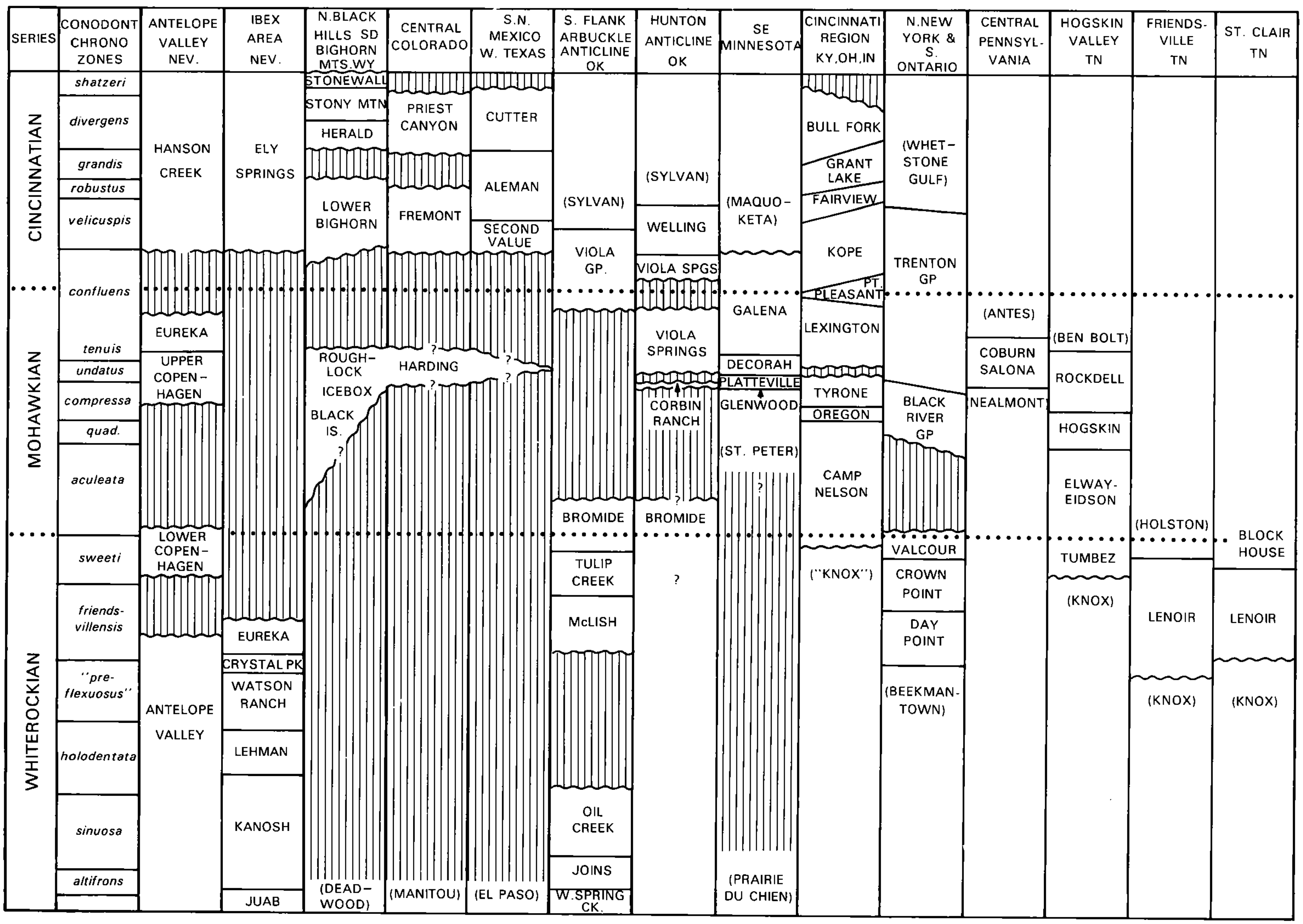


Figure 20. Correlation of Middle and Upper Ordovician strata in parts of the United States, as determined by graphic correlation based on conodonts. Names in parentheses are those of under- or overlying units for which data usable in graphic correlation are not currently available. The positions of formations with names in parentheses are not intended to indicate correlation, but simply stratigraphic position. From Sweet (1984).

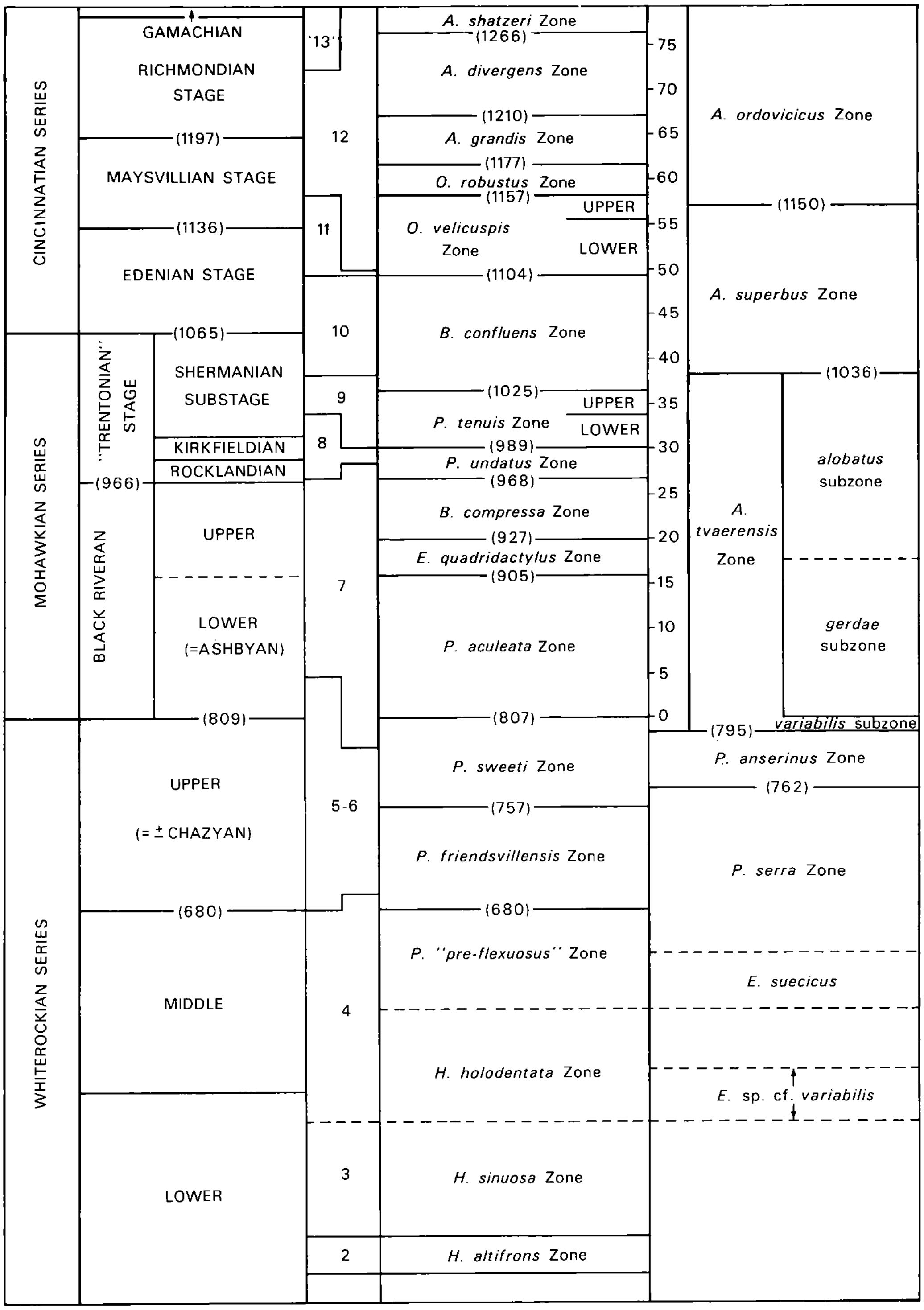


Figure 21. Middle and Upper Ordovician chronostratigraphic units (left two columns); conodont faunal un of Sweet and others (1971) (third column from left); conodont-based chronozones (fourth column); po: Whiterockian standard time units (fifth column); and North Atlantic conodont zones (right column) Bergström (1971). Vertical dimension and extent of all units determined from graphic correlation. Pr Mohawkian conodont chronozones are provisional in name and scope; Mohawkian and Cincinnatian con dont chronozones are defined in Sweet (1984). CSS and SRS values of important stratigraphic boundari are given in parentheses. From Sweet (1984).

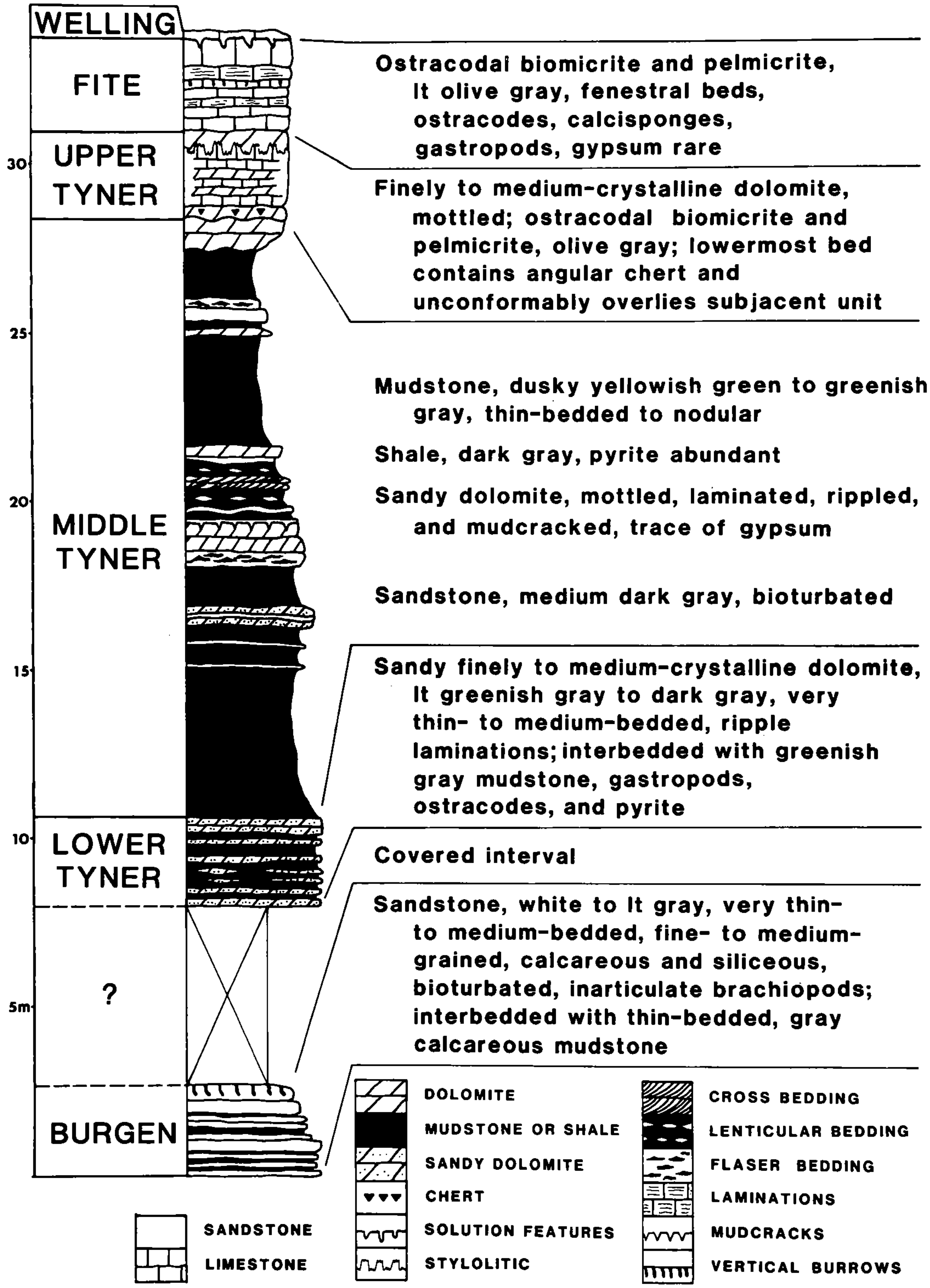


Figure 22. Composite stratigraphic section of the upper Burgen, Tyner, and Fite Formations in northeastern Oklahoma. Composited from four sections located along Oklahoma Route 10 in the Illinois River valley in T. 17–18 N., R. 22 E., Cherokee County, Oklahoma. The Tyner and Fite localities described by Amsden and Sweet are south of this area in T. 13–16 S. (see Amsden and Sweet, 1983, p. 2, for map of both areas). From Bauer (1989).

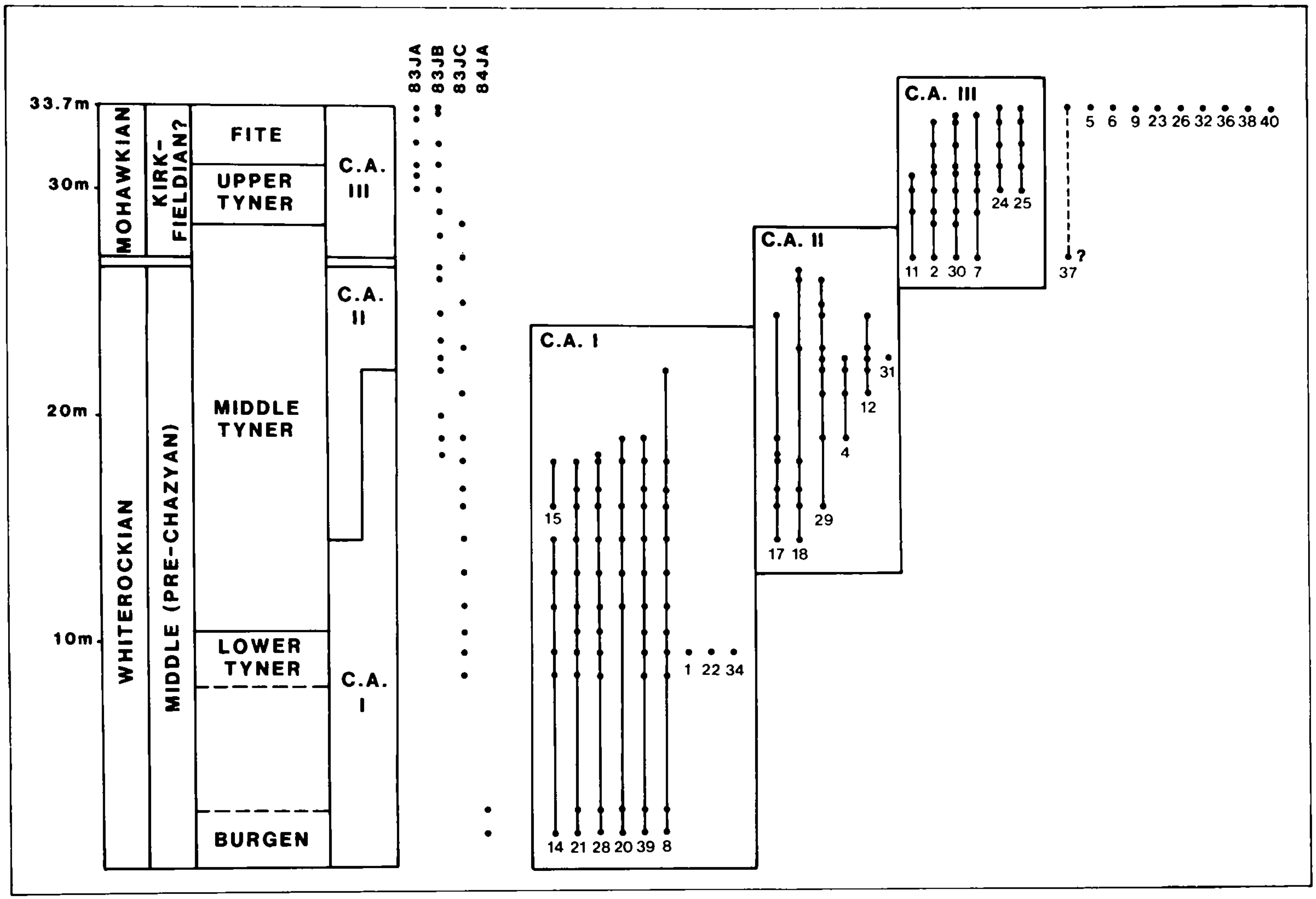


Figure 23. Conodont ranges in the composite stratigraphic section of the upper Burgen, Tyner, and Fite Formations, Illinois River valley, Cherokee County, Oklahoma, showing conodont association ranges and chronostratigraphic classification. The individual measured sections are designated 83JA, 83JB, 83JC, and 84JA; the dots under each section designation show the approximate levels of collections from that section. Conodont associations are arbitrarily designated C.A. I, C.A. II, and C.A. III (note that C.A. I has nothing to do with conodont alteration index, which is commonly abbreviated CAI). From Bauer (1989).

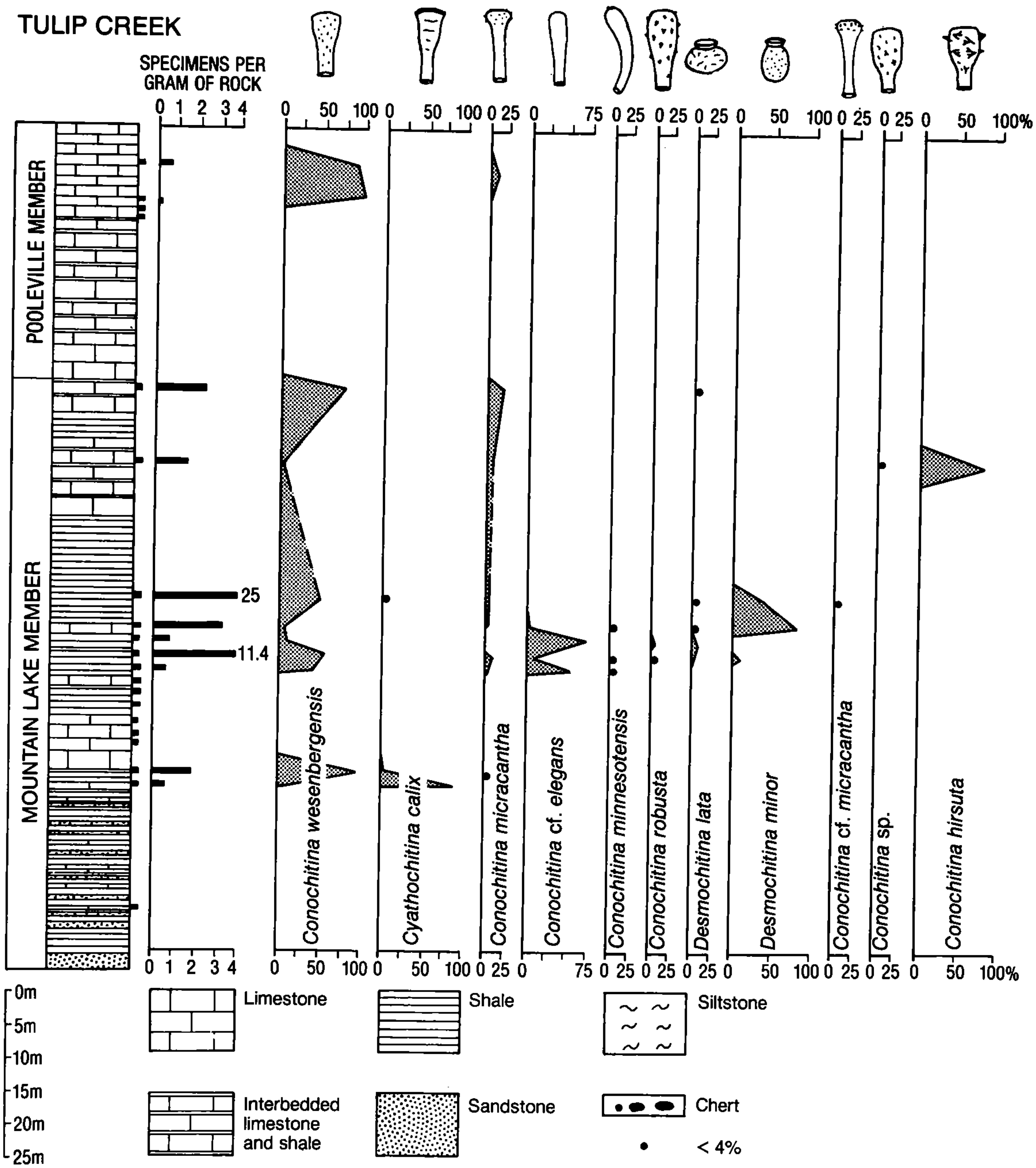


Figure 24. Lithology, sample levels, and relative frequency of chitinozoan species in the Bromide Formation, Tulip Creek section, immediately west of Interstate 35, south flank of Arbuckle anticline, Arbuckle Mountains, Oklahoma (see Fig. 12). Relative frequency is calculated on the total number of specimens recovered in each 50-g sample. Simplified lithology is after Sprinkle (1982). From Grahn and Miller (1986).

mentary sequences are more accurately located on the continental margins, where hiatuses are of lesser magnitude, than on the craton, where the same hiatuses are of considerable magnitude. For the Arbuckle anticline succession, we suggest that the upper boundary of the Sauk sequence of Sloss (1982) is better placed at the top of the Oil Creek Formation, within the mid-Whiterockian hiatus, rather than at its more traditional position at the top of the Arbuckle Group. As a generalization, the top of the Sauk sequence elsewhere is probably also better placed at the apparently widespread mid-Whiterockian hiatus, rather than the traditional approach of assigning any strata of Middle Ordovician age to the overlying Tippecanoe sequence.

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EVIDENCE OF PALEOKARSTIC PHENOMENA AND BURIAL DIAGENESIS IN THE ORDOVICIAN ARBUCKLE GROUP OF OKLAHOMA

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ABSTRACT.—Cores of Ordovician-age Arbuckle Group carbonates from Oklahoma were examined for evidence of paleokarst. The depositional and diagenetic fabric of the rock was analyzed and compared with outcrop analogs to illustrate the nature of sedimentary, karstic, and diagenetic facies. Burial diagenesis and hydrothermal alteration have in many cases obscured the original fabric of these rocks.

Arbuckle rocks in different tectonic settings and stratigraphic intervals in the subsurface of south-central and north-central Oklahoma display surprisingly similar suites of karstic and diagenetic phenomena. Dissolution cavities, solution-enlarged fractures, collapse breccias, and vugular porosity are present in many cores and attest to the predominance of fabric-destructive processes in the development of Arbuckle paleokarst. Collapse breccias and sediment-filled solution features bear striking resemblance to outcropping analogs. Primary speleothemic precipitates were not readily observed; either they were not precipitated or were obscured by later dolomitization. Phreatic cements were more commonly encountered than vadose cements.

A complex history of exposure, subsidence, and diagenesis is recorded in these rocks. Although the actual physical manifestations of paleokarst are not difficult to identify, interpretation of the genesis and age of these features is decidedly problematic. Arbuckle carbonates have been exposed to surficial weathering for periods of variable intensity and duration numerous times in geologic history. Paleokarst horizons may have developed subjacent to disconformities within and between formations of the Arbuckle Group and where these rocks subcrop beneath regional unconformities. This complex hierarchy of unconformities can produce numerous porous horizons whose preservation potential may depend on subsidence rates rapid enough to prevent extensive low-temperature phreatic cementation, thereby preserving the open pore network of the karst profile.

Burial diagenesis is evidenced by the multi-event dolomitization of these rocks. Ferroan and nonferroan "growth-zoned" baroque and limpid phreatic dolomite cements commonly occlude vugular and fracture porosity. Host-rock carbonates have been extensively replaced or neomorphosed. Cathodoluminescent microscopy and chemical staining indicate that "growth-zoned" baroque dolomite is commonly uniform in composition and was precipitated under mildly reducing conditions. Dolomite cementation was arrested by the migration of oil into the remaining pore space.

INTRODUCTION

The limestones and dolomites of the Arbuckle Group have produced oil and gas from a wide variety of settings throughout the state of Oklahoma. While production is commonly structurally controlled, it can be demonstrated that porosity development is often due to modification of Arbuckle carbonates by (1) erosion at an unconformity surface, (2) fracturing and brecciation related to cavern collapse and/or tectonism, and (3) hydrothermal alteration of host rock in the subsurface. Recent unexpected and significant oil and gas production from Arbuckle paleokarst reservoirs (Shirley, 1988) in the Anadarko and Arkoma basins emphasizes the need for immediate, detailed inquiries into the nature of Arbuckle paleokarst.

Objectives

The primary objective of this study is to provide an inventory of paleokarstic features present in Arbuckle rocks from the subsurface in Oklahoma. The petrology, possible genesis, and subsequent diagenesis of these features are described; a complex set of dolomite types is presented and explained in relation to the burial history of these rocks.

This study is part of an on-going program at Oklahoma State University to investigate the geology of Arbuckle paleokarst in Oklahoma, both on the surface and in the subsurface.

Previous Investigations

It has long been recognized that fracturing and the establishment of vugular porosity play an important role in the productivity of Arbuckle reservoirs (Bartram and others, 1950; Walters, 1958; Gatewood, 1979). However, the study of these features in the context of karst facies is still in its infancy.

Paleokarstic sink holes and sand-filled caves have been identified in Cotter Dolomite outcrops in northeast Oklahoma by Gore (1952). Collapse breccias have been described in the Kindblade Formation in the Arbuckle Mountains by Tapp (1978) and in the Cool Creek Formation by Ragland and Donovan (1985). Post-Permian to recent karstification of Arbuckle rocks in southern Oklahoma has been documented by Decker and Merritt (1928) and Curtis (1959). Donovan (1987) described speleothems coeval with oil migration in a small Permian paleokarstic reservoir on Bally Mountain, in southwest Oklahoma.

Disconformities within and between the various formations of the Arbuckle Group have been recognized by Ireland (1955), Chenoweth (1968), Derby (1969), and Reeder (1974), although discussion of the role these events might play in karst formation has been lacking to date.

Arbuckle paleokarst in the subsurface of Kansas has been documented by Merriam and Atkinson (1956) and Walters (1958). Paleokarst features in the Arbuckle-equivalent Ellenburger Group carbonates in Texas have been described by Ijirigho and Schreiber (1986) and Kerans (1988).

Methods and Procedures

Forty-two cores of Arbuckle Group rock housed at the Oklahoma Geological Survey Core and Sample Library in Norman, Oklahoma, were examined in a cursory manner to determine if the rock had in any way been affected by the physical or chemical processes that evidence karstification. Cores that exhibited unusual and possibly karstic features were procured for more detailed analysis. Those that did not exhibit these features were briefly described and cataloged for future reference. Seven cores from a variety of locations in Oklahoma were, on the basis of the initial macroscopic examination, selected for intensive study. Additionally, one other core was studied; it is the property of Oklahoma State University. Figure 1

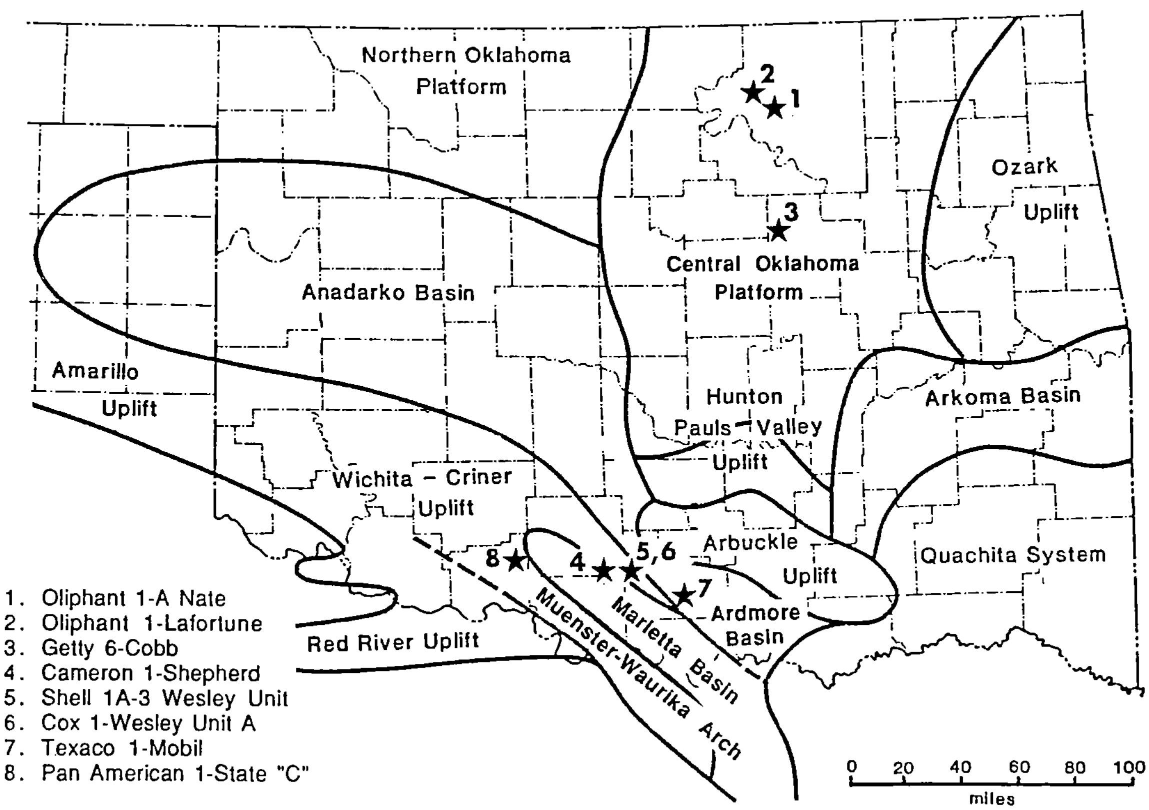

Figure 1. Locations of the cores (★) used in this study relative to the tectonic provinces of Oklahoma (after Al-Shaieb and Shelton, 1977; Arbenz, 1956).

TABLE 1. — CORE LOCATION AND INTERVAL INFORMATION

Core	Location	County/ field	Cored interval (depth, ft)	Cored formation	Depth below T/Arb. (ft)	Strata overlying Arbuckle
Oliphant 1-A Nate	15–T24N–R7E SW¼SW¼SE¼	Osage Naval Res.	–2,860 to –2,871	Cotter	0—at top	Burgen (Simpson)
Oliphant 1-Lafortune	8–T25N–R6E SE¼NW¼NE¼	Osage Burbank	–3,334 to –3,361	Cotter	0—at top	Burgen, (Simpson)
Getty 6-Cobb	3–T17–R7E SE¼SE¼	Creek Cushing	–2,466 to –2,517	?	0—at top	Bartlesville
Cameron 1-Shepherd	20–T3S–R4W SE¼NW¼NE¼	Jefferson Dixie Area	–6,278 to –6,296 (poor recovery)	?	3,640	Deese
Shell 1A-3 Wesley Unit	3–T4S–R3W NW¼SW¼SE¼	Carter Healdton	–3,584 to –3,633	Kindblade[a]	1,050	Joins, (Simpson)
E. L. Cox 1-Wesley Unit A	3–T4S–R3W NE¼SW¼SW¼	Carter Healdton	–3,806 to –3,983	Kindblade[ab]	870	Joins (Simpson)
Texaco 1-Mobil	1–T5S–R1W NE¼SE¼	Carter Wildcat	–6,300 to –6,310 –6,519 to –6,534	Kindblade[a]	420 640	Atoka
Pan American 1-State "C"	36–T2S–R10W N½NW¼NE¼	Cotton Wildcat	–7,475 to –7,500	Kindblade[b]	1,775	Joins (Simpson)

[a]Correlation based on well log signature.
[b]Correlation based on conodont biostratigraphy.

shows the locations of the cores in relation to the tectonic provinces of Oklahoma. Pertinent data for each core is presented in Table 1. Each core was slabbed and described and data were recorded on a petrolog form that was specifically designed for description of both depositional and diagenetic (karstic) features. Thin sections were prepared for petrofabric analysis utilizing both polarizing and cathodoluminescence microscopes. Many thin sections were stained with alizarine red-S and potassium ferricyanide to distinguish ferroan and nonferroan varieties of dolomite and calcite (see Adams and others [1984] for procedure). A scanning electron microscope coupled with energy dispersive X-ray analyzer (SEM/EDXA) was used to investigate the crystallinity and chemical composition of individual crystals of each cement species. Preliminary analysis of stable carbon and oxygen isotopes was performed on 35 samples of host rock and dolomite and calcite cements by Paul Wagner, at Amoco Production Co. Research Center in Tulsa, Oklahoma. Conodont elements garnered from insoluble residues of selected intervals were identified by Scott Ritter of the Oklahoma State University School of Geology.

GENERAL GEOLOGY

Arbuckle Group Stratigraphy

Figure 2 depicts the stratigraphy of the Arbuckle Group in southern and northeastern Oklahoma. Note especially the locations and relative magnitudes of unconformities within the Arbuckle Group and between the Arbuckle Group and the overlying Simpson Group.

Arbuckle rocks have suffered exposure and erosion numerous times since their deposition. The possibility exists (and must be explored) that porosity and karst may have developed at any or all of these unconformities, given sufficient time of exposure and environmental conditions. The effects of pre-Simpson erosion on Arbuckle strata in

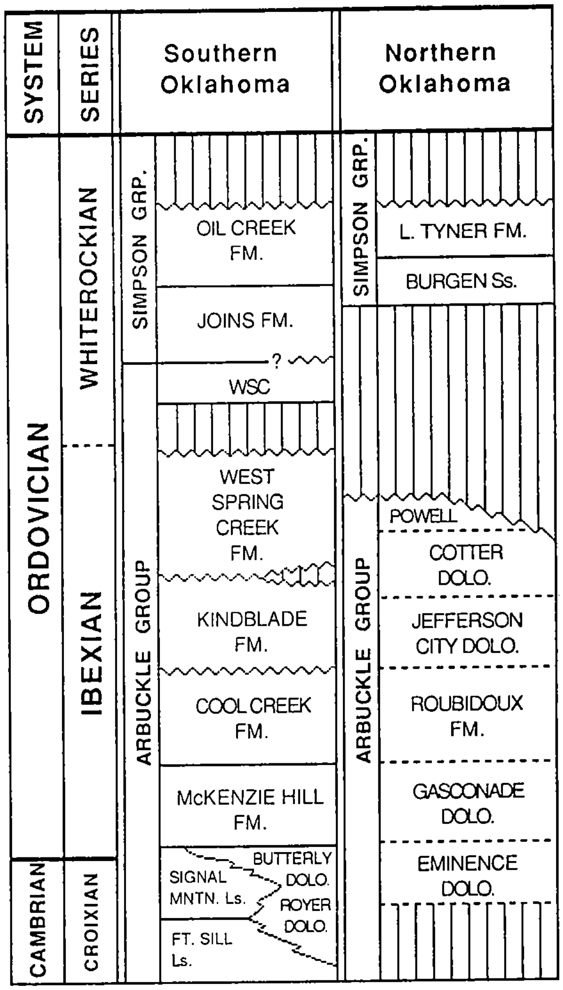

Figure 2. Stratigraphic column showing correlative formations of the Arbuckle Group in southern and northeastern Oklahoma (modified after Chenoweth, 1968; Ross and others, 1982).

northeastern Oklahoma are well documented (Walters, 1958; Reeder, 1974), but may be in question in southern Oklahoma where the Joins Formation of the Simpson Group overlies the West Spring Creek Formation with apparent conformity (Reed, 1957; Derby, 1969; Latham, 1970). Due to the rapidly subsiding, more mobile nature of the southern Oklahoma paleodeep, the "pre-Simpson unconformity" may not be represented by a single discrete surface at the top of the Arbuckle, but may instead be represented by subtle disconformities within the West Spring Creek Formation (see Derby, 1969).

Although they are lithologically similar, carbonates of the Arbuckle Group in southern Oklahoma are predominantly limestone, while those on the northern Oklahoma platform are dolomite (Ham, 1969). It is worth noting that all of the cores examined in this study were, regardless of location, extensively dolomitized.

Arbuckle Depositional Facies

Arbuckle rocks are characterized by numerous and recurring shallowing-upward, peritidal carbonate cycles. Subtidal mudstones and wackestones, intertidal bioclastic packstones and grainstones, and restricted, upper intertidal algal boundstones are commonly observed facies. Exquisite algal stromatolites and thrombolites occur in the Cool Creek Formation, and also in the Kindblade and West Spring Creek Formations more infrequently. Supratidal, evaporite-dominated sabkha facies are not well developed; either they were never deposited or were removed by erosion subsequent to deposition. In many instances a "typical" shallowing-upward cycle culminates with an intraformational conglomerate which probably represents a period of subaerial exposure, desiccation, erosion, and redeposition. Such erosive events may be correlative to hardgrounds and karst surfaces elsewhere in the Arbuckle environment (Donovan and others, 1983). Shallowing-upward sequences are seldom delimited above by supratidal facies such as interlaminated evaporites and dol-algal mats. Evaporites are encountered far more frequently in the literature (Reed, 1957; Latham, 1970; Gatewood, 1978; St. John and Eby, 1978; Beales and Hardy, 1980; Ragland and Donovan, 1985) than in either the outcrop or subsurface. Nodular and remobilized fracture-filling evaporites were noted in very few cores from southern Oklahoma; interestingly, these cores do not show signs of paleokarstification, nor do they exhibit the coarse dolomite textures found in the principle cores of this study. Evidence for evaporite presence in these cores is scant and equivocal, such as submillimeter-size anhydrite inclusions in chert nodules, and a few silicified salt hopper crystals recovered from insoluble residues from the Cox-Wesley core.

PALEOKARST FEATURES OBSERVED IN CORES OF ARBUCKLE ROCK

Breccias and Conglomerates

Breccias and conglomerates are lithotypes common to each of the Arbuckle Group cores studied. Where possible, the genesis of each conglomerate or breccia was inferred from the fabric of the rock and its relation to surrounding structures. Breccias were described with a combination of genetic terminology and textural modifiers, as proposed by Norton (1917) and Ijirigho and Schreiber (1986). Blount and Moore (1969) outlined certain criteria for differentiating between genetic breccia types.

Crackle Breccia

The term "crackle breccia" was introduced by Norton (1917) to describe incipient brecciation in extensively fractured rocks where the fragments have not been dislodged, rotated, or otherwise moved to any appreciable degree. This is a structural term that bears no genetic connotations. Ijirigho and Schreiber (1986) modified this term to describe crackle breccias with open or slightly dilated fractures, which they termed "microdil breccias." Kerans (1988) noted that these breccia-types, which he termed "fracture breccia" commonly occur in the "cave roof facies" of karstified Ellenburger Group carbonates in West Texas. Usually, this type of breccia is indistinguishable from—and indeed, may be nothing more than—intensely, and perhaps tectonically, fractured rock. It is only through its spatial relations to other karst features that this type of breccia has any use in karst-facies diagnosis.

Figure 3. Crackle-brecciated rock subjacent to an intra-Arbuckle disconformity. Many fractures are enlarged and filled with micritic sediment. (Oliphant-Lafortune core, –3,352 ft.)

Definitive crackle breccias were observed locally grading into mosaic or rubble collapse breccias in many of the study cores. Figure 3 shows crackle-brecciated rock subjacent to an intra-Arbuckle disconformity, in the Oliphant-Nate core. Early, crackle-breccia porosity is commonly occluded by infill sediment or cement.

Collapse Breccia

Collapse breccia is a term used to imply a genetic origin for breccia texture(s) that result from structural collapse of a previously open megapore or cavern network. The structural integrity of the host rock may be compromised by the dissolution of matrix carbonates and/or interbedded soluble evaporites. Foundering may be induced by gravity at the surface if the host rock is sufficiently eroded internally, or by the application of overburden subsequent to burial.

Collapse breccias are characterized by marked heterogeneity and angularity of clasts (which may be derived from numerous and lithologically dissimilar overlying stratigraphic units), poor sorting, with interstitial cement or matrix in a dominantly clast-supported fabric (Blount and Moore, 1969). Collapse breccias may exhibit the "mosaic," "random," and "rubble" textures of Norton (1917) and Ijirigho and Schreiber (1986).

Many of the breccias noted in this study exhibited fabrics suggestive of formation by solution and collapse. A few of the characteristic breccia fabrics observed in cores are described below.

The Pan American-State core from Cotton County (Fig. 4) is comprised nearly entirely of collapse breccia. The clast population—which ranges from granule to boulder size and is very poorly sorted—is comprised of at least six different lithologies, the most exotic of which is a silicified ooid grainstone that is not present as in-place host rock anywhere else in the core. The breccia exhibits a predominantly random fabric. Large, fractured clasts exhibit a mosaic fabric internally. The breccia matrix is a dark-brown, clayey, dolomitic mudstone.

Heterolithic, rubble collapse breccias as much as 4 ft thick were encountered in the Texaco-Mobil core (Fig. 5). These breccia immediately overlie crackle-brecciated host rock.

The Cox-Wesley core from Healdton field exhibits unique and complex "zebroid" brecciation (Beales and Hardy, 1980) and coarse dolomite cementation (Fig. 6A). This type of brecciation has been attributed to host-rock expansion due to displacive cementation subsequent to evaporite dissolution (Beales and Hardy, 1980). However, neither relict nor extant evaporites were noted in the brecciated interval. In places, the brecciated zone exhibits a random-fragment packing suggestive of collapse (Fig. 6B). The very angular breccia fragments are cemented with a 1-cm-thick rind of isopachous saddle dolomite.

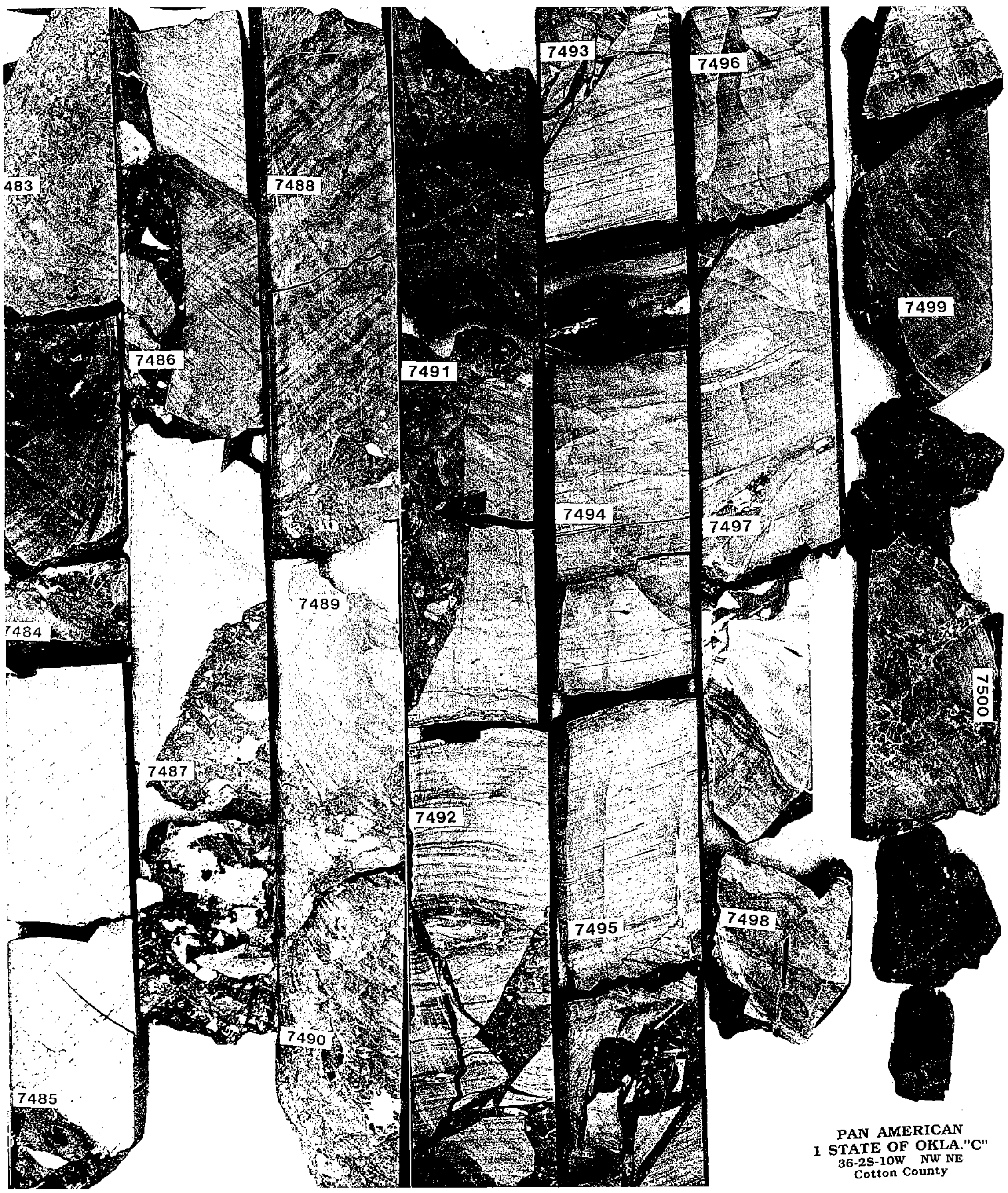

Figure 4. Heterolithic collapse breccia exhibiting random texture (RB) and mosaic texture (MB). (Pan American-State core.)

Figure 5. Rubble collapse breccia. (Texaco-Mobil core, –6,521 ft.)

Cavern-Fill Parabreccia

The term "cavern-fill parabreccia" was proposed by Lynch (1990) for poorly sorted, matrix-supported breccias that exhibit structures suggestive of subterranean deposition. (The precedent for this terminology is the paraconglomerate of Pettijohn [1975], which, with the exception of clast shape, this lithotype texturally resembles). The term is genetic, although structurally it is analogous to Norton's (1917) "pudding breccia," and also to the "pheno breccia" of Ijirigho and Schreiber (1986). Cavern-fill parabreccias are deposited by aqueous media flowing through an open mega-pore or cavern network within a karstified terrain. Phenoclasts may calve off the roof of the cavern in which they are deposited, or be transported from locations of collapse elsewhere in the cavern network (probably during periods of turbulent flow and rapid discharge after storm events). The fine matrix is composed of cave mud or infiltrated sediment. These breccias differ from true solution-collapse breccias in that they are comprised mostly of matrix that has filled an uncollapsed cavern. Analogous deposits have been observed by the authors in recent karst in the Lake of the Ozarks region of Missouri.

Figure 7A,B shows two interesting cavern-fill parabreccias in the Cox-Wesley core at depths of –3,939 and –3,972 ft. The parabreccias are lithologically similar; each is comprised of very angular phenoclasts of chert and light blue-gray limestone supported in a matrix of dark-brown, clayey, dolomitic mudstone. The parabreccia at –3,939 ft contains conspicuous limestone phenoclasts that are unlike any other lithology encountered in the cored interval. This breccia was deposited within an eviscerated chert bed, from whence the chert phenoclasts in both this deposit and the one at –3,972 appear to have been derived. Thus it would appear that an open solution-tunnel network of at least 33 vertical feet (and of indeterminable lateral extent) was present within this sequence of rocks. There are no remnants of this pore network that are open currently as effective, channel porosity.

Sedimentary Breccias and Conglomerates

The terms "sedimentary conglomerate" and "sedimentary breccia" are used to describe carbonate rudstones having rounded or angular clasts, respectively, that are formed by erosion of an exposed carbonate terrain adjacent to, or possibly within, the environment of deposition. While perhaps not specifically related to karstification, these deposits do evince surficial erosion during the time of Arbuckle sedimentation. This type of breccia is analogous to the "depositional breccia" of Blount and Moore (1969).

A 1-ft-thick sedimentary conglomerate comprised of well-rounded pebbles of dark-gray mudstone within a light-gray, detrital quartz-sand-rich carbonate mudstone matrix was observed in the Oliphant-Lafortune core (Fig. 8A). The noncontemporaneity of clasts and matrix is indicated by the fact that some of the clasts are internally fractured, and none of them contain the quartz sand that is prevalent in the matrix. The clasts exhibit some rough imbrication which, together with the detrital quartz, suggests the possibility that this is a subaqueous channel-like deposit. The pre-burial formation of this conglomerate is indicated by the transection of both the clasts and the matrix by rather conspicuous stylolites.

The Getty-Cobb core from Cushing field exhibits a sedimentary breccia that contains phenoclasts of dolomudstone and silicified ooid grainstone(?) in a finely laminated, green shale matrix

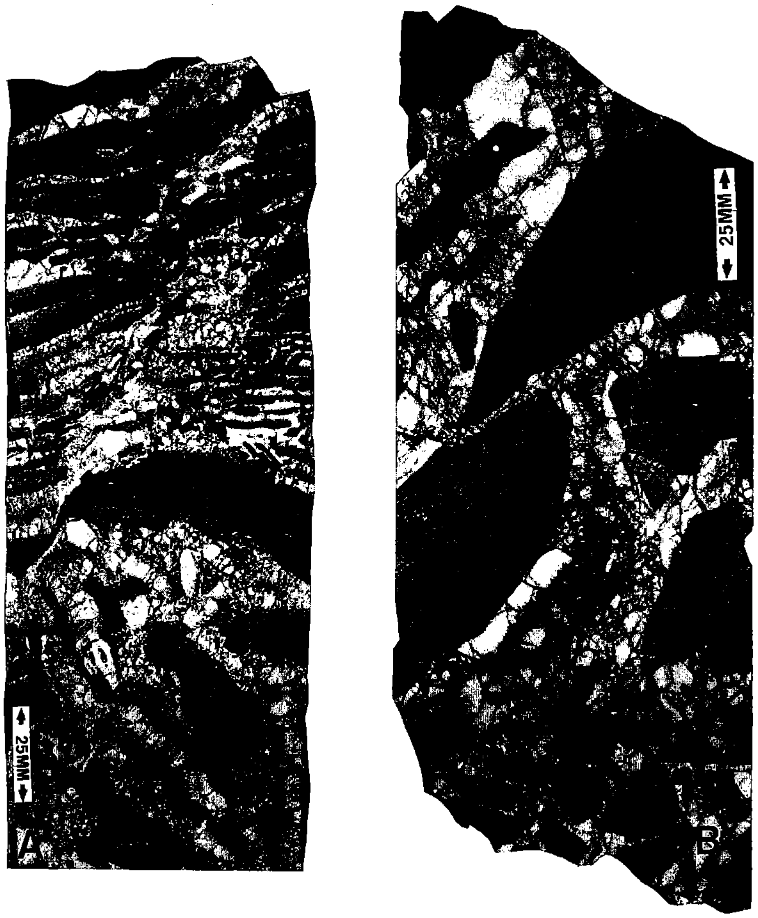

Figure 6. *A*—Zebroid breccia. Dark laminae are laths of host rock, white laminae are subisopachous crusts of columnar and saddle dolomite. (Cox-Wesley core, –3,954 ft.) *B*—Random collapse breccia cemented with isopachous saddle dolomite. (Cox-Wesley core, –3,949 ft.)

(Fig. 8B). The brecciated interval is delimited above and below by transitional contacts with the host rock.

Dissolution Features

The most commonly distinguishable type of dissolution observed in this study was that which was limited to the millimeter or centimeter scale, or those features that are small enough to be contained within the diameter of the core sample. Dissolution vugs, solution-enlarged fractures, and solution channels were observed in the study cores; more often than not the original pore space was occluded by infill sediment or cement.

The type of porosity created in karst terrains depends strongly upon the fabric and lithology of the host carbonate. In low-permeability carbonates such as Arbuckle rocks, vugular and solution-channel porosity (Choquette and Pray, 1970) may be created by the restriction of ground-water flow through high-permeability conduits such as fractures or bedding planes. Extensively karstified terrains may exhibit cavern porosity, which Choquette and Pray (1970) define as the smallest opening that an adult human can enter, or, if the rock is encountered in subsurface drilling, as any opening large enough to cause a noticeable drop in the drill bit.

The vugular porosity encountered in the rocks of this study typically range from 0.5 to 2 mm in diameter, and occasionally as large as 2 cm. The distinction between very large vugular porosity and channel porosity is often arbitrary, especially when determined from small-diameter core samples.

Striking examples of solution-induced channel porosity and matrix-vugular porosity were observed in the Oliphant-Nate core from Osage County. The fine (<1 mm in diameter) vugular

Figure 7. Cavern-fill parabreccias from the Cox-Wesley core. *A*—Parabreccia within an eviscerated chert (CH) bed at –3,939 ft. Angular phenoclasts are blue-gray dolomicrite. *B*—Parabreccia within a tan dolomicrite host (HR) at –3,972 ft. Chert phenoclasts (CH) derived from chert bed at –3,939 ft exhibit early, internal fractures that are healed with dolomite cement. Late, open tectonic fractures transect host rock, parabreccia matrix, and chert clasts.

porosity is notably concentrated in halos that extend a few centimeters outward from the eroded walls of the solution channels (Fig. 9A). Geopetal sediment, rhombic dolomite cement, and pyrobitumin (in order of emplacement) commonly fill the vugular and channel pore space.

A detailed view of a large vug in the Cox-Wesley core (Fig. 9B) shows clayey sediment geopetally filling the open pore. The orientation of the infill sediment relative to the dipping beds of the host rock suggests that the pore formed—or, was at least filled—after the Pennsylvanian tectonism that induced the folding.

Cements

The morphology of cementation in the karst environment is controlled largely by the amount of water present. In the unsaturated (vadose) zone, cements precipitate from water that gravitates toward the bottoms of grains or is bound by capillary pressure at grain contacts—hence, these cements often display characteristic pendant (also termed microstalactitic, or dripstone) and meniscoid fabrics. "Phreatic" cements precipitate in solution-filled cavities and tend to form isopachous coatings on pore walls and grains. Speleothemic precipitates (stalactites, stalagmites, dripstone curtains, etc.) composed of laminated calcium carbonate are typically formed in mature-stage karst close to the water table in the lower vadose zone (Esteban and Klappa, 1983).

Karst speleothems such as stalactites and stalagmites were not observed in any of the cores studied; either they were not precipitated or their presence was masked by later diagenetic events. Additionally, the tremendous improbability of actually coring one of these features cannot be discounted.

Cathodoluminescent microscopy revealed the presence of vadose-type microstalactitic calcite spar cementing breccia clasts in the Texaco-Mobil core (Fig. 10A,B).

The macro- and microscopically visible cements in all of the cores studied are principally subisopachous, pore-lining saddle and rhombic dolomite. Although these cements are not a product of low-temperature karst cementation (see discussion below), certainly their isopachous morphology suggests precipitation under predominantly phreatic conditions.

Infill Sediment

The phrase "infill sediment" refers to the sediments that are deposited anywhere within the karst profile, including breccia matrix and pore-filling sediments.

The lithology of the infilling sediment may reflect the origin of the sediment and, roughly, the timing of its deposition within the karst system. Arbuckle-age micrite was commonly observed filling solution-enlarged and crackle-breccia fractures (see Fig. 3). Additionally, some solution tubes were filled with what appears to be Simpson Group sand, below the eroded top of the Arbuckle (Fig. 11). It is perhaps significant that the detrital feldspars, polycrystalline quartz, and metamorphic rock fragments that are nearly ubiquitous in Pennsylvanian sandstones were not encountered in any infill sediment in the study cores. The parabreccia matrix described earlier is typically comprised of an admixture of terrigenous clastic clays, carbonate mud, and silty (possibly detrital?) dolomite.

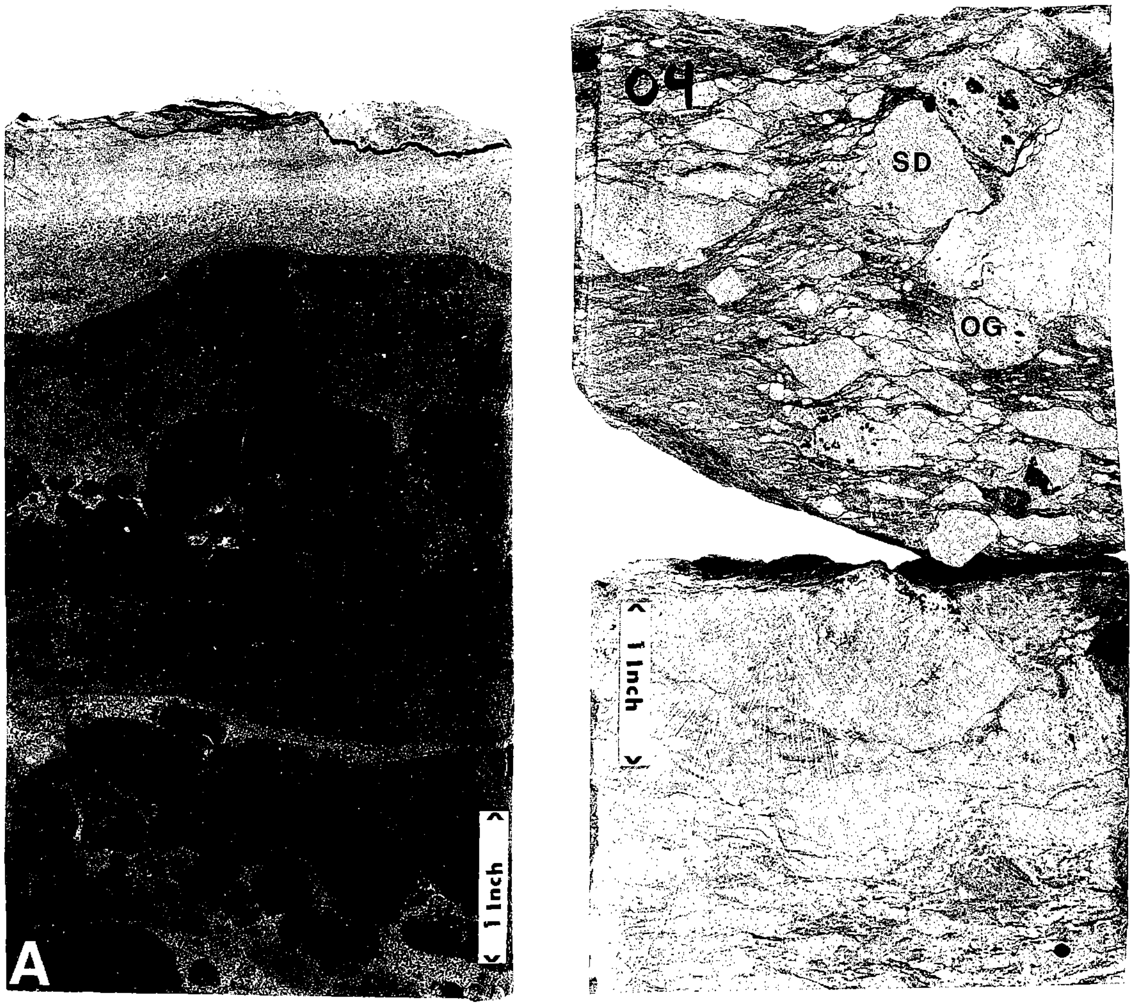

Figure 8. *A*—Sedimentary conglomerate comprised of dark-gray dolomicrite pebbles within a sandy, dolomicrite matrix. (Oliphant-Lafortune core, –3,349 ft.) *B*—Sedimentary breccia comprised of clasts of tan, sucrosic dolomite (SD) and silicified ooid grainstone (OG), within a green shale matrix. (Getty-Cobb core, –2,504 ft.)

DOLOMITE TYPES AND SEQUENCING

The dolomite in the study cores was differentiated into eight types on the basis of crystal habit and fabric, relation to host rock, and mineralogy as determined from energy dispersive X-ray analysis, cathodoluminescence, and staining. Five types of dolomite are considered matrix replacive, or recrystallized host rock, and include: (1) host matrix dolomicrite/dolomicrospar, (2) hypidiotopic-rhombic (sucrosic) dolomite, (3) clay-associated hypidiotopic-rhombic dolomite, (4) saddle-rhombic dolomite, and (5) xenotopic dolomite. Cements that were precipitated in open pore space include (1) saddle dolomite, (2) columnar dolomite, and (3) idiotopic-rhombic dolomite. The characteristics of each type of dolomite are discussed below. Many of the dolomite types described in this study are similar to those considered by Lee and Friedman (1987) as being indicative of deep burial-diagenesis in Ellenburger Group carbonates in Texas.

Matrix-Replacive Dolomites

The host rock or matrix carbonate in each of the cores studied is predominantly, if not entirely, composed of dolomite. The episodic and pervasive nature of the dolomitization has made it difficult in some instances to determine whether dolomite is syndepositional, neomorphic after a dolomitic precursor, or replaced calcitic matrix.

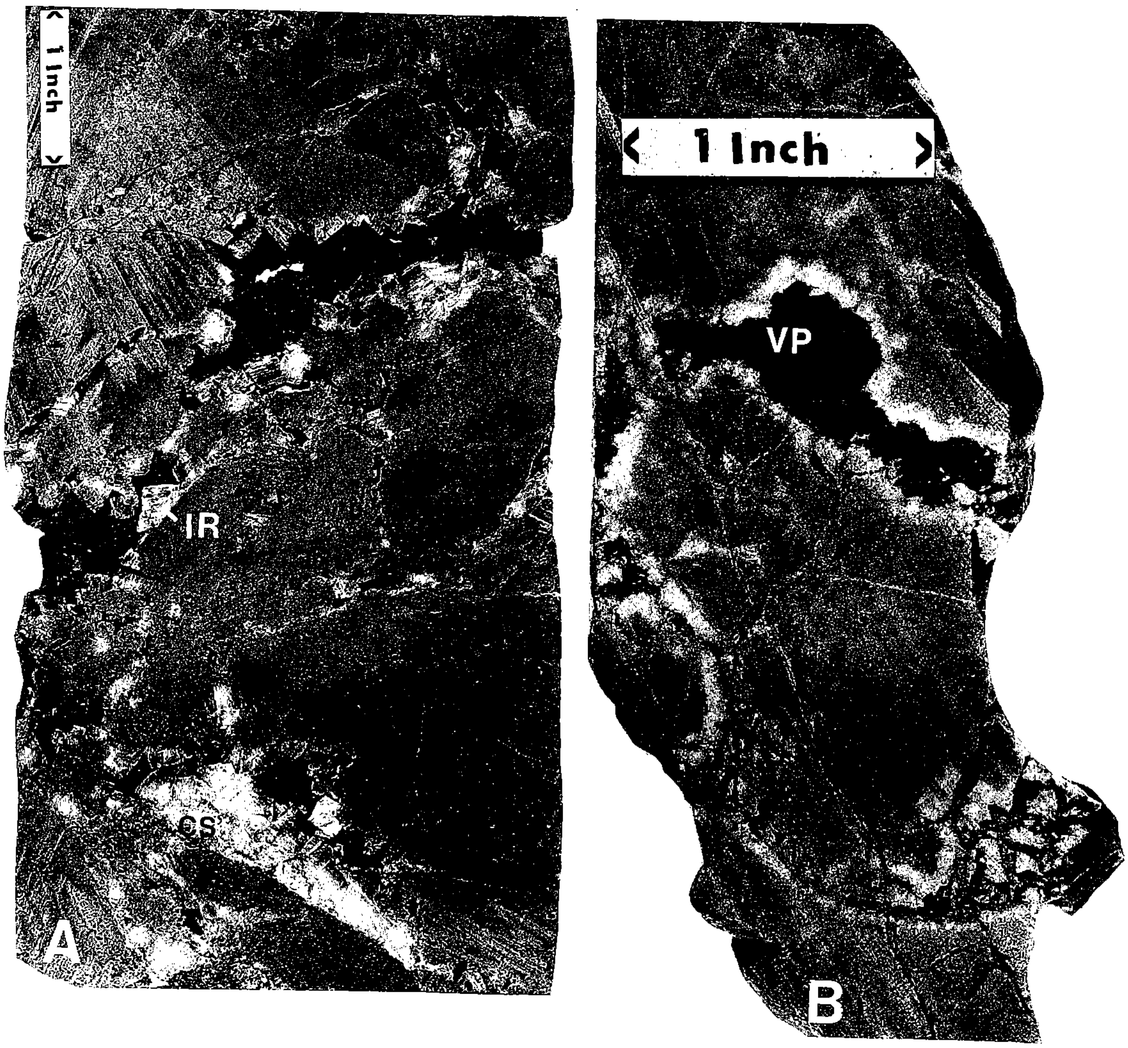

Figure 9. Examples of karstic porosity: *A*—Solution channels filled with clayey infill sediment (CS), idiotopic-rhombic dolomite cement (IR), and pyrobitumin (black, in pore space). Note vugular porosity halo (highlighted by black pyrobitumin mottling) surrounding the solution channels. (Oliphant-Nate core, –2,863 ft.) *B*—Enlarged vugular porosity (VP) with infilling sediment geopetally oriented with respect to present structural dip (indicated by black line). (Cox-Wesley core, –3,902 ft.)

Host Rock Dolomicrite/Dolomicrospar

The peritidal carbonate mudstones in these Arbuckle rocks are composed principally of very finely crystalline dolomicrite and/or dolomicrospar. Individual crystals range from 0.01 to 0.05 mm in diameter, and are usually anhedral in shape. These matrix dolomites exhibit dull red luminescence and a dirty appearance under plane polarized light (see Fig. 12A). The fine crystal size and pervasive nature of this dolomite suggest a primary, syndepositional or perhaps eogenetic origin. Petrographic evidence indicates that this type of dolomite formed prior to fracturing, brecciation, and stylolitization. This type of dolomite was observed in all of the study cores.

Hypidiotopic-Rhombic (Sucrosic) Dolomite

This type of dolomite is characterized by a porous fabric of loosely interlocking hypidiotopic to idiotopic rhomb-shaped crystals (Fig. 12B). The highest intercrystalline porosity values (7–13%) of any of the dolomite textures described in this study were encountered in this dolomite. This friable dolomite is what is commonly referred to as "sucrosic dolomite" in the subsurface vernacular. Individual euhedra range in size from 0.1 to 0.3

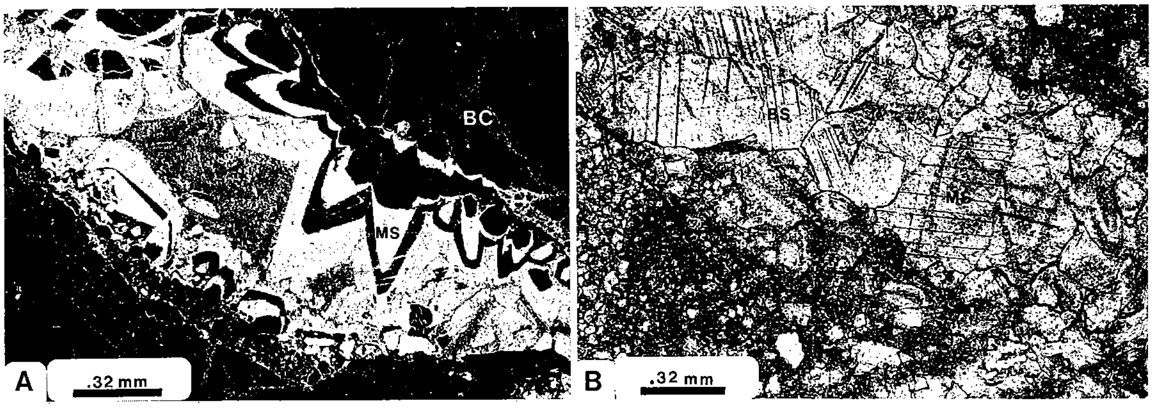

Figure 10. Thin-section photomicrographs of microstalactitic calcite spar. *A*—Cathodoluminescence reveals the delicate spar crystals (MS) that grew downward from the bottom of a breccia clast (BC). *B*—The presence of the microstalactitic spar (MS) is masked in this plain polarized light photomicrograph by later blocky, equant spar (BS). Faint growth zonations are visible internally. (Texaco-Mobil core, –6,305 ft.)

mm, and are commonly comprised of a dirty, inclusion-rich core and a limpid outer rim. This dolomite is slightly ferroan, and has dull red luminescence. Crystals of this dolomite are noticeably larger than those of the clay-associated hypidiotopic dolomite described below, which they otherwise closely resemble.

Sucrosic dolomite is a common constituent in the fractured and brecciated host rock of the Getty-Cobb core. It is suggested that this dolomite is a product of eogenetic diagenesis.

Clay-Associated Hypidiotopic-Rhombic Dolomite

The presence of inclusions of clay in this matrix-replacive dolomite suggests that it formed by replacement of a precursor lithology. This dolomite is limited in areal extent (on a millimeter or centimeter scale) to terrigenous clastic clay-rich seams in the carbonate host (Fig. 12C). This type of dolomite is typically fine crystalline, ranging in size from 0.05 to 0.20 mm. Individual crystals may be inclusion-free (clear throughout) or inclusion-rich (dirty throughout), or may exhibit fine growth zonation due to the inclusion of impurities imparted from the replaced host during crystal growth. Little or no mineralogical variance was observed across inclusion-rich and inclusion-free zones, and the dolomite tends to have an even, dull red luminescent quality. This type of dolomite was encountered in the Cox-Wesley and Shell-Wesley cores from Carter County.

Saddle-Rhombic Dolomite

This interesting dolomite texture was observed only in the Shell-Wesley core. This type of dolomite is characterized by medium to coarse (0.4–1.0 mm) subhedral crystals that form a well-cemented, interlocking hypidiotopic fabric, with interstitial silty clay. The crystals are uniquely shaped, appearing rhombic in outline but with curved crystal faces reminiscent of pore-filling saddle dolomite cement (Fig. 12D). Radke and Mathis (1980) suggest that this type of dolomite is a matrix-replacive form of saddle dolomite, which is supported by the observation that the dolomite in this study contains inclusions of clay and quartz silt. Extinction is strongly undulose, often with opposing corners going extinct at the same time. This type of dolomite is characteristically growth zoned, although its luminescence is usually a non-zoned dull red.

Xenotopic Dolomite

The term "xenotopic" was proposed by Friedman (1965) for diagenetically altered carbonate rocks having a fabric comprised predominantly of anhedral crystals.

Fine- to medium-crystalline xenotopic dolomites were observed in the Texaco-Mobil, Cox-Wesley, and Shell-Wesley cores. Crystals range in size from 0.1 to 0.5 mm, and are usually arranged in a tightly interlocking fabric with little or no intercrystalline porosity. "Ghosts" of spherical and elliptical grains (perhaps ooids) were observed within the dolomite crystals, attesting to its replacement origin (Fig. 12E). In addition to replacing grainstones, the uniformly dirty appearance of some of this dolomite suggests that it also replaced wackestone/mudstone (Fig. 12F). Individual anhedra tend to be smaller in the mudstone-replacive variety, although both varieties tend to appear dull red under cathodoluminescence.

Anhedral crystals that grow into pore space often exhibit subhedral to euhedral terminations.

Additionally, some of the anhedra exhibit faint internal rhombic growth zonation, which suggests that crystallinity diminished as the dolomite crystals grew together in competition for the available pore space. This feature is more commonly observed in those xenotopic dolomites that replace grainstone than in the wackestone/mudstone-replacive variety. It would appear that the space available for crystal growth and the initial lithology of the replaced host carbonate exert some degree of control over the texture of the final dolomite.

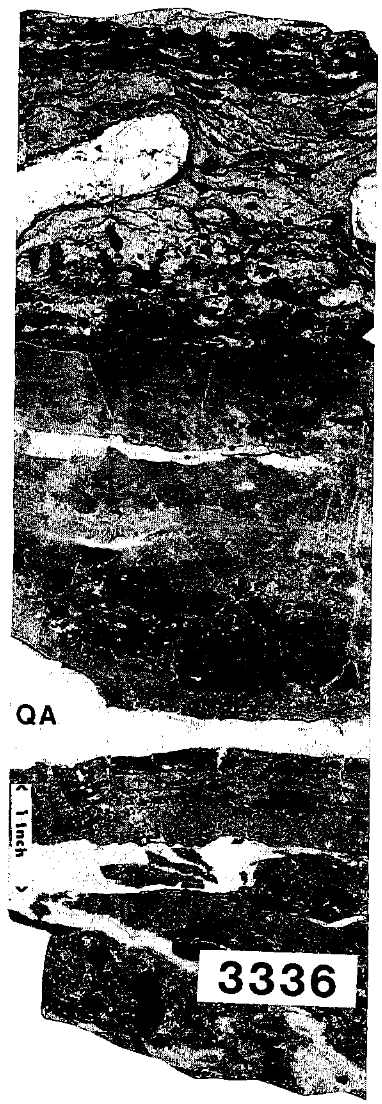

Figure 11. Solution channels filled with quartz-arenite sandstone (QA) of the overlying Burgen Formation (Simpson Group). (Oliphant-Lafortune core, –3,336 ft.)

Pore-Filling Dolomites

Three different morphologies of pore-filling dolomite cement were observed in the study cores. These dolomites exploit many different types of porosity, including vugular, channel and fracture porosity, and breccia-intergranular porosity.

Saddle Dolomite

Spear-type saddle dolomite (Radke and Mathis, 1980) occurs as a void-filling cement in the Cox-Wesley, Shell-Wesley, and Texaco-Mobil cores from Carter County, and in the Cameron-Shepherd core from Jefferson County. Typically, it occurs as subisopachous crusts of hypidiotopic to idiotopic crystals oriented normal to pore walls or grain boundaries (Fig. 13A). Saddle-dolomite crystals, which range in size from 0.5 to 5.0 mm, commonly exhibit pronounced undulose extinction. Individual crystals may be growth-zoned or, less commonly, clear or dirty throughout. Under cathodoluminescence, this type of dolomite exhibits subtle zonation comprised of dull red, bright red, yellow and non-luminescent laminae, attesting to the slight variance in chemical composition across each layer (Fig. 13B).

Radke and Mathis (1980) postulate that saddle dolomite forms at temperatures >80°C, and thus would imply a deep burial origin—or a hydrothermal origin at shallower depths—for this dolomite. It is interesting to note that the isotopic signature obtained for the saddle dolomites in this study (from rocks at –3,000 to –6,000 ft) is nearly identical to that observed by Lee and Friedman (1987) in saddle dolomite from Ellenburger carbonates at –15,000 to –20,000 ft.

Columnar Dolomite

This type of dolomite is similar in appearance to the saddle dolomite described above, except that the columnar variety has a greater length-to-width ratio than the saddle variety. Columnar dolomite also occurs as normal-to-substrate isopachous crusts (Fig. 14). Strong, sweeping undulose extinction characterizes this dolomite. This type of dolomite is predominantly inclusion-rich, appearing dirty throughout. This dolomite also exhibits dull red luminescence. Columnar dolomite was only encountered in the "zebroid" breccia interval in the Cox-Wesley core from Carter County.

Idiotopic-Rhombic Dolomite

Clear, equant rhombs of dolomite occur in the vugular pore space of the Oliphant-Nate, Cameron-Shepherd, and Pan American-State cores. Individual crystals range from 0.5 to 5 mm in diameter, have straight to slightly undulose extinc-

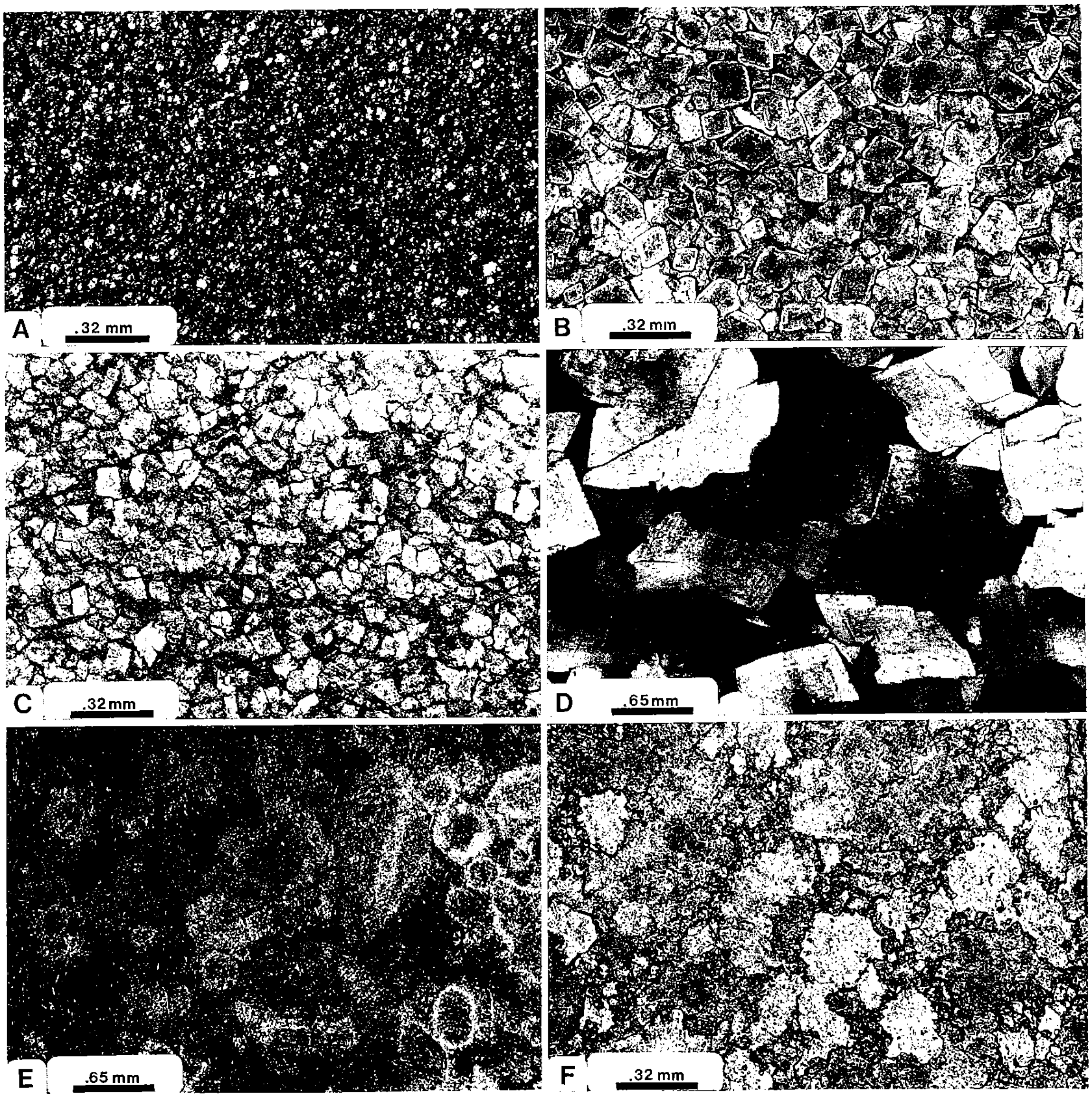

Figure 12. Thin section photomicrographs of host-rock-replacive dolomite types: *A*—host dolomicrospar, *B*—sucrosic dolomite, *C*—clay-associated hypidiotopic-rhombic dolomite, *D*—saddle-rhombic dolomite, *E*—grainstone-replacive xenotopic dolomite, *F*—mudstone-replacive xenotopic dolomite. (All photos taken under plain polarized light.)

tion, and are typically nonluminescent. They occur as vug-lining, though not isopachous, cements (see Fig. 9A). Individual crystals are not interlocked to the extent observed in the saddle dolomites.

Dolomite Paragenesis

Paragenesis of the various dolomite types was inferred from the mutual cross-cutting relationships exhibited by types, by their occurrence relative to the structural fabric of the rock, and by consideration of certain crystal textures that are known to be temperature-dependent. Although dolomite-formation temperatures have not yet been verified by fluid-inclusion analyses, preliminary investigations into the stable-isotopic signature of these dolomites suggests that most of the dolomites formed at elevated temperatures. Additionally, the mineralogy of all dolomite types, as indicated by energy dispersive X-ray analysis, is slightly ferroan.

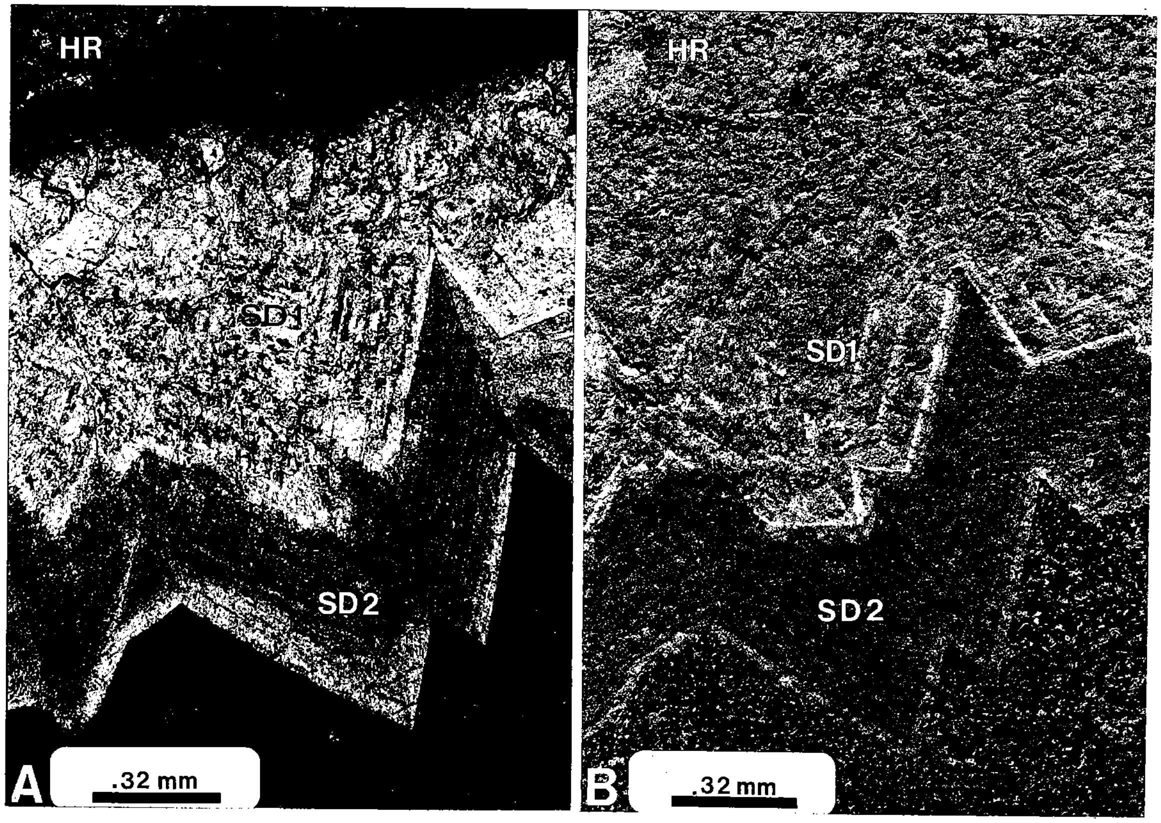

Figure 13. Thin-section photomicrographs of pore-filling saddle dolomite: At least two generations cf cement are present and discernible under plain light (*A*) and cathodoluminescence (*B*). First-generation cement (SD1) is adjacent to host rock (HR), is solution pitted and has less-distinct growth zoning, and has alternating bright red and yellow luminescent zones. Second-generation cement (SD2) was precipitated as an overgrowth on the earlier cement. The second-generation dolomite exhibits more-intense growth zoning, and dull red luminescence.

The first dolomite to form was the fine-crystalline dolomicrite/dolomicrospar, either as sedimentary dolomite within the environment of deposition or by replacement of lime mud at shallow burial depths. The occurrence of the sucrosic hypidiotopic dolomite within oomoldic porosity and replaced matrix carbonates suggests that it is a product of eogenetic diagenesis. The clay-associated hypidiotopic-rhombic form probably began forming within clay-rich seams subsequent to increasing burial; the concentration of this dolomite along clay seams may reflect incipient stylolite formation. With increasing depth of burial, Arbuckle Group rocks were exposed to elevated formation temperatures and perhaps to heated subsurface brines; at this time the xenotopic dolomite probably began to form. Gregg and Sibley (1984) suggest that xenotopic dolomite precipitates at temperatures > 50°C. The idiotopic-rhombic pore-filling dolomite cement may have precipitated from fluid of similar, or slightly cooler, temperature.

The precipitation of columnar and saddle dolomite probably commenced in stages, from fluid heated to >80°C (Radke and Mathis, 1980). This would tend to suggest that this dolomite formed either at depths sufficient to encounter formation temperatures >80°C, or that heated subsurface brines ascended through available conduits such as fractures to dolomitize the rock at shallower depths. Hagni (1976) has shown that Mississippi-Valley-type ore deposits nearly always contain sparry saddle dolomite as the most prevalent gangue mineral, and that the single most common factor among this type of deposit is their occurrence at or near the surface. The ubiquitous presence of growth zonation in this dolomite type suggests that it accreted layer by layer, over an incalculable period of time. Petrographic evidence strongly suggests episodic migration of a fluid capable of periodically dolomitizing and corroding the carbonate host (see Fig. 15).

The absolute timing of the pervasive saddle-type dolomitization is difficult to assess. It is cer-

Figure 14. Thin-section photomicrograph of pore-filling columnar dolomite. This cement occurs as drusy, isopachous crusts on laths of host rock (HR) in the zebroid breccia interval, in the Cox-Wesley core. (Cross polarized.)

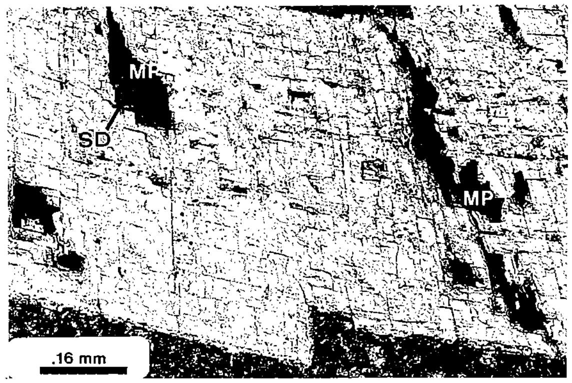

Figure 15. Thin-section photomicrograph of solution-pitted saddle-dolomite cement. Silty dolomite (SD) geopetally infills the solution micro-porosity (MP). (Plain polarized light.)

tainly plausible to propose that the extensive fracture systems which developed during Pennsylvanian tectonism could have acted as conduits for ascending, dolomitizing hydrothermal fluids. This premise is substantiated by Kranak (1978), who recognized Pennsylvanian-age Mississippi-Valley-type sphalerite mineralization in the Butterly Dolomite in Murray County, Oklahoma. It is likewise possible that the dolomitization could have occurred earlier by the same general process. We suggest that the petrographic and structural data can explain the episodic ascension of dolomitizing hydrothermal fluids, subsequent to the collapse stage of karstification, and Pennsylvanian tectonism. Dolomitization was probably a long-ranging process that continued throughout the burial history of the rock.

ARBUCKLE PALEOKARST MODEL AND EVOLUTION

Karst facies in the Arbuckle Group cores of this study are typified by collapse breccias and dissolution features; low-temperature speleothemic cements were not observed in appreciable amounts. The paucity of karstic cements may indicate that Arbuckle paleokarst developed in a semi-arid environment with a low water budget.

The establishment of karst probably began with the dissolution of massive Arbuckle carbonates along open ground-water conduits such as fractures and bedding planes. Solution-enlarged fractures would develop into channel porosity, and perhaps into cavern porosity. Although cavernous porosity was not directly observed in this study, its former presence can be inferred from extant features such as collapse breccias. Certainly, the 20+ ft of heterolithic collapse breccia in the Pan American-State core necessitates a precursor cavity of similar vertical extent, into which the roof rock was able to collapse. The lateral extent of such features is perhaps unknowable, but for comparison purposes it is interesting to note that Curtis (1959) has cataloged present-day Arbuckle caves in Murray County that extend as much as 14,050 ft in total length. Such caves typically exhibit linear trends that may indicate a joint-controlled origin. It has been suggested that karst-facies elements such as collapse breccias may attain thicknesses of 50–100 ft, and may be present hundreds of feet beneath a paleoexposure surface (Kerans, 1988).

The timing of Arbuckle paleokarstification is less obvious than the features that result from it. Data presented in this and other studies suggest that the paleokarstification of Arbuckle strata may have been an episodic phenomena that occurred (in varying degrees) when Arbuckle rocks were exposed at local or inter-regional unconformities. Numerous authors (i.e., Gore, 1952; Ireland, 1955; Merriam and Atkinson, 1956; Walters, 1958; Kerans, 1988) have documented the presence of paleokarst in Arbuckle and equivalent rocks at the pre-Simpson unconformity in Oklahoma, Texas, and Kansas. Walters (1958) has shown that the porosity present in Arbuckle reservoirs along the central Kansas uplift is directly related to Pennsylvanian karstification that probably modified pre-Simpson paleokarst.

Data presented in this investigation suggests that the establishment of Arbuckle paleokarst may be due to multiple episodes of subaerial exposure and erosion during, and subsequent to, the time of Arbuckle deposition. The presence of similar and biostratigraphically equivalent conodont species (Ritter, personal communication, 1989) in both the matrix and clasts of cavern-fill parabreccia and collapse breccia, and the complete lack of younger (Simpson, Pennsylvanian, etc.) fauna suggests that karstification occurred during Arbuckle time. An argument for karstification at the pre-Simpson unconformity can be made for the cores from Osage County that contain Simpson-derived sandstone in solution channels and vugs. Note that these cores also exhibit intraformational disconformities and sedimentary breccias that are clearly of Arbuckle age. With the exception of a thin, sandy residual zone at the top of the Getty-Cobb core, no evidence for Pennsylvanian erosion was found in the cores of this study. This in no way negates the possibility of Pennsylvanian karst occurring in other areas where Arbuckle rocks subcrop beneath the pre-Permian unconformity (for example, see Donovan, 1987). Walters (1958) speculated that the entire thickness of the Arbuckle Group in Kansas served as a vast regional aquifer during the Pennsylvanian, much as it does presently in the Tri-State region of Kansas, Missouri, and Oklahoma (Macfarlane and Hathaway, 1987).

Preservation of karst porosity appears fortuitous and uncommon, based on the rocks in this study. Solution-channel porosity is commonly occluded by infill sediment, while the interparticle porosity in collapse breccias is typically cemented or filled with sediment. Data from the Oliphant-Nate core suggests that migration of hydrocarbons into solution channels and vugs effectively arrested dolomite cementation, thereby preserving the open pore network. Much of the effective porosity currently present in these rocks appears to be intimately related to diagenetic modification of the karst profile as it was buried and then structurally modified by later tectonism.

SUMMARY AND CONCLUSIONS

Paleokarstification of upper Arbuckle Group rocks is evidenced in cores from Oklahoma by (1) breccia facies that form as a result of collapse, subterranean sedimentation, and surficial erosion; (2) vadose cements (infrequently); and (3) dissolution structures, such as enlarged vugular and channel porosity, and solution-enlarged fractures. The presence of Arbuckle-age conodonts in breccia matrix (and the complete lack of younger fauna) suggests that much of the karstification observed may be of Arbuckle age.

A complicated burial-diagenetic history for these rocks is indicated by the variety and paragenesis of matrix and pore-filling dolomite types. Eogenetic, matrix-replacive dolomite types suggest early diagenesis, while coarse-crystalline, xenotopic matrix-replacive varieties indicate burial diagenesis at temperatures >50°C. Pore-filling saddle and columnar dolomites likely precipitated from ascending hydrothermal fluids that may be related to hydrocarbon migration. That these fluids were alternately constructive (precipitating) and destructive (dissolving) is indicated by the presence of compositionally zoned and solution-pitted dolomite cements. Breccia-intergranular porosity (where preserved) and tectonic fractures provided the necessary conduits for fluid migration. Dolomite cementation was apparently long-ranging, but was arrested by the migration of hydrocarbons into pore space.

ACKNOWLEDGMENTS

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THE ARBUCKLE GROUP—RELATIONSHIP OF CORE AND OUTCROP ANALYSES TO CYCLIC STRATIGRAPHY AND CORRELATION

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Correlation of ramp-type and even platform-type carbonates is possible with an understanding of the relationship of depositional environment and related sequences and boundaries to log response. Hopefully, subtle relationships of facies to log character can be used to develop not only local correlations but regional correlations as well.

Lower Paleozoic carbonates such as the Arbuckle Group and Hunton Group which are dominated by ramp- to platform-type peritidal facies can be subdivided into cycles based on facies changes and depositional sequences (Fig. 1). Researchers have usually related these subdivisions to subsidence and eustatic sea-level changes. Although not thoroughly understood, episodic tectonic activity has been assumed to be negligible on such broad and stable shelves.

Depositional sequences commonly are in some form of the following upward-shoaling order of facies: open marine—subtidal—intertidal—supratidal. Open-marine facies are argillaceous to "marly" limestone; delicate, thin-shelled fossil fragments within such deposits are generally intact. Within the Arbuckle Group, open-marine facies have not been recognized; however, in the Hunton this facies is relatively common. Subtidal facies are less argillaceous and usually bioturbated. In the upper subtidal, shoals of bioclastic and oolitic material are also present. The Hunton is composed predominantly of this type of rock, and large amounts of the Arbuckle are also shallow subtidal. Upper subtidal/lower intertidal, a further facies subdivision within the Hunton, is dolomitized and preferentially develops porosity (Fig. 2). The intertidal facies normally consists of millimeter-scale laminations; within the Arbuckle, algal stromatolites are common in this facies. Supratidal facies, uncommon in both groups, occur more frequently in the Arbuckle. Thin laminae of anhydrite, desiccation features, and fenestral features are characteristic of the supratidal facies.

In the Silurian portion of the Hunton Group, core analyses of sequences (when compared to log signature) indicate that the higher energy facies, such as the upper subtidal/lower intertidal, tend to contain fewer siliciclastics, and therefore have a "cleaner" gamma-ray response corresponding with higher resistivity. Conversely, lower energy facies, such as the open marine or upper intertidal, often contain more siliciclastics and generally have a "dirtier" gamma-ray response and lower resistivity. This pattern is also known in many other sequences, such as in the Lodgepole of the Williston basin. The Frisco Formation (Lower Devonian) of the upper Hunton Group is composed of bryozoan- and crinoidal-rich mud-mound complexes with no dolomite or siliclastics; the result is a very "clean" gamma-ray response.

In the Arbuckle, however, the relationship of facies to log response is not always clear, because the facies tend to be restricted to upper subtidal and intertidal environments which do not always show a marked change in clastic input. In many areas, pervasive dolomitization of all facies masks the original textures and the resulting log characters. In general, within the Arbuckle, siliciclastic content is usually greater in the intertidal deposits where it is either dispersed wind-blown clay, silt, and very fine sand or sand and silt concentrated in shallow tidal channels. Such thin clastic intervals cause high gamma-ray deflections which should be useful in correlations. These are already known to be useful markers in the Ordovician Trenton of the Michigan basin and in the Siluro-Devonian (Hunton equivalent) in west Texas. Another subregional marker is a black shaly dolomite present in the Pan American Tackett well and correlatable through much of the Arkoma basin (Fig. 3).

Once the relationship of facies to log response is established, minor sequences can be correlated locally and groups of sequences, bracketed by third-order eustatic sea-level changes, can be correlated regionally. It is hoped that such moderate-scale sequence groupings in the Arbuckle by the use of Fischer diagrams can be used as a regional or even an interregional correlation tool, following the lead of recent studies of Cambrian-Ordovician strata in the Appalachians and west Texas.

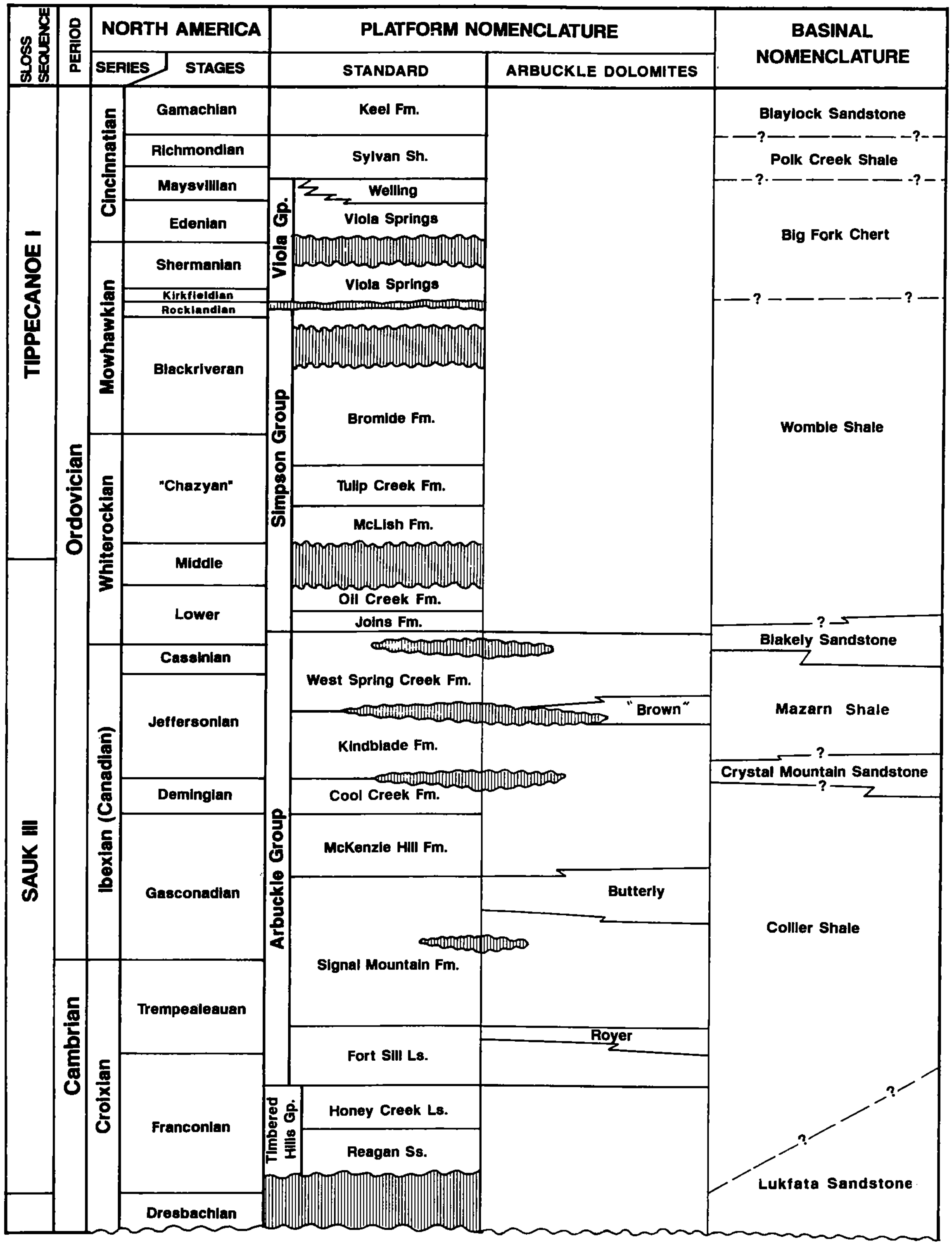

Figure 1. Lower Paleozoic stratigraphic column for Oklahoma, showing platform and basinal (Ouachita trough) nomenclature.

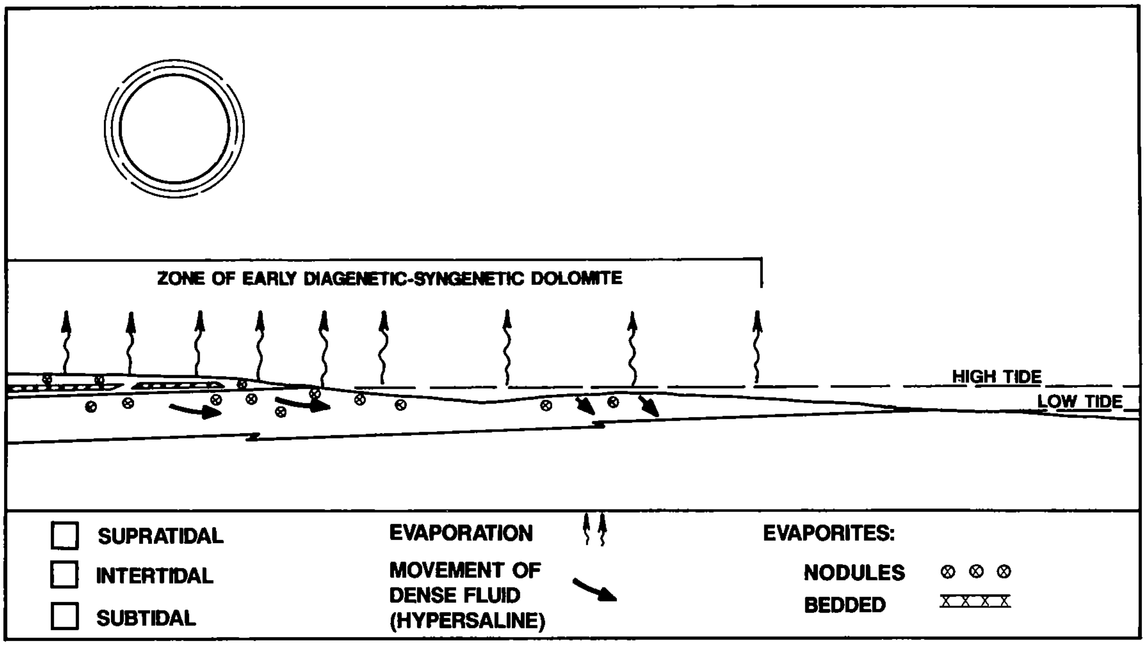

Figure 2. Syngenetic dolomitization.

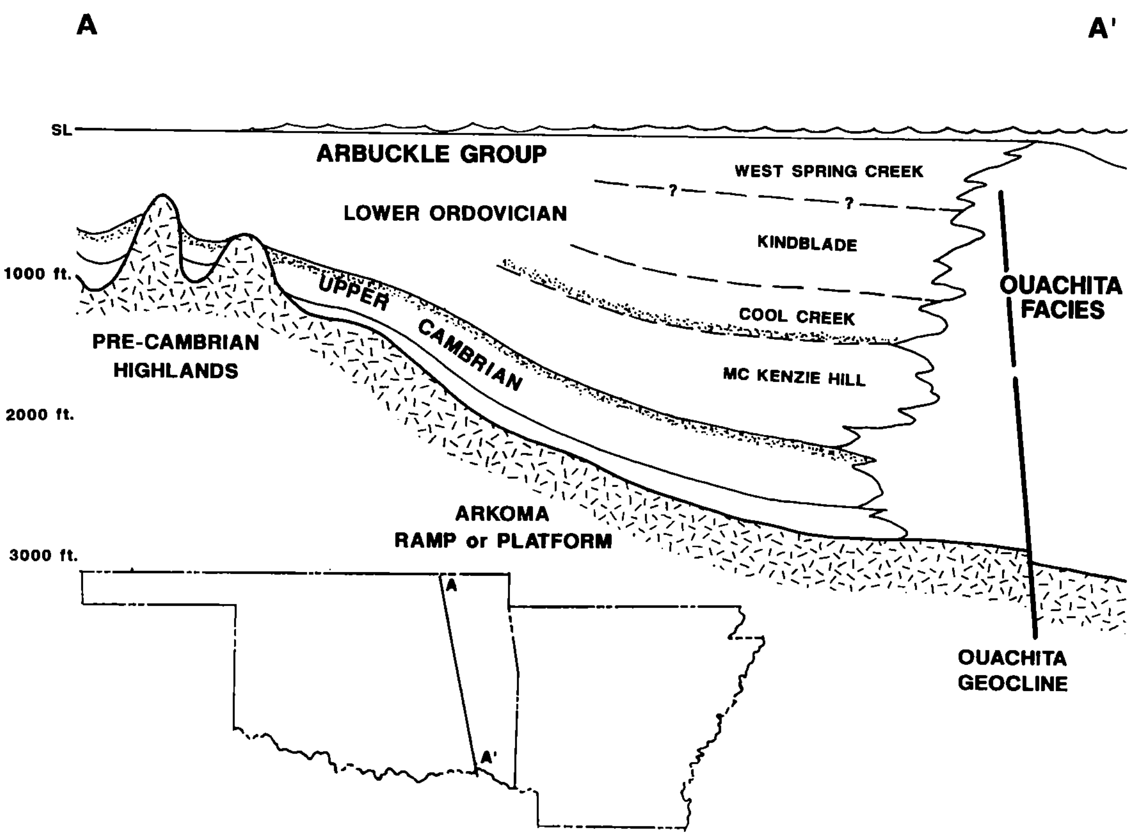

Figure 3. Correlation of the Arbuckle Group across the Arkoma basin.

PETROGRAPHY AND GEOCHEMISTRY OF MASSIVE DOLOMITE FROM THE UPPER ARBUCKLE GROUP, SLICK HILLS, SOUTHWESTERN OKLAHOMA

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ABSTRACT.—Petrographic and geochemical analyses of massive dolomite units (400 m thick) from the upper Arbuckle (Early Ordovician) shallow-water marine carbonate sequence, Slick Hills, southwestern Oklahoma, define two distinct kinds of dolomite. Dolomite I, with exposed thickness up to 350 m, displays dull luminescence, or is nonluminescent. Dolomite I is characterized by nearly stoichiometric compositions (average 50.4 mol% $CaCO_3$ and 49.6 mol% $MgCO_3$), low Fe and Mn contents (average 0.05 mol% Fe and 0.01 mol% Mn), light $\partial^{18}O$ values (average –7.3‰, PDB; Fig. 1), and slightly lower $^{87}Sr/^{86}Sr$ ratios (average 0.70879) relative to associated limestone (average 0.70892). Volumetrically minor dolomite II has bright-red luminescence and occurs mainly in transition zones between dolomite I and associated limestone. Dolomite II is characterized by Ca-rich compositions (average 52.0 mol% $CaCO_3$ and 47.5 mol% $MgCO_3$), high Fe and Mn contents (average 0.30 mol% Fe and 0.20 mol% Mn), heavy $\partial^{18}O$ values (average +1.7‰, PDB), and radiogenic $^{87}Sr/^{86}Sr$ ratios (average 0.70925) relative to associated limestone.

Different petrographic and geochemical characteristics reflect different formation histories of the two types of dolomite. The similarity of $^{87}Sr/^{86}Sr$ ratios of most dolomite I samples to that of Early Ordovician seawater, as well as the thick, massive occurrence of dolomite I in a shallow-water marine carbonate sequence, suggest that dolomite I probably originated from seawater during early diagenesis. However, the light $\partial^{18}O$ values of dolomite I and the slightly lower $^{87}Sr/^{86}Sr$ ratios relative to Early Ordovician seawater demonstrate that dolomite I was modified, probably during burial. ^{18}O-depleted meteoric water, buffered by younger carbonates with $^{87}Sr/^{86}Sr$ ratios lower than Early Ordovician seawater, caused the modification during formation of numerous unconformities which characterize the younger rocks.

Dolomite II occurs sometimes as cements within dolomite I, and was formed later than dolomite I. The dominantly radiogenic $^{87}Sr/^{86}Sr$ ratios of dolomite II require that dolomite II formed from ^{87}Sr-enriched fluids. The heavy $\partial^{18}O$ values and high Fe and Mn concentrations require that dolomite II formed from ^{18}O-enriched fluids under reducing conditions in the shallow subsurface. These late fluids probably derived their magnesium from dissolution (solution/compaction) of deeply buried Arbuckle dolomite and were probably derived from the Anadarko basin, north of the Slick Hills. The late fluids migrated through porous dolomite I strata and dolomitized the limestones near limestone/dolomite I contacts, resulting in the distribution of dolomite II in transition zones between dolomite I and associated limestone.

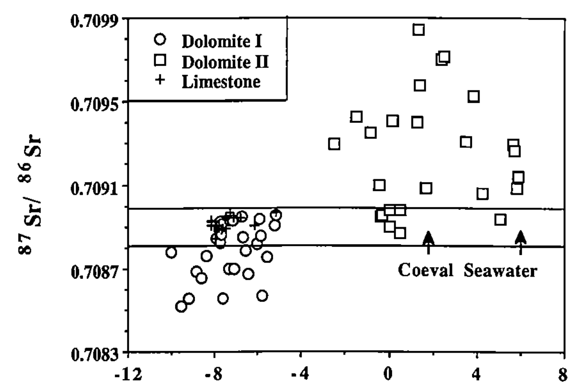

Figure 1. $\partial^{18}O$ versus $^{87}Sr/^{86}Sr$ plot. Horizontal lines define the $^{87}Sr/^{86}Sr$ of coeval seawater (Burke and others, 1982).

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Oklahoma Geological Survey Circular 92, 1991

RECENT DEVELOPMENTS AT WILBURTON FIELD, LATIMER COUNTY, OKLAHOMA

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Recent deeper drilling in Wilburton Field has resulted in new production from *in situ* Spiro and Cromwell sands, and from the Arbuckle Group in a structurally high fault block beneath the old-field pay in the overthrusted Spiro. ARCO Oil and Gas first found production from *in situ* Spiro beneath a low-angle thrust fault in sec. 36, T. 5 N., R. 17 E. As of the summer of 1989, nine wells now produce from the reservoir. Two wells have been completed below the Spiro in the *in situ* Cromwell reservoir, and others are thought to be productive. Three wells are currently drilling for Spiro and Cromwell.

Structural data generated by drilling the *in situ* Spiro and Cromwell pays revealed a southwest dipping, upthrown fault block with nearly 2,000 ft of vertical relief. ARCO Oil and Gas tested this feature with their #2 Yourman in sec. 15, T. 5 N., R. 18 E., discovering prolific gas production from the Arbuckle Group. The Arbuckle reservoir consists largely of vugs and solution-enlarged fractures in a clean dolostone. Conventional neutron-density logs show very low porosity, but cores and Formation Microscanner images reveal the vugs, some of which are several inches across. Permeability is created by a pervasive system of younger fractures and microfaults. All current production is from the uppermost 700 ft of the Arbuckle, with a gas column of at least 1,100 ft. As of mid-1989, six Arbuckle wells have been completed, two are in completion, and two are drilling.

Oklahoma Geological Survey Circular 92, 1991

SEISMIC EXPLORATION FOR CAMBRIAN–ORDOVICIAN OBJECTIVES, WILBURTON AREA, SOUTHEASTERN OKLAHOMA

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ABSTRACT.—Recent gas discoveries near Wilburton show that Arbuckle carbonates can be prolific reservoirs. Production is from structurally high blocks bounded by near-vertical faults. Since there is no direct relationship between shallow and deep structure, new Arbuckle trends and prospects will be defined using high-quality seismic data.

Modern seismic acquisition in the area is done with a dynamite source, a large number of recording channels, and closely spaced geophone groups. Data quality is better north of the Choctaw fault, where the structure is simpler. Greater effort is required to image the complex geology farther south.

In processing, migration velocities must be selected carefully, as they affect the apparent size of a structure. Seismic interpretation should be based on simultaneous examination of migrated and unmigrated sections, and care should be taken when interpreting sections which exhibit sideswipe.

Recently acquired seismic data illustrate the structural style in the Wilburton area.

INTRODUCTION

For the last two years the Ouachita frontal fairway of southeastern Oklahoma (Fig. 1) has been one of the most active onshore exploration provinces. The major exploration targets have been Spiro and Wapanucka reservoirs on thrusted structures, and Arbuckle reservoirs on structurally high blocks. The prolific Cambrian–Ordovician wells at Wilburton field have particularly captured the imagination and have spurred the search for new prospective structures. This paper reviews the structural style of the Wilburton area and summarizes the elements of seismic exploration for Arbuckle objectives. Recent papers addressing seismic exploration or interpretation of this area include Perry and Suneson (1990), Reeves and others (1990), and Milliken (1988).

GEOLOGIC MODEL AND SEISMIC EXPRESSION

Bertagne and Leising (1990) presented a model for the Wilburton area based on interpretation of 100 mi of recent seismic data. They identified four tectonic units (Fig. 2):

Unit 1
Broad Synclines/Tight Anticlines North of the Choctaw Fault

Unit 1 is composed of Atokan and younger rocks. The lower boundary of the unit is a décollement near the base of the Atoka. This décollement shallows to the north. Seismic events from the synclines are continuous, whereas the tight anticlines are underlain by triangle zones lacking coherent reflections. These zones are inferred to consist of a series of N-vergent thrust faults, which cause thickening of the sedimentary section, overlain by a backthrust.

Unit 2
Steeply Dipping Beds Above the Choctaw Fault

The Choctaw fault, which marks the base of unit 2, juxtaposes rocks of similar velocities and so does not have a strong seismic expression. A further consequence of the velocity distribution is that significant velocity pullups are not observed beneath the Choctaw fault. This is in marked contrast to the major faults of many other thrust belts.

Seismic data from above the Choctaw fault generally exhibit steep, S-dipping events. The top of the Wapanucka produces a strong reflection which is readily identified even in highly deformed areas.

Unit 3
Thrusted Spiro–Wapanucka–Cromwell Strata Beneath and North of the Choctaw Fault

The top of this unit is the same décollement which marks the base of unit 1. The base of unit 3 is a décollement near the basal Springer shale.

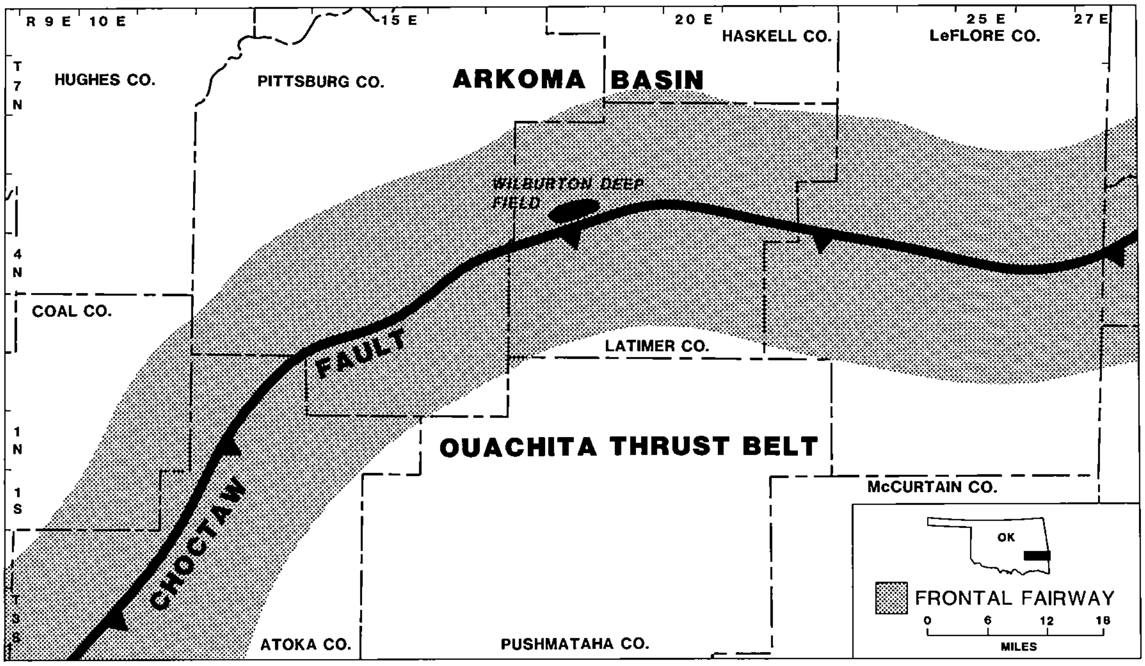

Figure 1. Location map of Wilburton field and Ouachita frontal fairway.

Spiro, Wapanucka, and Cromwell strata dip to the south and are repeated by N-vergent thrust faults. The thrusted structures are often easily identified because of the strong reflection generated at the Spiro/Wapanucka interface.

Unit 4
Deep Fault Blocks

Unit 4 is composed of deep blocks bounded by near-vertical faults which offset the lower Atokan through Precambrian (Figs. 3,4). These faults are commonly down-to-the-south and normal, although reverse faults are also observed. The blocks are generally identified by tracing the limits of the reflection correlating to the Viola. Occasionally, this event is hard to discern, making interpretation difficult. The absence of the event may be due to facies changes which result in smaller reflection coefficients. The deep blocks appear to have influenced the location of younger thrust fault ramps.

There is continuing discussion of whether the northern field-bounding fault at Wilburton is normal or reverse. Recent seismic data (Fig. 3) suggest a normal fault, although selected reflections within the greater Wilburton block may indicate reverse faulting. The controversy should be resolved when relevant well data have been released.

A major down-to-the-south normal fault, offsetting the Arbuckle section by thousands of feet, is present south of Wilburton, beneath the Choctaw fault. This fault will probably mark the southern economic limit of the Arbuckle play in the Wilburton area.

SEISMIC EXPLORATION
Acquisition

Modern seismic data are acquired with a dynamite source, a large number of recording channels, and closely spaced geophone groups. When designing an Arbuckle seismic program, line location and acquisition parameters should be optimized for the deeper section. When possible, dip lines should not be placed near tear faults, since this can interfere with data quality. Magnetic data can help locate these faults.

Data quality is better north of the Choctaw fault, where the overall structure is simpler; it is more difficult to image Arbuckle structure farther south. The data quality depends partly on topography: Better data are recorded on lines shot across topographic highs, whereas data acquired in valleys are commonly poor. As would be expected, increasing the fold and decreasing the group interval while maintaining a good offset distribution improves the data quality in difficult areas. It also significantly increases the cost of acquisition.

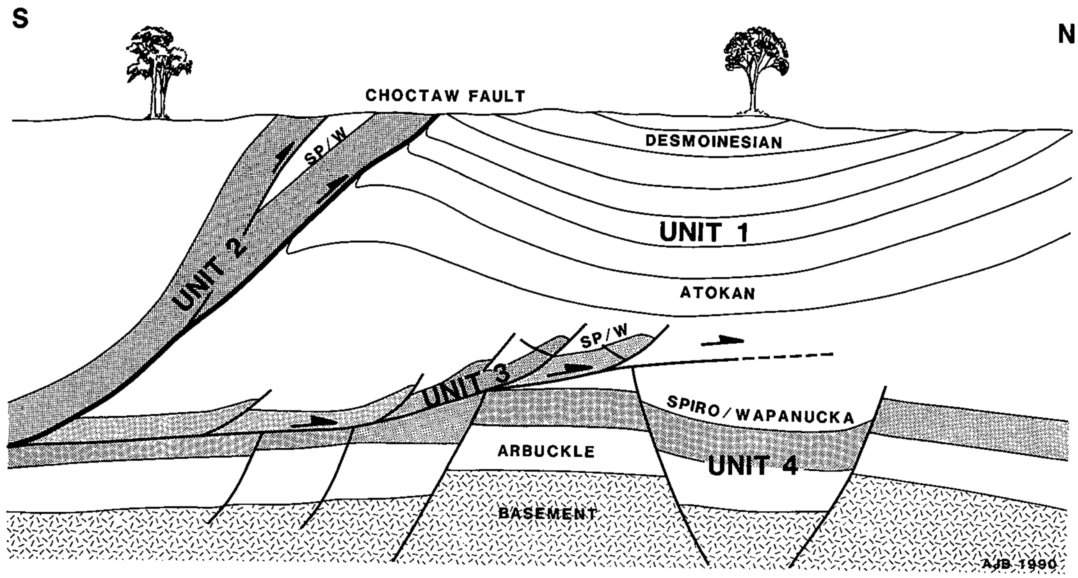

Figure 2. Schematic cross section showing structural style of the Wilburton area, and tectonic units discussed in the text.

As in other complex areas, the problem of sideswipe—that is, the recording of energy which is not from immediately below the surface seismic-line location—is encountered in the frontal fairway. The only complete solution to this problem is to record three-dimensional data; however, three-dimensional methods are typically considered only for field development. A possible compromise is to record multiline swaths (i.e., record on several geophone lines while shooting on a single source line). The net result is several closely spaced, separately processed CDP lines, which can give the true, three-dimensional dip of a strong seismic event. The cost of swath shooting is lower than three-dimensional acquisition.

Processing

Seismic processing is optimized when there is close interaction between the interpreter and the processor. Thus, critical decisions affecting the appearance of structures can be made with the interpreter's consent. Two steps where this input is particularly important are the migration and the stacking velocity analysis.

Migration (i.e., focusing of the seismic image) is probably the most sensitive step in the processing sequence. Careful selection of migration velocities is needed to avoid distortion or destruction of structures. Standard approaches for migrating data are: (1) simple post-stack migration, (2) dip moveout combined with post-stack steep-dip migration, and, (3) full pre-stack migration. These three processes are increasingly more "correct" and more computer-intensive. The interpreter needs to be familiar with the effect of each technique, as well as with their associated costs.

Other processing steps—such as the mid- and shortwave statics correction, post-stack deconvolution, and FX deconvolution—can also make important contributions to the overall data quality, depending upon the geology. However, the data are not as sensitive to these steps as to the migration and velocity analysis.

Interpretation

Since Arbuckle structures can be difficult to image, it is helpful to have as many displays as posssible when interpreting. Of particular importance are displays showing the different migration types, the effect of various migration velocities, and the effect of stacking selected parts of the cable (e.g., near-trace and far-trace stacks). In addition, we find that sections with a great vertical exaggeration provide a rapid overview and can allow new insights into the geology of the area.

As noted, sideswipe is commonly observed on seismic lines from the fairway. A typical manifestation is two seismic events which cannot be simultaneously interpreted in a geologically meaningful way. Since it is difficult to remove sideswipe on conventional lines, the interpreter must develop the sixth sense of knowing whether a given event is real or is sideswipe.

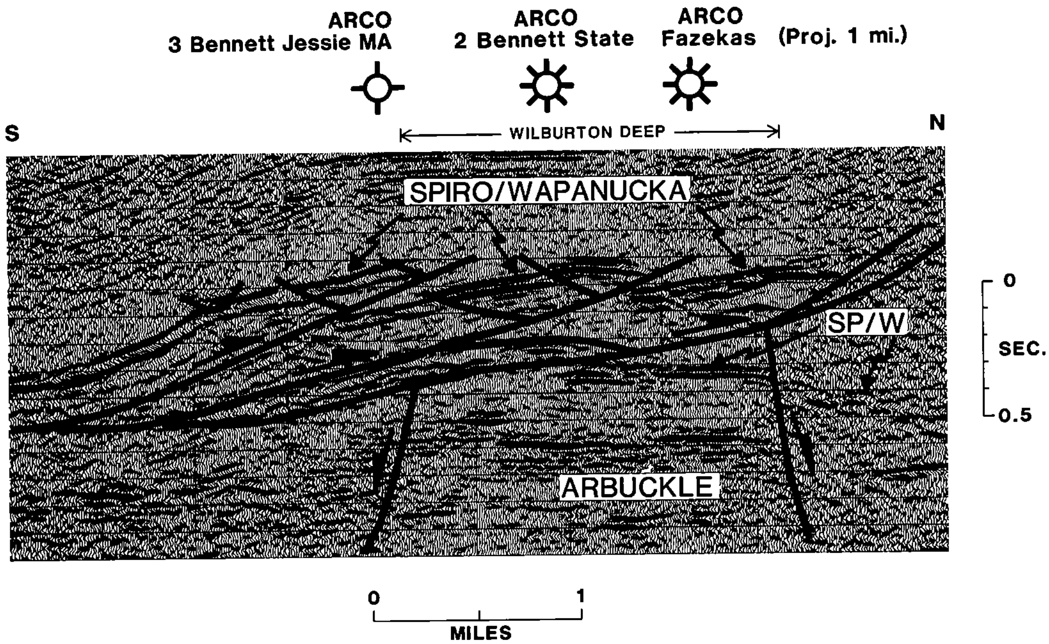

Figure 3. Seismic line showing the expression of the Wilburton field. Early production was from thrusted Spiro–Wapanucka structures; recent discoveries are from Arbuckle reservoirs.

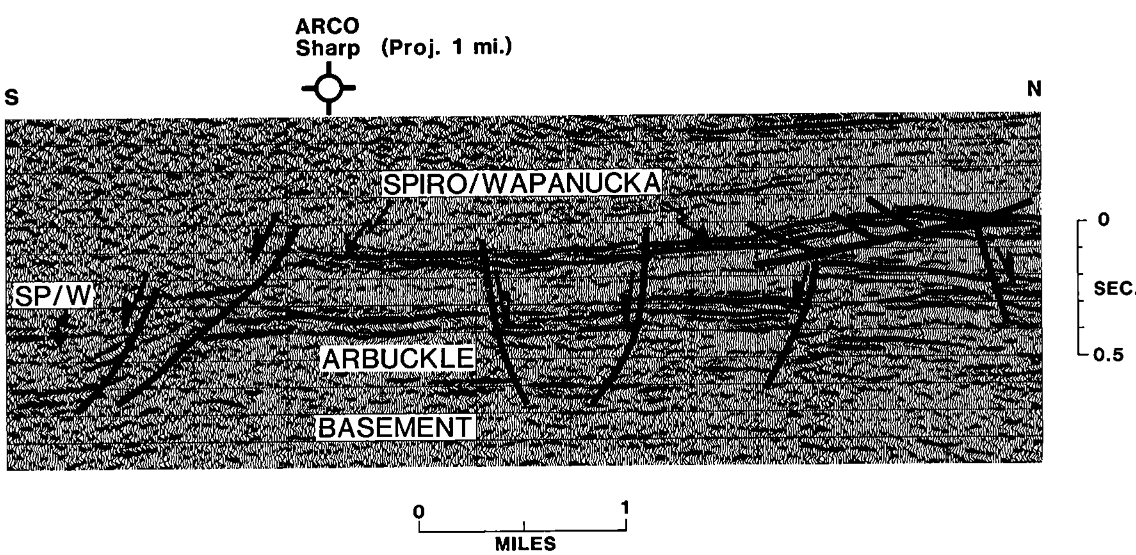

Figure 4. Seismic line located north of Wilburton, showing Arbuckle fault blocks.

CONCLUSIONS

The recent discoveries at Wilburton field have proved that Cambrian–Ordovician reservoirs of southeastern Oklahoma are important exploration targets. Competitive advantages will accrue to companies able to acquire high-quality seismic data, develop an optimized processing sequence, and work at the limits of interpretability of the resulting data.

Future exploration success in the Ouachita frontal fairway will come from imaginative interpretation of "difficult" seismic data, combined with an element of luck. In this respect, the Cambrian–Ordovician play is no different than numerous plays in other structurally complex areas.

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ARBUCKLE GROUP DEPOSITIONAL PARASEQUENCES, SOUTHERN OKLAHOMA

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ABSTRACT.—Outcrop and/or subsurface core studies of Fort Sill, Royer Dolomite, Signal Mountain, Butterly Dolomite, McKenzie Hill, Cool Creek, Kindblade, and West Spring Creek Formations reveal that most of the Arbuckle Group was deposited as a series of storm-dominated, shallowing-upward parasequences. These formations were deposited upon an extremely broad, nearly flat-lying carbonate ramp that formed the southern margin of the North American craton (Knox, Arbuckle, Ellenburger, and El Paso Groups) in the Late Cambrian and Early Ordovician.

Shallowing-upward parasequences were deposited in a cyclic manner, with individual fifth-order cycles only a few feet to tens of feet thick. These cycles record abrupt transgressions, caused by quick sea-level rise, that were followed by progradation of a paleo-shoreline as sea level gradually fell. Each cycle is subdivided into subtidal, near-shore to shoreline, and tidal-flat components. Subtidal and tidal-flat components can be of equal thickness or can be skewed with one component becoming dominant and the other subordinate. Intervening near-shore to shoreline settings are more common in incomplete shallowing-upward parasequences.

The transgressive base of each cycle often contains rip-up intraclasts that are occasionally imbricated and sometimes contain clasts eroded off the top of the previous cycle. The subtidal section is composed of medium to thick beds of either (1) bioturbated skeletal packstone or (2) mud-rich mudstone/wackestone; or the section is (3) stromatolitic/thrombolitic. Stromatolites/thrombolites, common in the Early Ordovician, occur as small, digitate features; low-lying, flattened mounds; club shapes; small domes or hemispheres; or broad and tall, domed shapes. Some thrombolites are so cryptic they can only be seen when silhouetted by intraclasts. Storm horizons of intraclasts and flat-pebble conglomerate are also common.

The transition from thicker subtidal beds into thinner tidal-flat beds is wavy to flaser-like bedding called "ribbon carbonate." Above ribbon carbonate are very thin beds, called "thick" laminations, which give way upsection to laminated bedding, called "thin" laminations. Thick laminations were deposited in embayments or ponds, or between segmented tidal flats. Thin laminations were deposited on tidal flats. Less-common intraclastic, ooid grainstone is found between thick and thin laminations. Supermature, bioturbated siliciclastics are also found between thick and thin laminations (West Spring Creek and Cool Creek Formations). Occasionally, siliciclastics were carried offshore by storms.

Only half of all parasequences are complete shallowing-upward events. Once understood, the vertical stacking of facies in a parasequence is predictable, so that complete and incomplete parasequences can be easily recognized. These distinctions are very important to recognize, since well-developed subtidal parts of a parasequence can form reservoir intervals when dolomitized.

INTRODUCTION

During the Late Cambrian through Early Ordovician, the North American craton was situated between 20° N. and 30° S. lat. (Fig. 1). The position of the craton placed it in a tropical to subtropical latitude. The southern margin of the craton was a broad, nearly flat-lying ramp (Fig. 2), which has been referred to as the "Great American Bank" by R. N. Ginsburg (Pray, 1981). The width of this ramp was as much as 250 mi in the Appalachians; 450 mi through the Midcontinent of Kansas, Oklahoma, and North Texas; and 200 mi across the Permian basin of southeastern New Mexico and West Texas. The carbonate ramp formed a belt which extended around the southern margin of the craton ~2,500 mi. Regional dip across the ramp was ~2 cm/km. Arbuckle regional dip is comparable to the present-day evaporite/carbonate setting along the Trucial Coast in the Persian Gulf, where average dip from the coastline across the shallow-marine ramp averages 2.5 cm/km.

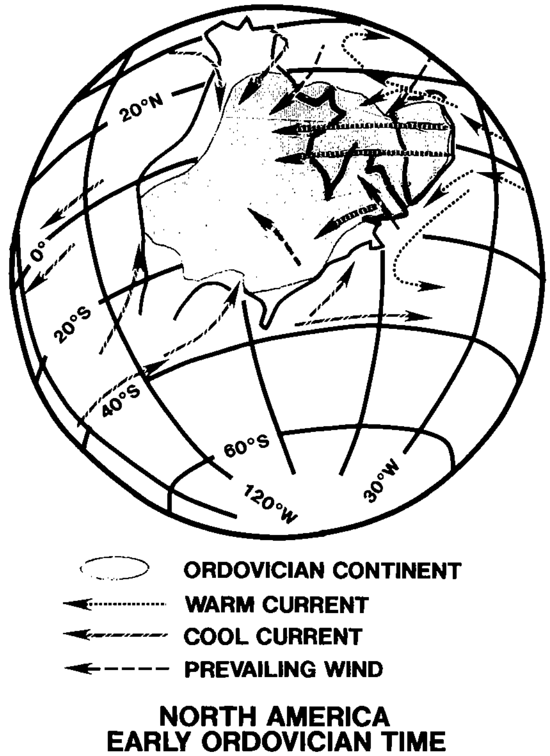

Figure 1. Early Ordovician position of the North American craton. Modified after Ross (1976).

Major Upper Cambrian–Lower Ordovician rock-unit names associated with various parts of this "Great American Bank," from east to west, are (1) Knox Group in the Appalachian Mountains, (2) Arbuckle Group in the Midcontinent, (3) Ellenburger Group in the Permian basin, and (4) El Paso Group in westernmost West Texas (Figs. 3–5). Along the western edge of the North American craton, in the Ibex area of western Utah, the Upper Cambrian is represented by the Orr and Notch Peak Formations, and the Lower Ordovician is represented by the Pogonip Group. The Arbuckle Group is subdivided into several formations; in ascending order, they are (1) Fort Sill Limestone, (2) Royer Dolomite, (3) Signal Mountain Formation, (4) Butterly Dolomite, (5) McKenzie Hill Formation, (6) Cool Creek Formation, (7) Kindblade Formation, and (8) West Spring Creek Formation (Fig. 6).

Deposition on the Great American Bank piled up an average of 1,000–3,000 ft of carbonate sediments (Fig. 3). However, in the southern Oklahoma aulacogen there was over-thickening (Arbuckle Group, 8,000 ft thick). The Arbuckle is dominantly limestone within the aulacogen and is extensively dolomitized outside of it. Also, Arbuckle solution-collapse breccias are far more common outside of the aulacogen than within it.

Several oil and gas fields are producing from Arbuckle and Ellenburger carbonates throughout the Midcontinent and the Permian basin. Because of the economic importance of these carbonates, a detailed study was initiated to gain a better understanding of Arbuckle deposition and diagenesis. This paper reports on sedimentologic results of outcrop and core study of the Arbuckle Group. Outcrop studies were conducted in the Arbuckle and Wichita Mountains (Figs. 4,5). Arbuckle core was studied from shallow burial depths (~4,000 ft) at Healdton field, and nonproductive deep burial depths (17,273–18,160 ft) in the Sun/Parker Mazur No. 1 (CSW¼NE¼ sec. 1, T. 3 N., R. 5 W., Grady County).

PALEOZOIC TECTONICS

The NW-trending southern Oklahoma aulacogen was first named by Shatski (1946) and is the deepest sedimentary trough in North America (Fig. 4). Ham and others (1964) named this same feature the "Southern Oklahoma geosyncline." In the Late Proterozoic or Early Cambrian, deposition of Tillman metasediments preceded formation of the aulacogen. This was followed by rift-related igneous intrusive and extrusive events in the late Early Cambrian through the Middle Cambrian (550–525 m.y.). The igneous activity was during initiation and development of the southern Oklahoma aulacogen (Gilbert and McConnell, this volume). Basalt/spilite, gabbro, rhyolite, and granite were either intruded (layered) or extruded within the developing aulacogen. As these rocks cooled and the aulacogen began to subside, they were initially covered by arkosic to subarkosic Reagan Sandstone and Honey Creek Limestone of the Timbered Hills Group in the Late Cambrian.

As subsidence continued in the Late Cambrian, through the Early Ordovician, and into the earliest Middle Ordovician, the aulacogen was filled by a thick sequence of Arbuckle Group carbonates. Following Arbuckle deposition, the Middle Ordovician through Early Mississippian was a time when shallow seas spread across the southern margin of the craton. Interspersed with carbonates are Simpson Group siliciclastic supermature sandstones in the Middle Ordovician, Sylvan Shale in the Upper Ordovician, and Woodford Shale in the Upper Devonian and lowermost Mississippian. The stratigraphic sequence above the Arbuckle Group consists, in ascending order, of the Simpson Group (siliciclastics, shales, and carbonates) in the Middle Ordovician; Viola Group carbonates and Sylvan Shale in the Upper Ordovician; Hunton Group carbonates in the uppermost Ordovician, Silurian, and Devonian; and Woodford Shale in the Upper Devonian and lowermost Mississippian. Unconformities are found on top of the Oil Creek Formation in the

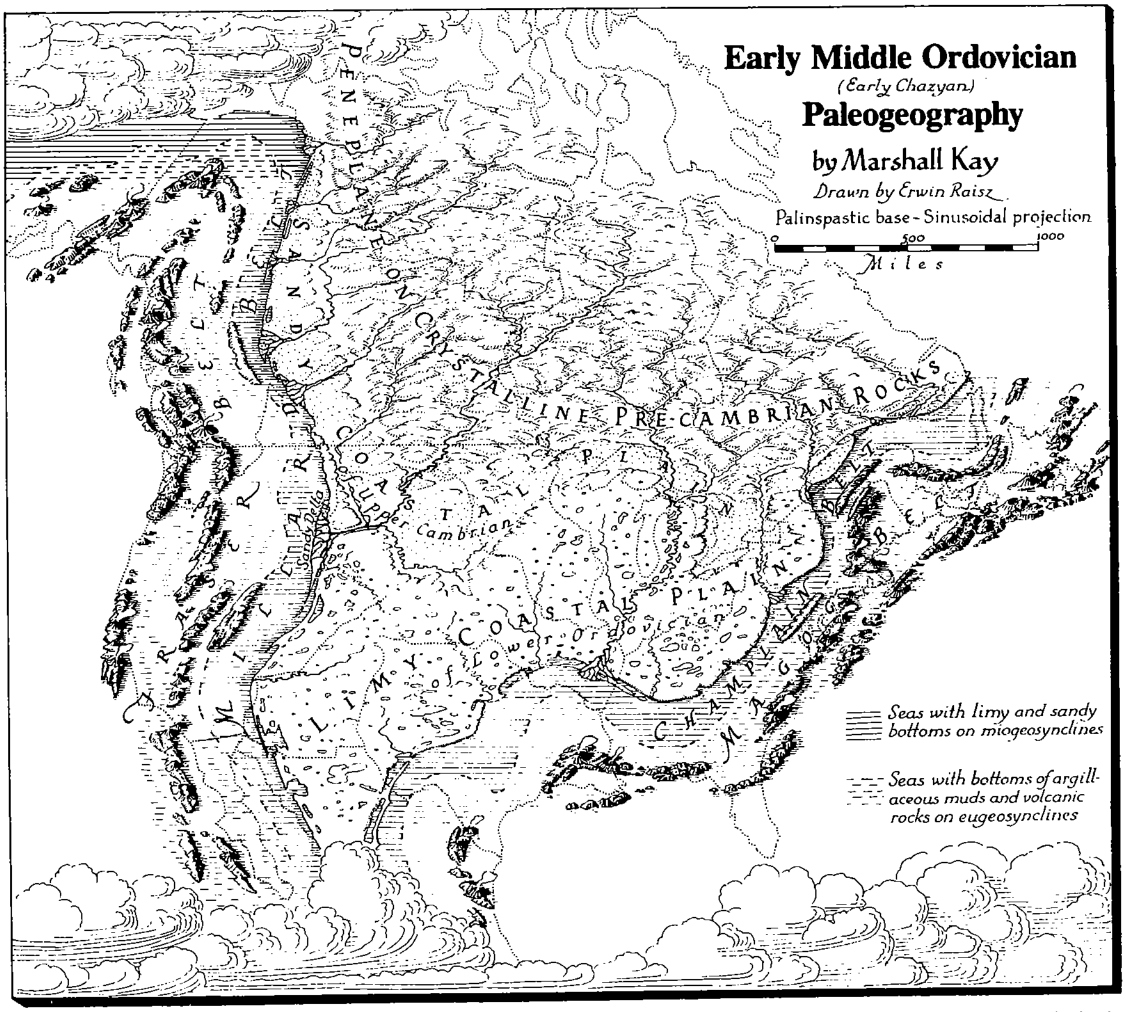

Figure 2. Early Middle Ordovician paleogeography of the North American craton. The limy coastal plain is the "Great American Bank" which developed in the Early Ordovician. Modified after Kay (1951).

lower Simpson Group, on top of the Bromide Formation in the upper Simpson Group, in the middle of the Viola Group; more-pronounced unconformities exist beneath the Frisco–Sallisaw Formations in the Hunton Group in the middle Lower Devonian, and between the Hunton Group and Woodford Shale in the Middle to Upper Devonian (Johnson and others, 1988,1989).

In the Late Mississippian through Late Pennsylvanian, a series of orogenic movements left a regional pre-Pennsylvanian unconformity and initiated development of an elongate basin. Pennsylvanian fold and thrust motion emplaced the Wichita and Arbuckle Mountains (Figs. 4,5). These large structures have been attributed to left-lateral transpression in the Wichita Mountains (Donovan and others, 1989), transpression in the Arbuckle Mountains (Perry, 1989), and compressional motion (Brewer and others, 1983; Brown, 1984) in both areas.

Uplifted mountain blocks, such as the Arbuckles and Wichita Mountains, shed gravels and finer detritus into basins. The gravel units here are named "granite wash," Collings Ranch Conglomerate, Post Oak Conglomerate, Vanoss Conglomerate, and Bostwick Conglomerate. Gravels derived from the Arbuckle Group, found in outcrop in the Arbuckle Mountains, are named the Collings Ranch Conglomerate. Specific facies of the Arbuckle, which will be discussed later, can be recognized in individual clasts.

In the Early Permian, some uplifts remained as localized highs, while basin subsidence continued with filling by carbonate, followed by anhydrite, shale, and salt. During the Permian, subaerially exposed Arbuckle formed collapse breccias and

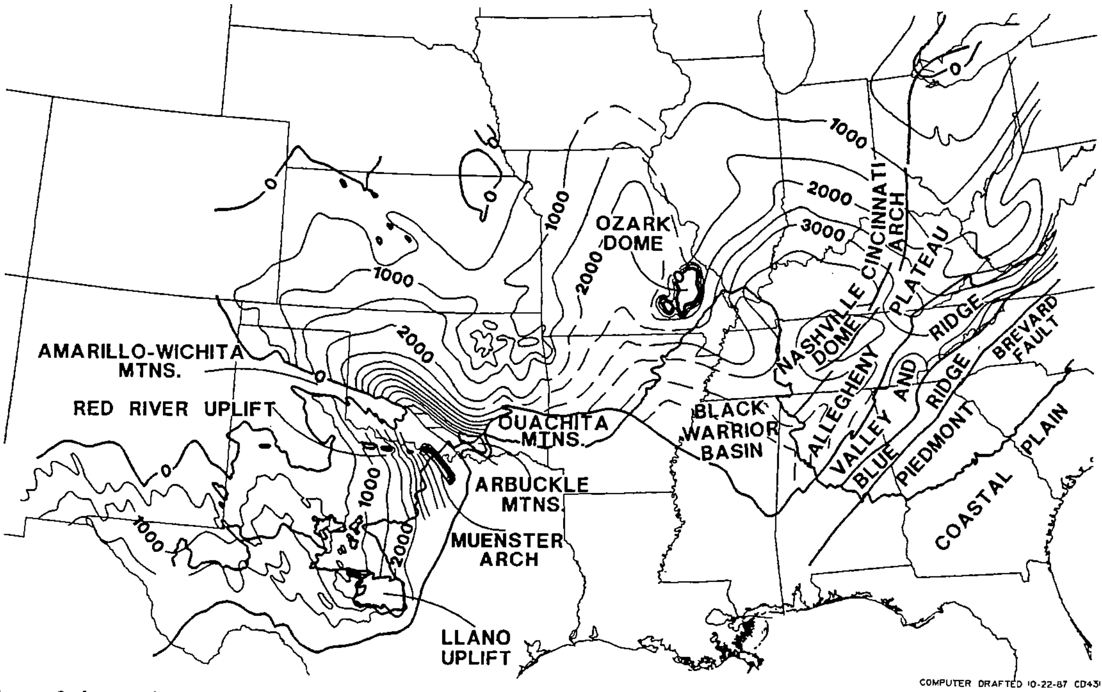

Figure 3. Isopach map of the Arbuckle Group and its lateral equivalents: the Knox Group in the Appalachian Mountains; the Ellenburger Group in the Permian basin of southeast New Mexico and West Texas; and the El Paso Group of West Texas. Thickness in feet.

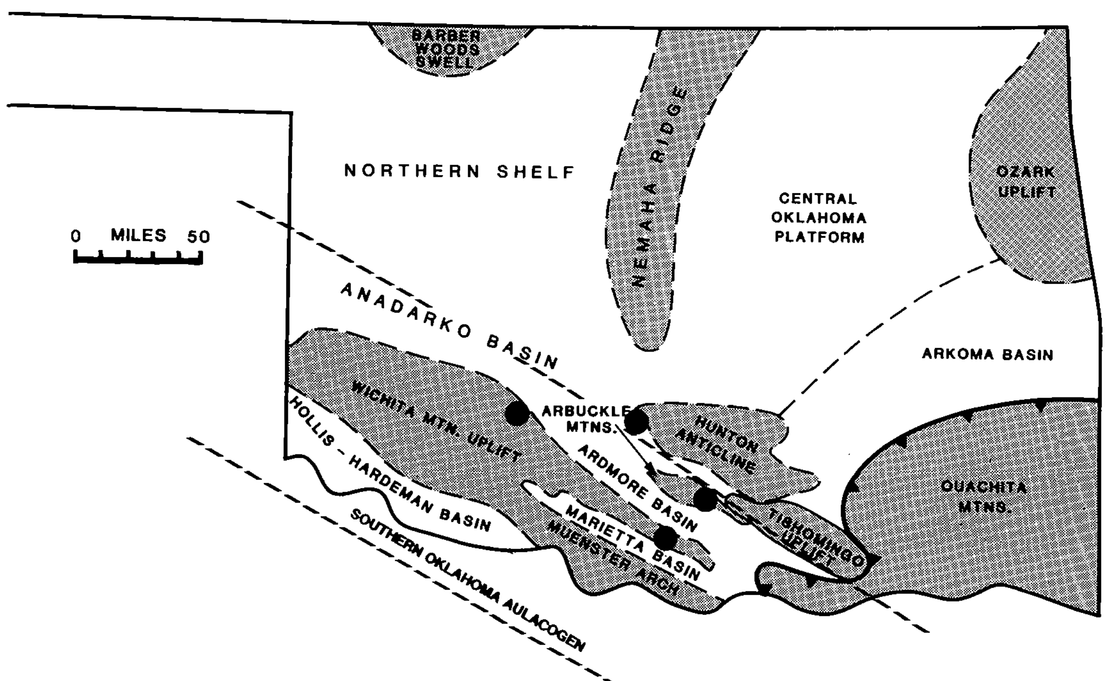

Figure 4. Major uplifts in Oklahoma. Black dots show study areas. Heavy dashed lines are the approximate boundaries of the southern Oklahoma aulacogen.

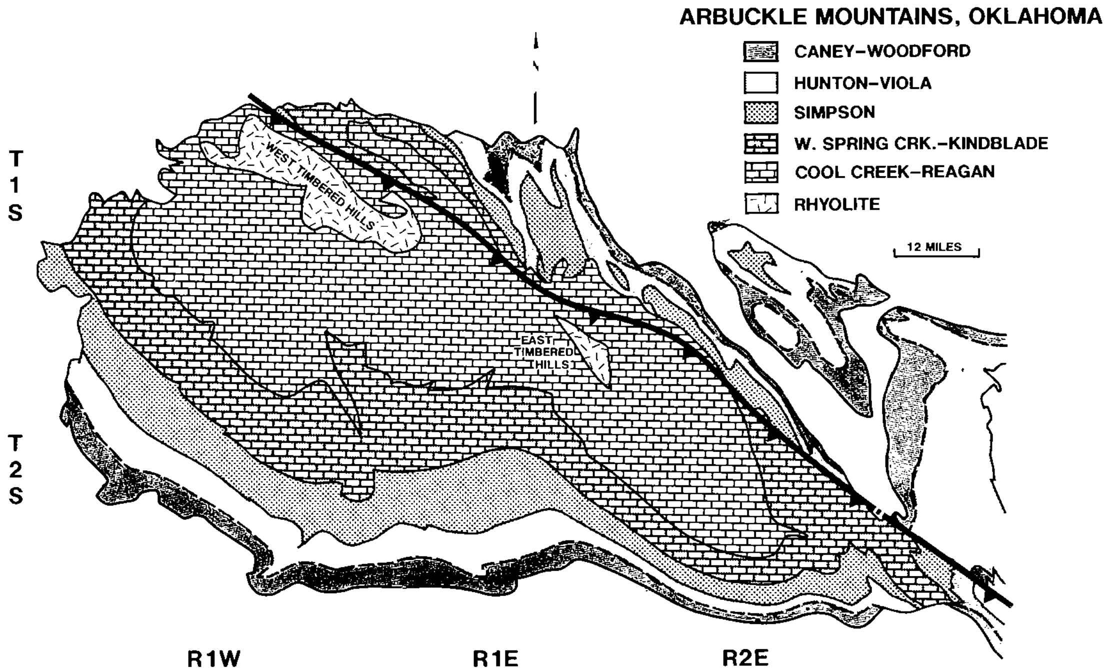

Figure 5. Outcrop map of the Arbuckle Mountains.

solution pipes in the Wichita and Arbuckle Mountains. In the Wichita Mountains, collapse voids were filled with Permian sediments, some fossiliferous (Donovan and others, 1988).

DEPOSITIONAL SETTING

The Arbuckle Group was deposited within three basic environments: (1) a shallow marine setting, (2) a near-shore to shoreline setting, and (3) a tidal-flat setting (Fig. 7). Transition from tidal flats through shoreline and near-shore settings into the shallow-marine was gradual, across a broad, very low-lying carbonate ramp.

Storms were common during deposition of the Arbuckle. A record of these storm events is left in the rock record as tempestite deposits of rip-up intraclasts and flat-pebble to imbricated flat-pebble conglomerate. Storm layers are generally only a few inches thick, and are found throughout subtidal to tidal-flat facies.

Arbuckle shallow-marine deposits form medium to thick beds which are occasionally massive (Fig. 7), while near-shore, shoreline and tidal-flat bedding is much thinner. Away from road cuts on open-range ranches, these subtidal beds form the characteristic "tombstone" topography of the Arbuckle Mountains. The shallow-marine setting is composed of three facies: (1) skeletal-rich packstone, (2) skeletal wackestone and mudstone, and (3) stromatolites and thrombolites. This indicates that within the shallow-marine setting there were either broad facies belts or more-specific niches that each facies occupied. Another possibility is that shallow-marine to near-shore settings may have experienced variations in salinity (schizohaline?).

The nearshore facies (Fig. 7) is composed of thin beds and wavy thin beds. The shoreline facies (Fig. 7) is composed of "flaser-like" ribbon carbonate, which forms very thin beds. Ribbon-carbonate deposition recorded bi-directional tidal exchange as cross-lamination. These two facies are common in some parasequences and are less common or absent in other parasequences. This would indicate that the near-shore to shoreline setting had a variable distribution.

The tidal-flat depositional setting was highly complex (Fig. 7). The outer parts of tidal flats apparently were segmented, or contained large embayments, or were continuous, though probably sinuous. Between and on even the most-continuous tidal flats there were embayments or ponds. These were preserved as very thin, parallel beds, sometimes containing cement-filled molds/vugs where evaporites were leached away, and are referred to as thick laminations. Tidal flats are preserved as laminated to slightly crinkly laminated beds, and are referred to as thin laminations. Upsection, a complete stratigraphic sequence would be thick laminations giving way to thin laminations. Interspersed between thick and thin laminations are occasional beds of ooid grainstone and bioturbated sandstone. Ooid grainstone and sandstone are not common; however, only one facies, either ooid grainstone or bioturbated sandstone, was deposited in an individual parasequence—never both. Ooid grainstone is also found in the shoreline to near-shore setting, which may indicate that large tongues of ooids were being washed back and forth by tidal exchange in embayments or channels out through the shoreline. These may have formed very thin ebb-and-flood tidal deltas that were channelized and oriented perpendicular to the general shoreline trend.

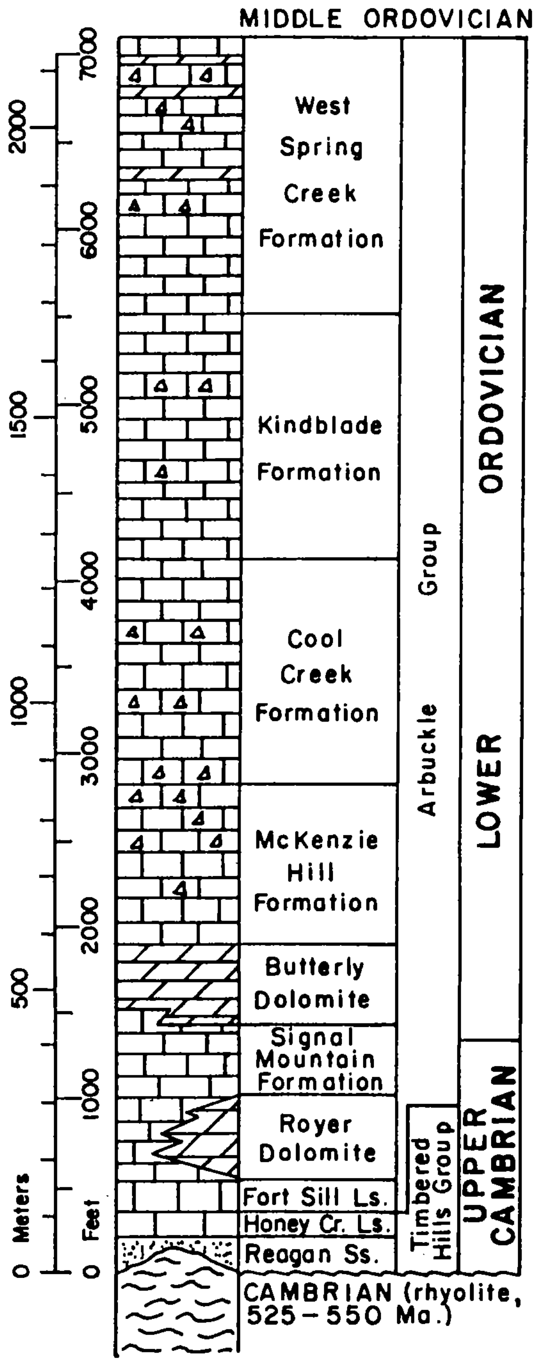

Figure 6. Stratigraphic column above Middle Cambrian Colbert Rhyolite, showing Upper Cambrian Timbered Hills Group and Upper Cambrian–Lower Ordovician Arbuckle Group stratigraphy in the Arbuckle Mountains. From Johnson and others (1988,1989).

Wave and tidal energy affecting the depositional setting may have been dissipated to some extent by the broad, low-lying carbonate ramp. However, there are some high-energy facies, such as ribbon carbonate and ooid grainstone, along the shoreline. Also, shallow-marine stromatolites and thrombolites are sometimes surrounded and capped by intraclasts and ooids. These are reminiscent of present-day stromatolites recently discovered by Dill and Shinn (1986a,b) and Dill and others (1989) in high-energy, ooid-generating tidal passages in the Bahamas. A strong, 150-cm/sec tidal exchange keeps all predators, except swimmers, away. In the Arbuckle Group, some stromatolites/thrombolites are not associated with grain-rich intervals and may have grown on the shallow sea floor; they probably populated the sea floor where predators, if any, were kept away due to increased salinities. This would be similar to the present-day setting in Hamelin Pool, Shark Bay, Western Australia. The Arbuckle shallow-marine setting did not become saline enough to precipitate bedded evaporites, although there is evidence of individual evaporite crystals (St. John and Eby, 1978; Ragland and Donovan, 1984,1985).

DEPOSITIONAL CYCLES

Although the Arbuckle Group is as much as 8,000 ft thick in the southern Oklahoma aulacogen, it has been found that Arbuckle deposition was made up of small-scale, short-lived, fifth-order carbonate cycles. These depositional parasequences, once deposited and physically/chemically compacted, are 1–104 ft thick, averaging 14 ft.

Rates of subsidence fairly well matched rates of carbonate deposition on the carbonate ramp. This allowed transgression and regression of the sea across vast distances on the southern margin of the North American craton. Each cycle is represented by a shallowing-upward parasequence consisting of three components: (1) a basal, shallow-marine unit deposited during transgression across the craton and initial regression off the craton; (2) an intermediate, near-shore to shoreline unit deposited as regression continued and a paleoshoreline prograded seaward; and (3) a tidal-flat unit deposited as the paleo-shoreline and tidal flat prograded across the southern margin of the

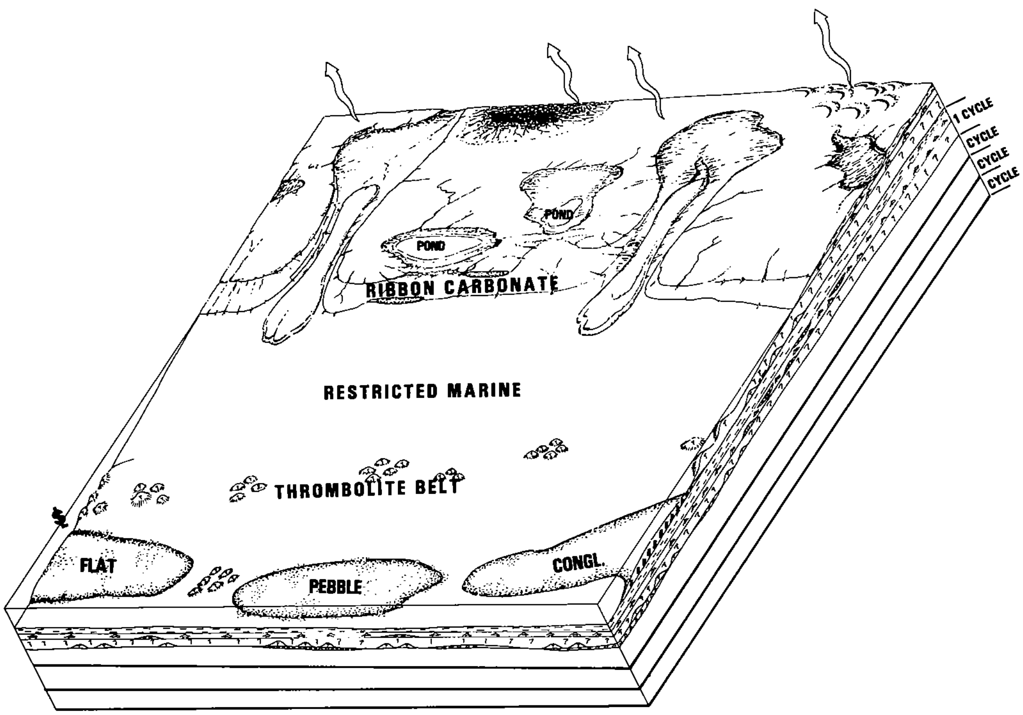

Figure 7. Depositional model of the Arbuckle Group in southern Oklahoma, showing a shallow-marine setting passing landward into a near-shore to shoreline setting, and ultimately into a tidal-flat setting. The sea alternately transgressed and regressed, depositing thin carbonate sequences.

craton. Deposition on the inner, middle, or outer parts of the carbonate ramp may have been variable and was controlled by the distance of individual transgressions, and whether the entire ramp or only parts of it were covered.

Therefore, if rates of subsidence and rates of carbonate deposition were equal, then the resulting depositional parasequence should be similar, or at least fairly similar. This, however, is not the case. Each depositional parasequence is unique in its set of facies variations, in its thickness, and in its completeness (or incompleteness) as a shallowing-upward parasequence. Several factors have affected deposition of each cycle within the Arbuckle Group; some of these factors could have been (1) the magnitude or height of each transgression, (2) the length of time during which each parasequence was deposited, and (3) rates of subsidence. Because of these factors, whether acting in combination or singly, the same depositional site received each cyclic carbonate deposit in a slightly different manner.

Individual carbonate parasequences are composed of either a complete or an incomplete shallowing-upward parasequence. In either case, each parasequence starts with a shallow-marine deposit, which may be skeletal packstone, skeletal wackestone to mudstone, or stromatolitic/thrombolitic. A medium to thick, occasionally massive, shallow-marine bed marks the transgressive phase of each parasequence. As sea level slowly fell and the subtidal carbonate factory began to create and deposit vast amounts of carbonate sediment, a regressive progradational event began and the paleo-shoreline migrated out across newly deposited subtidal beds. The first indication of regression in outcrop or core is a change in bed thickness from medium, thick, or occasionally massive bedding to thin, wavy-thin, or very thin bedding. Thin and wavy-thin beds represent a near-shore setting. Very thin "flaser-like" bedding, called ribbon carbonate, formed along the shoreline where tidal exchange was active. If a complete shallowing-upward parasequence was deposited, the near-shore to shoreline setting will be represented, but will not form a significant part of the parasequence. If an incomplete shallowing-upward parasequence was deposited instead, with no tidal-flat cap, the near-shore to shoreline facies can be as thick or thicker than the shallow-marine facies.

The final phase of development of a shallowing-upward parasequence is deposition of very thin parallel beds, thick laminations deposited in embayments and ponds, and thin laminations deposited on tidal flats, capping a complete shallowing-upward parasequence. These two facies are present in a complete shallowing-upward parasequence, whereas in an incomplete parasequence neither of them or only thick laminations may be present. Many cycles appear to record a deepening event at the top of an individual parasequence. Deepening (drowning) of the shoreline may have been the result of slightly faster rates of basin subsidence or an abruptly increased rate of subsidence. Whichever the case, the resulting cycle of deposition left an incomplete shallowing-upward parasequence. However, the next transgressive event still deposited shallow-marine beds above back-stepping (drowning) deposits, which indicates that the transgression was not related to basin subsidence.

BUTTERLY DOLOMITE

Butterly Dolomite cycles were studied in outcrop in the Arbuckle Mountains. The best outcrop available for examination is along Interstate 35 on the north flank of the Arbuckle Mountains, where the road cut in a hillside exposed the Butterly (E½SW¼NE¼NW¼ sec. 6, T. 2 S., R. 2 E., Murray County). A marker (no. 25) has been placed in the outcrop by the Oklahoma Geological Survey (OGS) and Ardmore Geological Society (AGS), identifying the outcrop as Butterly Dolomite. Here the Butterly is 59 ft thick and is broken up by several faults into small fault blocks (Fig. 8).

At this location the Butterly is composed of eight depositional parasequences (Fig. 8). Individual parasequences are 2.5–12.5 ft thick, averaging 7.3 ft thick. The total thickness of the Butterly is 297 ft (Fay, 1989). If the remaining covered portion contains similar parasequences of similar thickness, then some 40 depositional parasequences form the Butterly. Each of the eight exposed parasequences are complete shallowing-upward parasequences. All parasequences are composed of thick to massive, shallow-marine beds that are capped by tidal-flat thin laminations. No nearshore to shoreline facies (thick laminations) were deposited. Thick to massive, shallow-marine beds that form the greater part of each parasequence are 1.5–11 ft thick, averaging 6.5 ft. Tidal-flat thin laminations are 0.5–2 ft thick, averaging <1 ft. The entire outcrop has been dolomitized. Butterly depositional parasequences are very simple compared to parasequences in the rest of the Arbuckle Group. Walking the tombstone topography in the open range, away from road cuts, reveals that

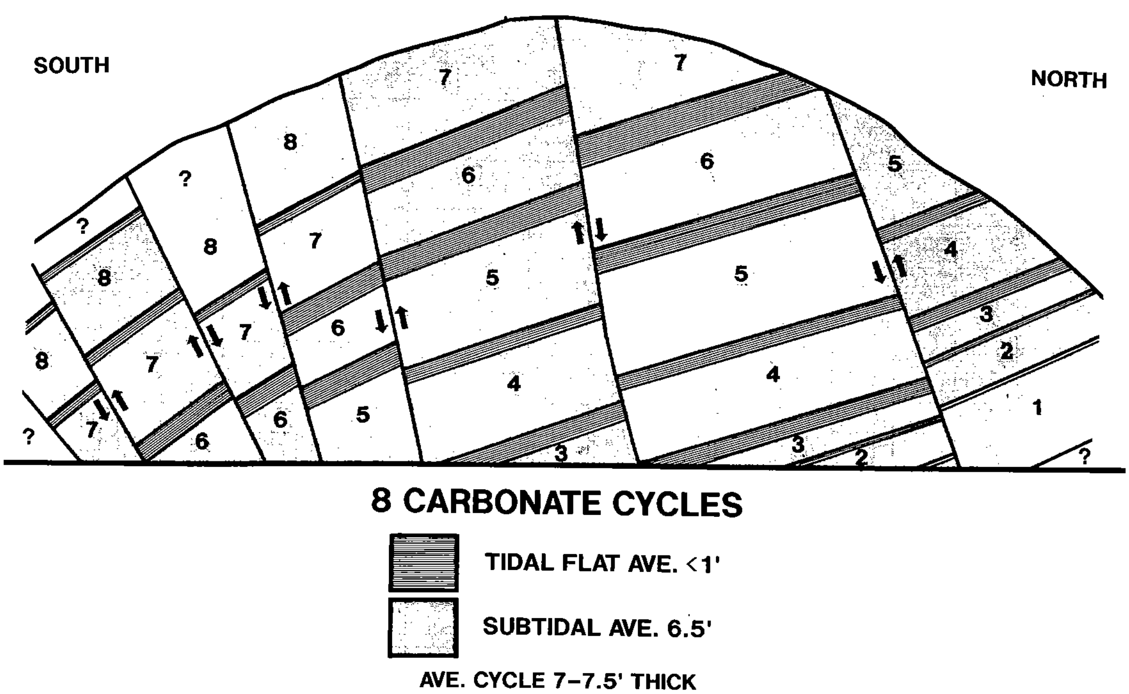

Figure 8. Butterly Dolomite parasequences in outcrop along Interstate 35, north flank of Arbuckle Mountains. Eight carbonate parasequences form this highly faulted outcrop. Each parasequence is so distinctive that it can be correlated from fault block to fault block; this shows relative motion of individual fault blocks.

Butterly parasequences contain much better-developed, more-dominant shallow-marine beds than the underlying Signal Mountain or overlying McKenzie Hill Formations.

COOL CREEK FORMATION

Cool Creek Formation parasequences were studied in outcrop at three localities in the Arbuckle Mountains. The best outcrop available is the road cut along the west side of Interstate 35 just north of the Carter/Murray county line, where a rest stop is located (CNE¼SE¼NE¼SE¼ sec. 13, T. 2 S., R. 1 E., Murray County). This locality is topographically the highest elevation along the highway. Here the outcrop extends from ~612 ft beneath the top of the Cool Creek to a point 1,142 ft below it. A marker (no. 12) has been placed in the outcrop by the OGS and AGS 700 ft below the top of the Cool Creek. Forty-two parasequences were measured at this location. Individual parasequences are 3–80 ft thick, averaging 11.75 ft.

Two other studied areas are inside the boundaries of Turner Falls Park. One outcrop is just beneath the overlook above Turner Falls (CE½E½ sec. 36, T. 1 S., R. 1 E., Murray County), next to the gift shop along Highway 77 (Fig. 9). Here a 42-ft-thick section contains four complete shallowing-upward parasequences and the top of a fifth.

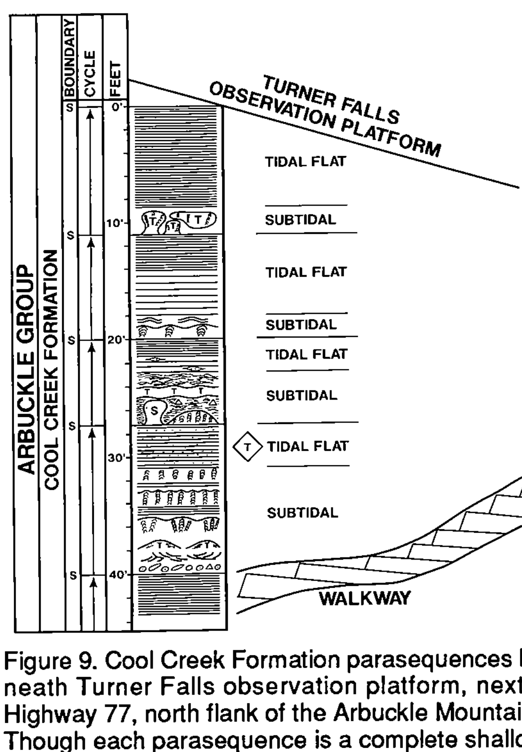

Figure 9. Cool Creek Formation parasequences beneath Turner Falls observation platform, next to Highway 77, north flank of the Arbuckle Mountains. Though each parasequence is a complete shallowing-upward parasequence, no two were deposited in the same manner.

These are directly beneath the observation platform, between the platform and the stone walkway down into the park (this walkway is now fenced off). These parasequences are 7–13 ft thick, averaging 10 ft.

The second locality of the Cool Creek inside Turner Falls Park is along the walkway to Turner Falls (CNE¼NE¼NW¼SE¼ sec. 36, T. 1 S., R. 1 E., Murray County), where a section 43 ft thick is exposed just east of a normal fault where Cool Creek is in fault contact with McKenzie Hill Formation. At this locality the Cool Creek outcrop is stratigraphically lower than the outcrop at Turner Falls overlook. Seven parasequences and the top of an eighth were recognized. Four of these parasequences are complete shallowing-upward parasequences and three are incomplete. These parasequences are 3–10.5 ft thick, averaging 5.1 ft.

Combining data from all three areas, Cool Creek depositional parasequences average 10.75 ft in thickness. Since a complete section of Cool Creek Formation is 1,300 ft thick (Fay, 1989), and assuming that average parasequence thickness is slightly less than 11 ft, the Cool Creek could contain 118 depositional parasequences.

KINDBLADE FORMATION

The Kindblade Formation was studied in outcrop on the south flank of the Arbuckle Mountains along Interstate 35 (CE½SE¼SE¼ sec. 13, T. 2 S., R. 1 E., Murray County, to CE½NE¼ sec. 24, T. 2 S., R. 1 E., Carter County). The Kindblade at this locality is 1,370 ft thick. (Fay, 1989, measured 1,440 ft at the same locality.) At this locality all of the Kindblade except the lower 100 ft is completely exposed along the highway. Fifty-seven sequences and part of another form the Kindblade. The partial sequence straddles the West Spring Creek/Kindblade contact. Depositional parasequences in the Kindblade are 1.5–94 ft thick, averaging 22 ft. Kindblade depositional parasequences tend to form incomplete shallowing-upward parasequences. Of the 57 parasequences that were recognized, only 14 are complete. Complete parasequences are much thinner than incomplete ones.

The Kindblade can be divided into a lower interval and a middle to upper interval. Lower Kindblade parasequences are thinner than middle and upper parasequences. In the lower Kindblade, complete parasequences are 2.5–19 ft thick, averaging 9.75 ft; incomplete parasequences are 3–94 ft thick, averaging 18.5 ft (Fig. 10). Middle and upper Kindblade complete parasequences are 4–41 ft thick, averaging 15.5 ft; incomplete parasequences are 1.5–98.5 ft thick, averaging 25 ft (Fig. 11).

Both complete and incomplete parasequences contain shallow-marine facies. Near-shore and shoreline facies are not common in complete

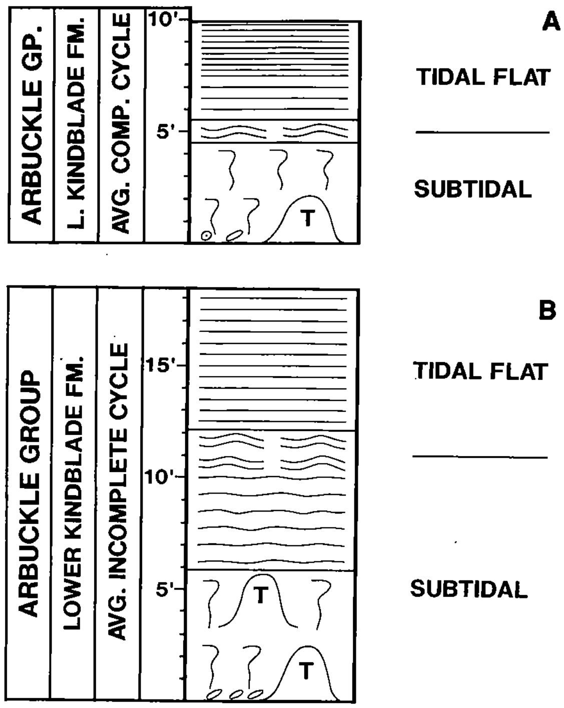

Figure 10. Lower Kindblade Formation parasequences, next to Interstate 35, south flank of the Arbuckle Mountains. Examples of complete (*A*) and incomplete (*B*) shallowing-upward parasequences are illustrated. Notice that lower Kindblade incomplete parasequences are thicker than complete parasequences.

cycles (Figs. 10,11). In incomplete parasequences, the near-shore and shoreline facies are common or are the dominant facies within a parasequence (Figs. 10,11). Tidal-flat facies and embayments and ponds within complete parasequences are well developed, whereas incomplete parasequences contain only embayment or pond thick laminations.

Variations in the development of complete versus incomplete shallowing-upward parasequences were constructed facies by facies. In some cases sea-level rise and fall and rates of subsidences were nearly balanced, producing a complete shallowing-upward parasequence. More commonly during Kindblade deposition sea-level rise and fall and subsidence were not balanced: Subsidence appears to have been faster. This tended to stack up a much thicker near-shore to shoreline set of facies and not allow tidal-flat facies to prograde across the craton, especially during deposition of middle to upper Kindblade parasequences.

WEST SPRING CREEK FORMATION

The West Spring Creek Formation crops out on the south flank of the Arbuckle Mountains in a long road cut along Interstate 35 (CE½NE¼ sec. 24, T. 2 S., R. 1 E., to SW¼NE¼SE¼ sec. 24, T. 2 S., R. 1 E., Carter County). In the subsurface, cores of the West Spring Creek were studied from Healdton field and from a deep well, the Sun/Parker Mazur No. 1 (CSW¼NE¼ sec. 1, T. 3 N., R. 5 W., Grady County). Outcrops in the Wichita Mountains (Bally Mountain) were also field-checked.

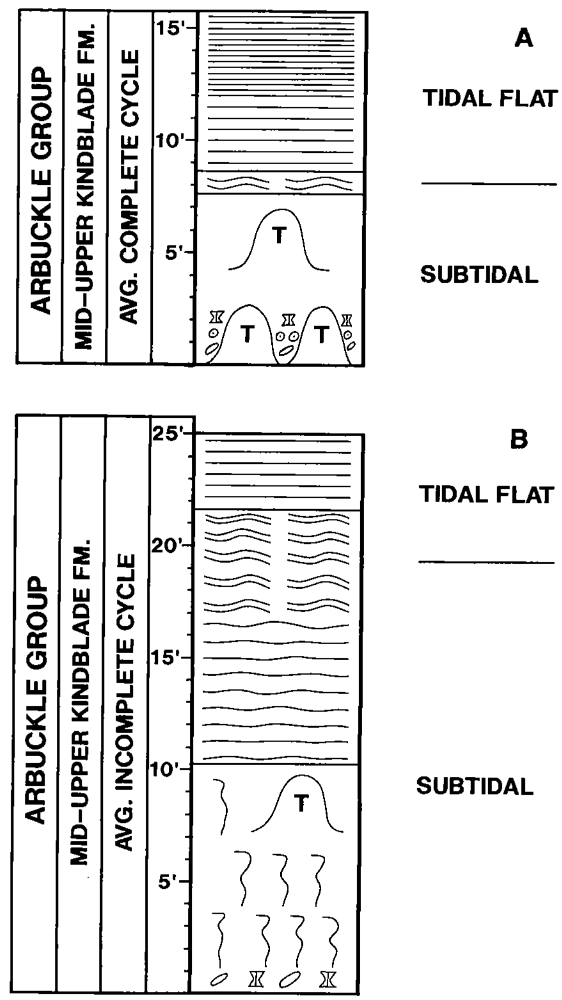

Figure 11. Mid-upper Kindblade Formation parasequences along Interstate 35, south flank of the Arbuckle Mountains. Examples of complete (*A*) and incomplete (*B*) shallowing-upward parasequences are illustrated. The Kindblade is composed of 57 parasequences, plus half of a parasequence that straddles the West Spring Creek/Kindblade boundary.

In the Arbuckle Mountains our measured section of the West Spring Creek is 1,520 ft thick, which is practically identical to the 1,528 ft measured by Fay (1989). The uppermost 67 ft of West Spring Creek is covered. The section is composed of 123 shallowing-upward parasequences and part of another, 69 of which are complete parasequences and 53 incomplete. In the Kindblade there were more incomplete shallowing-upward parasequences (75%), whereas in the West Spring Creek less than half (41%) are incomplete parasequences. These differences illustrate how the Kindblade was dominated by transgressive events, while the West Spring Creek was dominated by regressive events. Depositional parasequences in the West Spring Creek are <1 ft to 45 ft thick, averaging 12 ft. Seventy-three complete shallowing-upward parasequences were deposited and are 1.5–45 ft thick, averaging 12 ft; there are 50 incomplete parasequences 0.75–47 ft thick, averaging 11 ft. The West Spring Creek experienced a general regression, which left slightly thicker complete shallowing-upward parasequences and slightly thinner incomplete parasequences. Also, the lower part of the West Spring Creek is dominated by complete parasequences.

The upper 873 ft of the West Spring Creek in core from the Sun/Parker Mazur No. 1 well, 40 mi northwest of the Interstate 35 locality, contains 41 depositional parasequences. These parasequences are 2–104 ft thick, averaging 21 ft. Twenty-one cycles are complete shallowing-upward parasequences, and 20 are incomplete. These parasequences, on the average, are nearly twice as thick as West Spring Creek parasequences in the Arbuckle Mountains. Mazur No. 1 complete parasequences average 23 ft in thickness, thicker than incomplete parasequences, which average 18.75 ft. It also appears that more, but thinner, West Spring Creek parasequences were deposited in the Arbuckle Mountains and that fewer, but thicker, West Spring Creek parasequences were deposited in the Mazur No. 1. This is of great importance, since the Arbuckle Mountains are within the southern Oklahoma aulacogen, while the Mazur No. 1 well is north of the north bounding fault and outside the aulacogen.

Comparison of facies within and immediately outside the aulacogen shows that the lower West Spring Creek is dominated by shallow-marine sediments in the Arbuckle Mountains, and by tidal-flat deposits in the Sun/Parker Mazur No. 1 (Fig. 12). Also, ooid grainstone is more common in the Mazur No. 1 than in the Arbuckle Mountains. The upper West Spring Creek is dominated by tidal-flat deposits in both areas (Fig. 13). Near-shore to shoreline facies are better developed in the Mazur No. 1 than in the Arbuckle Mountains. Also, thicker, bioturbated siliciclastic beds are more common in the upper West Spring Creek in the Mazur No. 1 than in the Arbuckles.

CAUSE OF CYCLICITY

We estimate a total of ~400 parasequences were deposited in the Lower Ordovician part of the Arbuckle Group. That interval represents 15.5 m.y. of time, according to Ross (1976). Comparing the approximate total number of parasequences to time, without accounting for any hiatuses, reveals that an average parasequence may have been deposited in ~39,000 yr. This time per parasequence

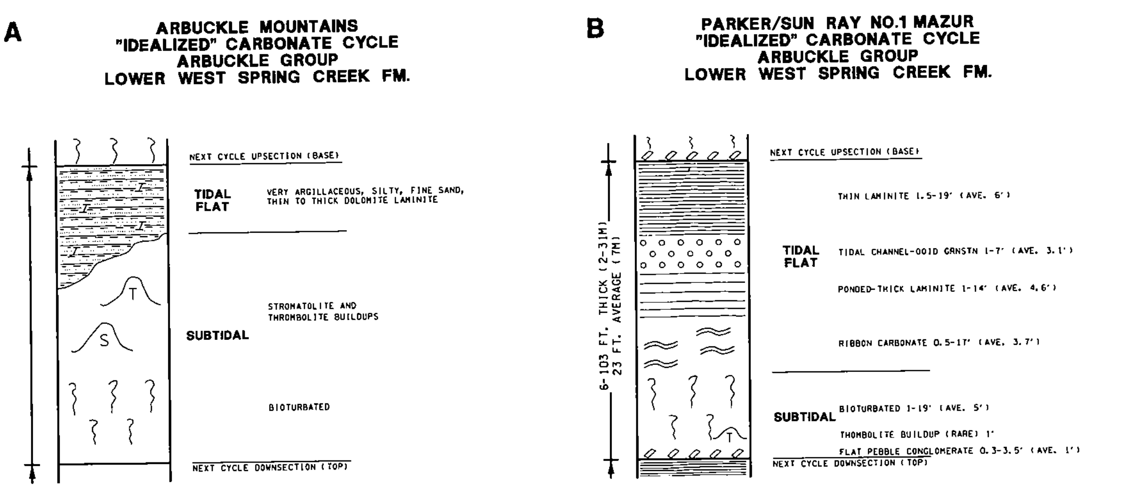

Figure 12. Average lower West Spring Creek parasequence in the Arbuckle Mountains (*A*) compared to an average parasequence in the Sun/Parker Mazur No. 1 well (*B*) in the deep subsurface 40 mi to the northwest, outside the southern Oklahoma aulacogen.

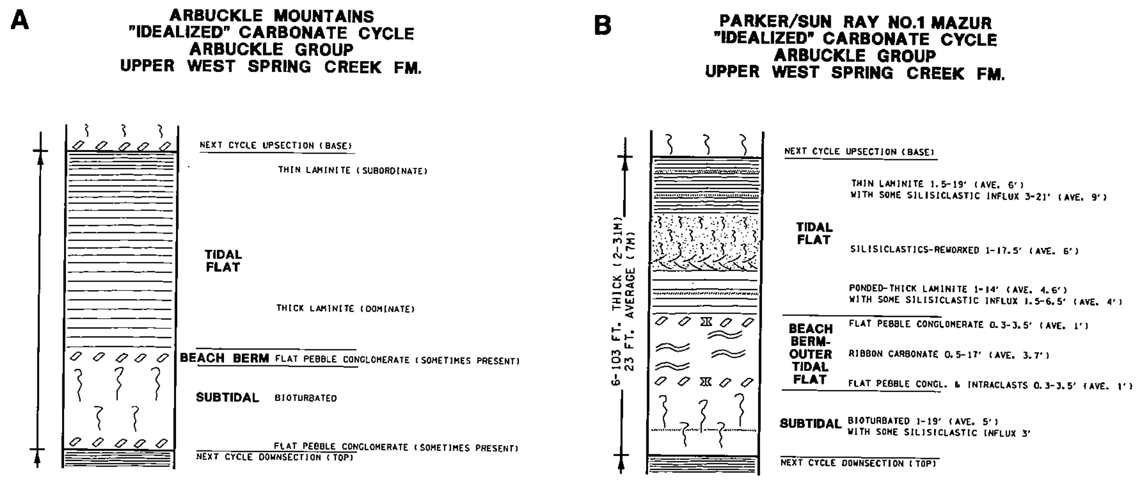

Figure 13. Average upper West Spring Creek parasequence in the Arbuckle Mountains (*A*) compared to an average parasequence in the Sun/Parker Mazur No. 1 well (*B*) in the deep subsurface 40 mi to the northwest, outside the southern Oklahoma aulacogen.

closely approximates perturbations in Earth's change in obliquity (tilt), which have a period of ~40,000 yr. A change in Earth's tilt of 3.5° could have had a dramatic effect upon climate and the length of individual seasons, creating either an overall warming or cooling trend. It seems possible to tie Arbuckle Group cyclic deposition to glacioeustatic rise and fall of sea level. Sea-level variations could have been caused by perturbations in Earth's spin, tilt, and orbit around the Sun, discovered by Milankovitch (1941), which are (1) 21,000 yr precession, or wobble of the Earth; (2) 40,000 yr changes in Earth's obliquity, or tilt; and (3) 100,000 yr eccentricity, or variations in Earth's orbit due to gravitational effects of other planets (see review by Fischer and others, 1988).

It is very interesting that, on an average, one of these three perturbations would fit the data from the Arbuckle Group so well. However, an unknown factor, which may have also been active, was the rate of sea-floor spreading during the Early Ordovician. This is mentioned because variations in rates of sea-floor spreading, within short periods of time, and the associated swelling of spreading centers, could cause sea-level rise, and perhaps fall, if rates were variable. How significant this may have been and what the time frame may have been, are not known (see review by Sloss, 1988). It also appears that rates of subsidence within the southern Oklahoma aulacogen were not uniform, but variable and perhaps spasmodic.

CONCLUSIONS

Arbuckle Group carbonates were deposited on a broad, flat-lying carbonate ramp. This carbonate ramp formed the southern margin of the North American craton. Deposition of carbonate sediment was in three basic environments: (1) shallow marine, (2) near-shore to shoreline, and (3) tidal-flat. The depositional environment was periodically ravaged by storms that left tempestites. Specific facies associated with each depositional environment are (1) shallow-marine skeletal packstone, skeletal wackestone and mudstone, or stromatolite/thrombolite facies; (2) near-shore thin to wavy beds and shoreline "flaser-like" ribbon carbonate; and (3) parallel to crinkly tidal-flat laminations (thin laminations), and very thin beds of coastal-flat embayments or ponds (thick laminations), sometimes containing evaporite molds and vugs. Ooid grainstone or bioturbated sandstone are also present; but they are not common, and were not deposited together.

The depositional setting was punctuated by small-scale, short-lived, fifth-order parasequences. Individual parasequences are 1–104 ft thick, averaging 14 ft. Complete shallowing-upward parasequences and incomplete parasequences both have been recognized. When rates of subsidence matched rates of accumulation of carbonate sediment, then rises and falls of sea level could cause transgression and regression across the broad carbonate ramp, depositing a complete shallowing-upward parasequence. When rates of subsidence increased, incomplete parasequences resulted. An individual parasequence can be divided into three parts: (1) a base composed of thick to massive, shallow-marine beds, deposited during transgression across the craton and initial regression; (2) an intermediate part composed of thin to wavy-thin near-shore beds and very thin "flaser-like" ribbon carbonate shoreline beds, deposited as regression continued and a paleo-shoreline prograded seaward; and (3) an upper part composed of tidal-flat laminations and crinkly laminations and very thin beds deposited in embayments or ponds as the paleo-shoreline and tidal-flat setting prograded across the margin of the craton.

Four hundred Arbuckle Group depositional parasequences may have been deposited in the Early Ordovician. Individual parasequences were deposited, on the average, every 39,000 yr; this period is similar to the period of change in Earth's obliquity, which averages 40,000 yr. Changes in Earth's tilt would cause climatic variations, with cooling and warming trends. Cooling trends would tie up ice on continents (Africa) near the South Pole during the Late Cambrian–Early Ordovician, causing a regression off the craton as sea level fell. Warming trends would melt ice and cause a transgression that would flood the craton. Rates of subsidence may not have been uniform through time, but may have been variable.

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ON THREE OUTCROPS OF THE SIMPSON GROUP IN THE NORTHWESTERN PART OF THE SLICK HILLS, SOUTHERN OKLAHOMA

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ABSTRACT.—Three exposures in the northwestern part of the Slick Hills (the "Sugar Hills") display the upper part (Bromide Formation) of the Simpson Group. The sequences are conformably overlain by limestones of the Viola Group. Bromide rocks consist of an upper sequence of platform carbonates (mudstones, wackestones, and packstones) (Pooleville member) and a lower sequence of quartz-rich sandstones, olive-green shales and thin limestones (including grainstones) (Mountain Lake member). The sequence is interpreted as the record of a long-term transgression following a low stand at the commencement of Bromide deposition. Details in the sequence suggest that the transgression was punctuated by short-term variations in sea level. Sands were moved by tidal currents which flowed northwest–southeast and constructed laterally variable beds as much as 10 ft thick.

Early diagenesis involved cementation by increasingly ferroan calcite and syntaxial quartz precipitation. Burial diagenesis included widespread replacement by ferroan dolomite and ankerite as well as continuing precipitation of silica. The latter did not completely obliterate porosity in the quartz-rich sandstones, which have primary porosities as much as 20%. In Permian times, the Sugar Hills were exposed at the surface following Pennsylvanian tectonism. The chief results of this uplift were widespread oxidation and dedolomitization of ferroan carbonates and the upward migration of hydrocarbons through the section from the underlying Anadarko basin.

INTRODUCTION

An Overview of the Simpson

The Simpson Group is one of the great stratigraphic units recognized in the lower Paleozoic of the southern Midcontinent. The oldest rocks in the group are dated as lower, but not basal, Whiterockian (basal Middle Ordovician), while the youngest are lower Mohawkian (Johnson and others, 1988). The group disconformably overlies the West Spring Creek Formation of the Arbuckle Group and is disconformably overlain by the Viola Group.

Internally the Simpson has been divided on lithostratigraphic grounds into five formations. From base to top these are the Joins, Oil Creek, McLish, Tulip Creek, and Bromide. Further, less formal, subdivision of these beds is a recognition of the lithologic complexity of the group, which is built of a variety of shales, quartz-rich sandstones, and thin limestones. A thick sandstone is taken as the base of each of the four upper formations (Decker and Merritt, 1931; Longman, 1981).

In general the Simpson is found throughout most of northern Oklahoma, the northern reaches of the Texas Panhandle, and in a 100-mi-wide belt trending E–W through northern Arkansas (Johnson and others, 1988). There is little doubt that the group formerly occupied a larger area but has subsequently been removed by Devonian and later Paleozoic erosion, following various epeirogenic uplifts. Thickness isopachs of the group closely follow the contours of the southern Oklahoma aulacogen (Johnson and others, 1988). The maximum thickness of the group is ~2,300 ft in the Ardmore area, close to the center of the aulacogen (Decker and Merritt, 1931), whereas thicknesses of ~330 ft are more typical of cratonic areas distant from the aulacogen axis.

The lithologic variety of the Simpson Group, which is in such marked contrast to the lithologic monotony of the underlying Arbuckle Group, has attracted a considerable amount of attention, especially because the various quartz-rich sandstones can be such prolific hydrocarbon reservoirs. Various authors have addressed questions, such as:

What is the reason for the extraordinary mineralogical maturity of the quartz-rich sandstones?

All the quartz-rich sandstones in the Simpson are remarkable for their mineralogical maturity—generally ≥95% of grains are quartz (e.g., Houseknecht, 1987,1988; McPherson and others, 1988; Pitman and Burruss, 1989; Pollastro, 1989). Most of this quartz is monocrystalline; presumably it is the end product either of multicycling of now-vanished formations (Dapples, 1955; Munsil, 1983), or of intense and/or long-lasting weathering during a single erosional-depositional cycle (or of some combination of the above). Given the established lower Paleozoic paleogeography of the area (Johnson and others, 1988), the most likely area of origin for the sediments lay to the north, where various granitic terranes were a probable source. There is little evidence for tectonic activity and associated uplift on the "proto-North American craton" in this area at this time, suggesting that the Simpson quartz sands, like those in the underlying Arbuckle Group, are the ultimate product of a long-lasting and complete weathering cycle (Donovan, 1986).

In this context, one of the most illuminating relationships is the unconformity between the Harding Formation (upper Simpson Group equivalent) and the underlying Precambrian basement in central Colorado. Although the Precambrian gneisses and granites are a potentially rich source of feldspar, the overlying Harding is an orthoquartzite which is virtually feldspar free; the only unequivocal contribution from the immediate basement consists of a few pebbles and granules of vein quartz. The absence of detrital feldspars suggests an extraordinary completeness of weathering, with the result that the only detrital fragments remaining were quartz and clays (mostly illite).

What is the mechanism by which large amounts of siliciclastic detritus were periodically transported across a craton that was blanketed by platform carbonates?

Most, if not all, authors who have analysed the Simpson suggest that the environment of deposition of the siliciclastics was a shallow-marine setting. This attribution is easiest to defend in the case of the carbonate sequences in the group, wherein an extraordinarily rich shelly fauna provides a strong argument for the existence of carbonate platforms (e.g., Decker and Merritt, 1931; Fay and Graffham, 1982; Longman, 1981,1982). Environmental definition has been somewhat more problematical in the case of the siliciclastics. Basing his arguments on established sedimentary parameters, Longman (1981) has recognized shoreface, foreshore, tidal flat, and "a host of other less easily recognized environments associated with barrier islands and very shallow water terrigenous shelf environments" in the Bromide Formation. In an analysis of a single quarry within the Oil Creek Formation, McPherson and others (1988) have suggested that most of the sequence at this location is built of storm-generated deposits. On the other hand, stratigraphically equivalent sandstones exposed a few miles to the west, in the Buckhorn asphalt district, appear to have been deposited as tidally influenced sand bodies. In essence it would appear that siliciclastics were worked into a variety of shallow-marine settings, differentiated by slight variations in bathymetry.

While the final site of sedimentation of the siliciclastics is reasonably well established as being marine, the initial transport mechanism of the grains appears to have been eolian. The chief evidence which supports this contention is the texture of sand grains in the Oil Creek Formation in the quarry noted above (McPherson and others, 1988). This quarry is remarkable for the preservation of primary porosity in the sandstones; primary surface textures are frosted (Denison and Ham, 1973) and grains are very well rounded. In essence, the Simpson sands are generically similar to other orthoquartzites of Cambrian age in both North America and Europe in that their *fons et origo* is due to eolian sorting on coastal plains which were in essence biological deserts (Longman, 1981; Dalrymple and others, 1985). By extension of this argument, many particles of both quartz and clay in the interbedded shales were probably transported in dust storms.

Assuming that the argument in the preceding paragraph holds true, then it follows that the introduction of siliciclastics into the Oklahoma area was associated with periods of relatively low sea level. During subsequent transgressions the windblown detritus was reworked into various shallow-marine settings. Longman (1981) has suggested that, at least in part, such sea-level changes reflect local variations in subsidence rates, whereas McPherson and others (1988) accent the importance of global fluctuations in the position of sea level. Neither argument is exclusive of the other. Furthermore, the paleogeographic analysis of Schramm (1964) suggests that the various formation boundaries in the Simpson are regional unconformities. If this is true, then it is probable that quartz in the upper formations of the Group is at least partially recycled from lower levels.

Why do the sandstones locally maintain such extraordinarily high primary porosities?

One of the remarkable features of the various Simpson sandstones concerns the spectacular variations in porosity that they show. On the one hand they may be tightly cemented by quartz and carbonate, with vanishingly small porosity (Munsil, 1983). At the other extreme, in the celebrated case of the Oil Creek Sandstone in the silica-sand pit near Mill Creek, Oklahoma, they may be virtually uncemented with porosities of 22.5–30.5% and permeabilities of 2,000–2,500 md (McPherson and others, 1988). In the latter instance the porosity appears to be primary, with slight reduction due to mechanical compaction, pods of poikilitic calcite, and infiltrated detrital illite.

The preservation of such extraordinary porosity must, in the first instance, reflect the near absence of hydrologic gradient throughout the history of the rock (or, if substantial pore-fluid movement took place, that such fluids were undersaturated). Assuming that deposition of the Oil Creek was followed by a transgression, as argued above, stagnant pore waters might be expected to have prevailed during the early history of the rock. Subsequently, during the lower Paleozoic, integrity of porosity may well have been maintained because the deposit was essentially sandwiched between comprehensively cemented carbonate units. In addition, both Munsil (1983) and McPherson and others (1988) have highlighted the role of illite-grain coats in inhibiting syntaxial quartz precipitation.

Following Pennsylvanian tectonism, the Oil Creek Formation in the Mill Creek area was uplifted, thus escaping the long-term deeper burial which affected Simpson sandstones in the Ardmore and Anadarko basins. As Houseknecht (1987,1988) has shown in a study of Bromide sandstones presently buried at depths of 6,500–9,000 ft in the Norman, Oklahoma area, such burial was a potent cause of porosity loss through pressure solution of quartz grains, to the extent that the deposits which he studied were net exporters of quartz.

THE SIMPSON GROUP IN THE SLICK HILLS

The Slick Hills form the exposed portion of the frontal fault zone which lies between the Wichita uplift and the Anadarko basin. They are essentially Permian relief features which have recently been exhumed from beneath a cover of Lower Permian sediments (Donovan, 1986). At the northwestern edge of the the Slick Hills, four Permian-age inselbergs—the Sugar Hills—are the exposed tips of a structural block which dips to the north at about 20°. These hills are capped by carbonates of the Viola Group (Fig. 1); three of the hills are underlain by scanty outcrops of the upper part of the Simpson Group on their southern flanks. Two of these Simpson exposures were described by Decker and Merritt in 1931. Since that time, the most westerly of these exposures has deteriorated significantly (and is not redescribed here), while the other has been improved by quarrying. A third exposure has recently been opened up during road making and is noted here for the first time.

In this paper we describe the two Simpson Group exposures (Fig. 1), the significance of which lies in their structural position (emplaced, by thrusting, above the Anadarko basin) and in their geographic isolation relative both to the more substantial outcrops of the Arbuckle Mountains and Criner Hills and to most of the sections penetrated during drilling operations.

STRATIGRAPHY OF THE SUGAR HILLS EXPOSURES

Neither of the exposures shown in Figure 1 is complete; 25 ft and 110 ft respectively (with 35% of the latter covered). Quarrying in the latter (western) exposure has exposed a conformable contact with the Viola Group. This contact is not marked by a conglomerate at the base of the Viola, as indicated by Decker and Merritt (1931), but is sharp and shows neither relief nor evidence of substantial mechanical erosion. The contact is covered by Permian talus deposits in the (new) road cut exposure.

In both exposures the general stratigraphy shows a sequence of fossiliferous limestones (Table 1) underlain by a mixed sequence of sandstones, green waxy shales, and thinly bedded limestones. It is possible to correlate the two sections

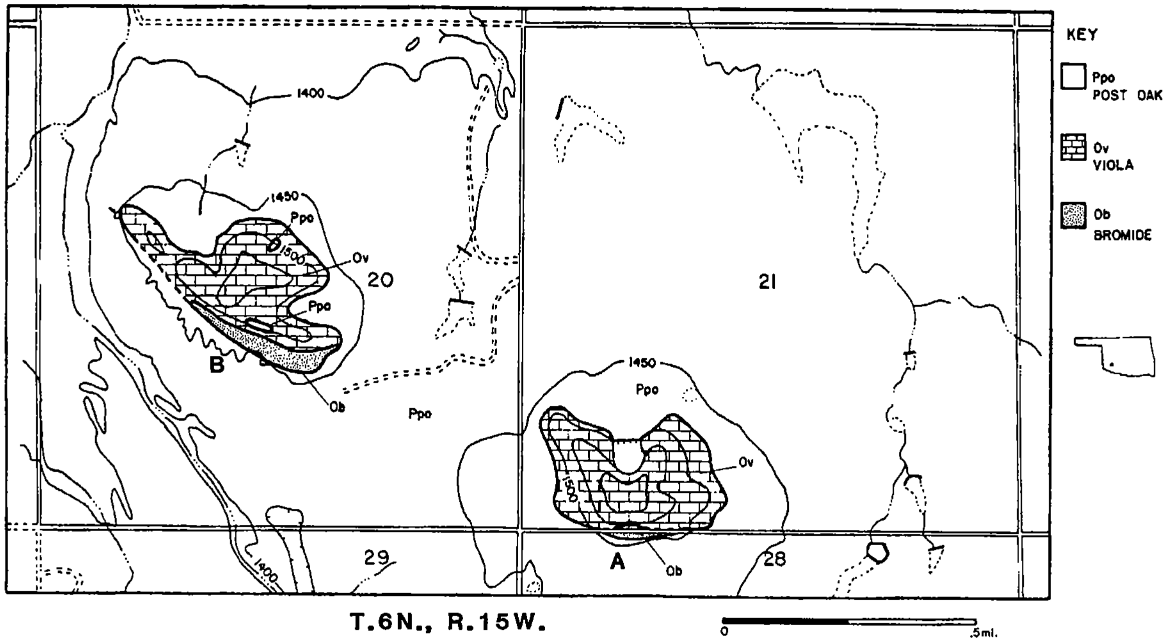

Figure 1. Map showing location of exposures A and B of the Simpson Group (Bromide Formation) in the two Sugar Hills outcrops in Kiowa County, southwestern Oklahoma.

and establish that the siliciclastics show some facies variation (Fig. 2).

Stratigraphic Position of the Exposures

The conformable relationship with the Viola and the fauna of the limestones indicates that the exposures can be assigned to the Bromide Formation (Decker and Merritt, 1931). In their general features, the two exposures are similar to descriptions of the Bromide elsewhere in southern Oklahoma, where Cooper (1956) recognized a lower Mountain Lake Member (shales, sandstones, and thin limestones) and an upper Pooleville Member (limestone). Longman (1981) suggested that the contact between the two members be taken as the junction between fossiliferous micrite (Pooleville) and "any and all shale and biosparite beds." Using this criterion the top 30 ft of the Sugar Hills succession is assigned to the Pooleville. This is much less than the thicknesses recorded by Longman (1981) in the Arbuckle Mountain area; it is uncertain how much of the member may have been removed prior to Viola deposition (either by solution or mechanical erosion). Detailed facies correlation between the Wichita and Arbuckle Mountain areas is not possible (see below).

Lithofacies Description

The most comprehensive description of the Bromide Formation is that of Longman (1981) who described a "more or less predictable vertical sequence" in the Arbuckle Mountain area. In the Mountain Lake Member this consists of six lithofacies: basal sandstone, transitional sandstone, cystoid shale, inner shelf, basinal, and shelf-edge biosparite. In the Pooleville Member he recognized five lithofacies: basinal limestone, diverse fauna biomicrite, limited fauna biomicrite, burrowed micrite, and birds-eye micrite.

In general, similar facies can be recognized in the Mountain Lake portion of the Sugar Hills section, even though the exposure is not complete (Fig. 2). The principal difference is the existence of a thick quartz sandstone ~3 ft below the top of the member. This sandstone is present in both the Sugar Hills sections examined, but is better developed in the eastern (road cut) exposure. The Pooleville in the Sugar Hills is not only much thinner but also is less variable than the section described by Longman.

The lowest beds exposed in the Sugar Hills comprise ~8 ft of fine-grained, tan-weathering, quartz-rich sandstone. Near the base this sandstone displays trough cross bedding in sets as much as 12 in. thick, with reactivation surfaces, and bipolar vectors oriented 115–305°. Other sedimentary structures include horizontal lamination and a single surface of symmetrical ripple marks near the top of the unit.

Separated from the lower sandstone unit by a slight gap is a vivid orange/red-weathering, fine-grained sandstone. This rock consists of 30–70%

TABLE 1. — TAXA IDENTIFIED FROM SLABS AND THIN SECTIONS*

Cnidaria
- *Tetradium ?cellulosum* Bassler, 1950

Bryozoa
- *Stictopora (Rhinidicyta) fenestrara* Ulrich, 1836
- *Hallopora ?macrostoma* Loeblich, 1942
- *Monotrypa ?undulata* Nicholson, 1879
- *Prasopora fritzae* Loeblich, 1942
- *Hemiphragma irrasum* (Ulrich, 1893)
- *Trepostomina crassa* Bassler, 1952
- *?Monticuliporella croneisi* Loeblich, 1942
- *Anolotichia spinulifera* Loeblich, 1942

Brachiopoda
- *Glyptorthis costellata* Cooper, 1932
- *Schizambon* sp.
- *Sowerbyella variabilis* Cooper, 1956
- *Öpikina formosa* Cooper, 1956
- *Multicostella convexa* Cooper, 1956
- *Multicostella magna* Schuchert and Cooper, 1932
- *Hesperorthis sulcata* Cooper, 1956
- *Linguella glypta* Cooper, 1956
- *?Cyclospira parva* Cooper, 1956
- *Dinorthis subquadrata* Cooper, 1936
- *Oxoplecia gouldi* Ulrich and Cooper, 1936

Mollusca
- *Orthoceras* sp.

Arthropoda
- *Lochodomas mcgeheei* Decker, 1931
- *Encrinurus punctatus* Emmrich, 1849

Echinodermata
- *Apodasmocrinus daubei* Warn and Strimple, 1977
- cf. *Doliocrinus pustulatus* Warn, 1982

Dasyclad algae
- *Ischadites iowensis*

*Mostly from the Pooleville Member.

quartz with subordinate amounts of green clay clasts, oolitic phosphatic grains, abraded carbonate grains (including bryozoans, brachiopods, and trilobites), highly weathered glauconite peloids, and variable amounts of greenish clay matrix (much of which appears to have formed by the squashing of clay clasts). The dominant cement is ferroan carbonate, mostly dolomite, which has been dedolomitized and oxidized (hence the overall color of the bed). The bed displays cross bedding, horizontal lamination, and horizontal trace-fossil burrows. A single phosphatic hardground occurs near the top of the unit.

Above another small gap lies a complex 16-in. limestone built of layers of biomicrite separated by hardgrounds. Fossils are mostly broken and include brachiopods, bryozoans, ostracods, echinoderms, molluscs, calcispheres(?), and assorted spines; as much as 20% of the rock consists of very fine-grained quartz grains. The matrix is nonferroan micrite. One layer exhibits burrows that have been infilled by microspar. In detail this thin limestone sequence is very diverse; not only does the size of individual fossil populations vary greatly from layer to layer, but also the relative proportions of quartz and the various faunal elements change.

A poorly exposed sequence of fractured olive-green shales overlies the limestone. Some thin limestone beds toward the top of the shales resemble the limestone described in the previous paragraph, in that they are potpourris of quartz and fossil detritus. However, in most cases these limestones are rather dirty grainstones cemented by ferroan and nonferroan carbonate.

The shales are overlain by a quartz-rich sandstone sequence which is the lowest unit that can be correlated between the two hills. In the eastern hill the sandstone is more or less homogenous, consisting of medium-scale trough cross beds 1–8 in. thick. Numerous green clay clasts help define the cross bedding. Cross-bed vectors are bipolar and generally similar in trend to those found in the lower sandstone (i.e., 115–305°). The sandstone in the western hill is divided by a green shale bed. The sandstone unit below the interbedded shale is characterized by both trough and, less commonly, planar cross bedding. The upper sand unit has a sole-marked base and displays horizontal burrows.

Strata above the sandstone sequence are a very variable sequence of thin green shales, thin limestones (mostly grainstones with an abundant biota and variable amounts of quartz), and a single quartz-rich cross-bedded sandstone with an erosive base (seen only in the eastern hill). Some of the limestones display trough cross bedding in sets as much as 6 in. thick; some shell imbrication is present. The quartz sandstone bed displays abundant *Skolithos* burrows on its top surface.

The contact between the Mountain Lake and Pooleville Members is placed ~30 ft below the Viola/Bromide contact, at the base of a ledge-forming biomicrite bed which contains abundant *Receptaculites* (Fig. 2). At the eastern hill this bed contains 10–15% quartz sand and is overlain by ~3 ft of interbedded biomicrite and green shale, followed by a mostly covered sequence of thinly bedded biomicrites. However, on the western hill the Pooleville (which is completely exposed) contains practically no quartz and clay, and it consists

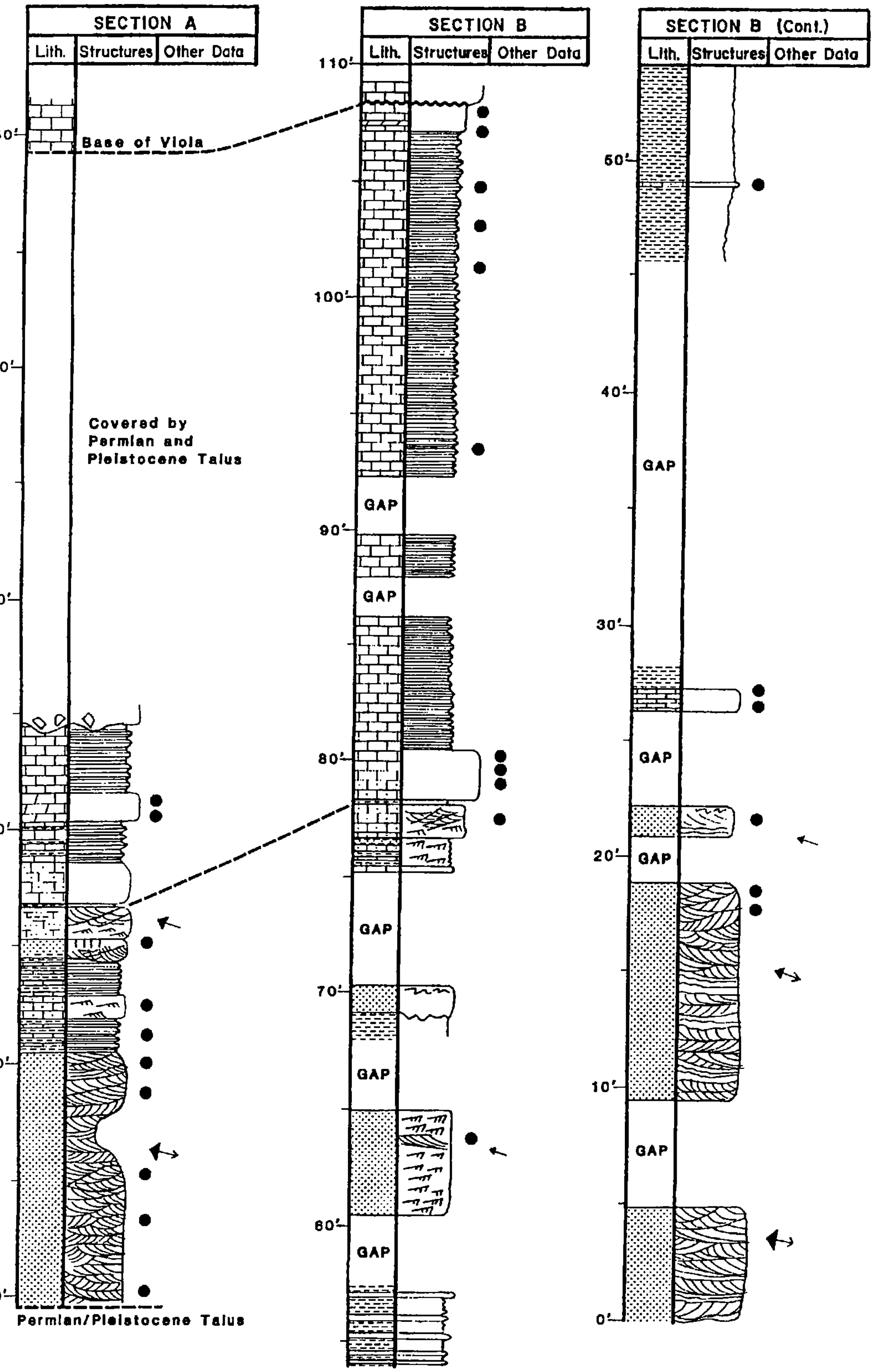

Figure 2. Log of the Bromide Formation as exposed in the eastern (section A) and western (section B) Sugar Hills. Symbols follow conventional usage (arrows give cross-bedding vectors, dots are thin-section sites). Dotted line connecting sections A and B is proposed boundary between Pooleville and Mountain Lake Members.

of highly fossiliferous biomicrites, containing several hardgrounds. The fauna in these beds is diverse (Table 1) and variable in the size, sorting, and proportions of species from layer to layer. Many of the fossils are whole and show little sign of abrasion. The limestones have been partially replaced by ferroan dolomite, which is to some extent fabric selective, being particularly well developed in and around burrows.

Environment of Deposition

Quartz-Rich Sandstones

Well-washed, trough cross-bedded sandstones with reactivation surfaces, some *Skolithos* burrows, and bipolar cross-bedding vectors occur at two principal horizons (Fig. 2). These sandstones are interpreted as sand bodies molded by tidal currents that moved parallel to the long axis of the southern Oklahoma aulacogen. In both cases the northwesterly flow was stronger than that to the southeast. A similar pattern is developed in the underlying Upper Cambrian Reagan Formation in the same paleogeographical situation (Donovan, 1986). Little can be said as to the large-scale architecture of the beds, except that the upper sand body thickens to the east (Fig. 3).

Green Shales

Poorly exposed green shales, consisting of a mixture of chlorite and illite, constitute a considerable proportion of the section. They contain little or no organic matter, may be delicately laminated, but are mostly unburrowed and are interpreted as a below-wave and storm-base deposit laid down in quiet waters under mostly sterile bottom conditions.

Fossiliferous Lime Mudstones

Most of the limestones in the section are fossiliferous lime mudstones which can be classified either as biomicrites or wackestones (Fig. 4). Hardgrounds are common and many beds are burrowed. As noted previously, beds show a great deal of variety in the proportions of species which they contain. Furthermore the average size of individuals of the same species varies from layer to layer. Sorting is highly variable.

Limestones of this type are interpreted as carbonate platform deposits which were ultimately deposited in calm waters below wave base. The poor sorting shown by many of the deposits suggests periodic reworking of the sediments by storms. On the other hand, hardgrounds record periods of sea-floor stability and penecontemporaneous cementation. Some of these hardgrounds clearly acted as surfaces of colonization for a variety of organisms. The faunal variations noted above suggest that the platform was ecologically partitioned.

Fossiliferous Grainstones

Fossiliferous grainstones are found in two settings: (1) as thin, poorly sorted "dirty" beds associated with green shales; and (2) as well-sorted "clean" cross-bedded sandstones that contain variable amounts of quartz grains (Fig. 5). The former are interpreted as allochthonous storm

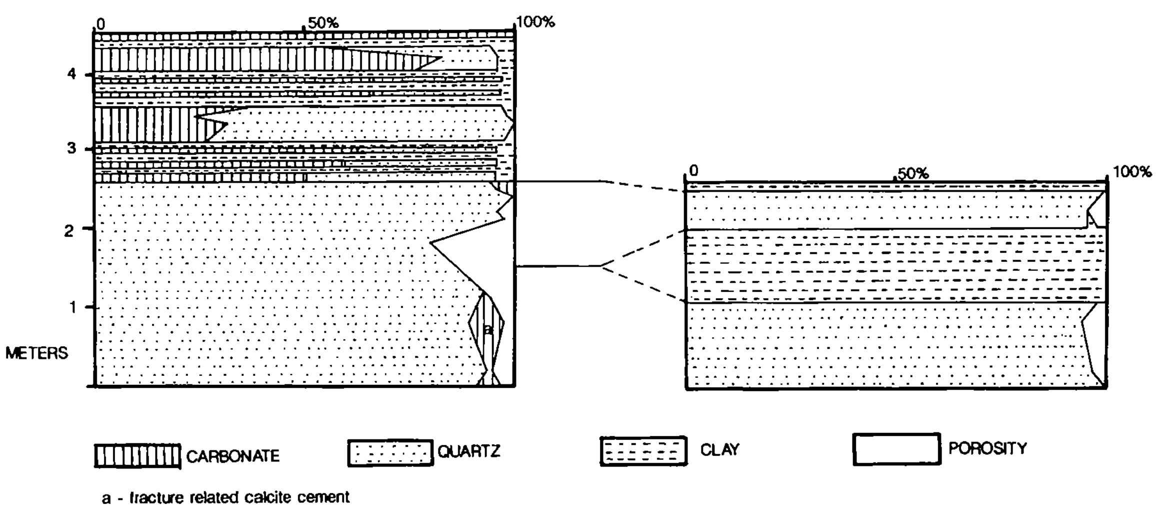

Figure 3. Petrographic comparison of the two sections (the right-hand section is the western exposure). The interval shown is the upper part of the Mountain Lake Member. Note the facies changes and the restriction of significant porosity to the interior of the quartz-rich sandstone beds.

Figure 4. Typical fossiliferous mudstone of the Pooleville Member. Large fossil is cf. *Apodasmocrinus daubei* (Warn and Strimple, 1977). Field of view is 4 mm wide.

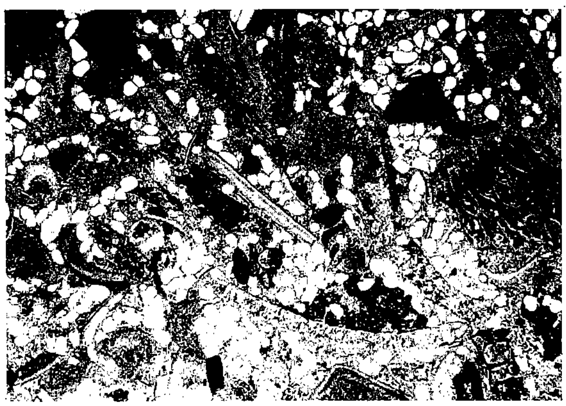

Figure 5. Sandstone of mixed provenance from the upper part of the Mountain Lake Member. Principal fossils are broken shell valves, echinoderms, and bryozoans. Notice packing of quartz grains between imbricate shell fragments. Field of view is 12 mm wide.

deposits proximal to a carbonate shelf (a similar interpretation to that of Longman, 1981). The latter are interpreted as tide-influenced sand shoals.

A Possible Sequence of Events

The general paleogeography of Mountain Lake times has been addressed by Schramm (1964), Longman (1981), and Munsil (1983), all of whom note that shale is more common along the axis of the southern Oklahoma aulacogen than on the surrounding cratonic areas and therefore suggest that the aulacogen was a somewhat deeper-water "gulf" at this time. Within this framework the Sugar Hills outcrops are an erosional remnant sited on the southwestern edge of the "gulf." To the south lay the old carbonate platform of the Texas arch.

Previous investigations have emphasized that the complexity of the Bromide has resulted from subtle shifts in various paleogeographic elements (Schramm, 1964). This emphasis is accepted here with two important caveats. Firstly, the presence of a phosphatic hardground suggests that parts of the section may well be condensed; thus it is reasonable to assume that variations in clastic sediment supply related to differing sea-level positions occurred.

Secondly, it is possible that interpretations of the Simpson Group as a whole may have to be revised in the light of the current debate as to the nature of the "fundamental stratigraphic unit." In this context it is pertinent to note that the underlying Arbuckle Group, which was deposited at a rate similar to the Simpson Group (Donovan, 1986), displays a record which is dominated by short (~6 ft) shallowing-upward cycles, each of which records a transgressive/regressive event. These cycles may very well be Milankovitch-driven and are similar to the punctuated aggradational cycles of Anderson and Goodwin (1990). These authors note that the existence of such cycles calls into question our current practice of applying Waltherian interpretations to classical formational divisions of the size of the Bromide Formation.

Such cycles have yet to be recognized in the Simpson Group; present models suggest that the introduction of large amounts of quartz detritus reflects relatively low sea-level positions and that a carbonate factory of the order of that recorded by the Pooleville Member is a record of a relative high stand (e.g., Longman, 1981). If this is the case, then the Bromide Formation as a whole would seem to be a record of a single long-term transgression. However, although not complete enough to permit a perfect "read," the Sugar Hills succession appears to record a more complex situation, involving several minor transgressive/regressive events and/or tectonically controlled adjustments. Thus it may be that equivalents to the cyclic patterns in the Arbuckle Group are present but are either masked by longer scale variations related to changes in sediment type or have yet to be recognized.

PETROLOGY

Petrologic complexity is the hallmark of these rocks. While it is possible to recognize three end members—quartz-rich sandstones, shales, and limestones—many of the rocks consist of mixtures of various types of grains (Figs. 5,6). Not surprisingly, the diagenetic history of these rocks is intricate.

Quartz-Rich Sandstones

Sandstones of this type consist principally of unstrained monocrystalline quartz grains, with minor but constant amounts (1–3%) of feldspar and phosphatic grains (including abraded conodonts and small nodules). These three components constitute a discrete detrital population. In addition, variable amounts of green shale clasts and various fossils are present in positions close to the contact of the sandstones with either shales or limestones.

Quartz, feldspar, and phosphate grains are generally of fine to very fine size and are very well rounded. Sorting is more variable, from moderate to good, depending to some extent on the nature of the sedimentary structures in the rock.

In some sandstones there are appreciable amounts of green clay matrix, but for the most part the primary grain populations were well washed. Subsequent cementation has involved principally quartz and, more locally, calcite and authigenic illite.

Quartz cement is seen as syntaxial overgrowths and is ubiquitous, except where inhibited by clay or calcite. Contacts between quartz grains are tangential, long, or slightly concavo-convex; sutured contacts, similar to those recorded by Houseknecht (1987,1988) in the Bromide near Norman, Oklahoma, are not present. In general it appears that the sandstones have been slight net importers of quartz.

Calcite cements are developed adjacent to areas where the quartz-grain populations are mixed with shell fragments. The earliest calcite cement is nonferroan; later cements are increasingly iron-rich. In such rocks the shells and quartz grains have acted as templates for calcite and quartz precipitation, respectively. This partitioning has resulted in complex and apparently contradictory timing relationships between the two cements (Fig. 7).

Clay is present as both rims of detrital illite and as authigenic sproutings of illite crystals. The amount of detrital illite varies from almost none to a complete infill of pore spaces between detrital grains. Wherever detrital illite is present at a grain boundary, it has inhibited the precipitation of quartz overgrowths. Authigenic illite lines pore walls in most of the sandstones examined, invariably postdating syntaxial quartz overgrowths. However, illite of this type is patchily distributed and is not found in the majority of pores. In addi-

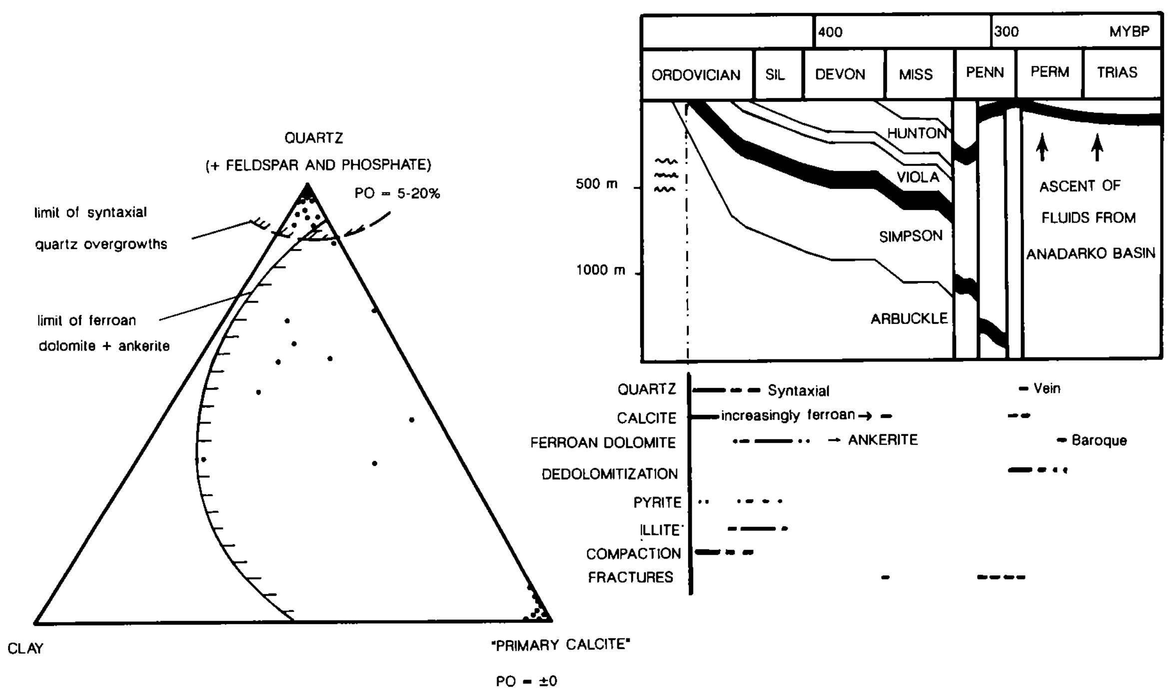

Figure 6. Summary of Bromide diagenesis. Triangular diagram shows range of sandstone types; various mixtures of siliciclastic and calcite grains with clay. Note how both syntaxial quartz overgrowths and porosity are restricted to clean quartz-rich sandstones. Burial curves (derived from Donovan, 1986) illustrate how the Simpson Group was gradually buried beneath Lower Paleozoic sediments until, in Pennsylvanian times, the southern Oklahoma aulacogen was dismembered into two parts, one of which was buried to great depth in the Anadarko basin, the other uplifted to the surface (and partly eroded) in the Wichita uplift.

tion to the illite, a trace amount of kaolinite is present in some pores; this clay also postdates quartz cementation and is very patchily distributed.

Limestones

Although most limestone textures are mud-supported, a spectrum of textures, from grainstones to very sparsely fossiliferous mudstones, is found; as noted previously, variable amounts of detrital quartz and clay may be present. In addition, one or two very "dirty," quartz- and clay-rich limestones in the Mountain Lake Member contain highly weathered glauconite grains.

Almost all carbonate grains observed are fossils, either complete or fragmented (Fig. 4); a few peloids and intraclasts are present, but no ooids were seen. Early calcite cement textures include syntaxial overgrowths around echinoid fragments and the infill of inter- and intra-particle porosity by drusy sparite. As noted previously, later calcite cements are increasingly ferroan, even within single drusy crystals.

Many of the limestones have been partially replaced by varieties of iron-rich carbonate; although this replacement carbonate is highly oxidized to iron oxide, both iron-rich dolomite and ankerite are present (Fig. 8). Some replacement is fabric-selective in that it is clearly related to burrows, whereas in other cases there is no obvious site control. In general, mud (i.e., micrite) is more commonly replaced than fossils, suggesting that, in detail, replacement was controlled by fluid movement along crystal boundaries. As a consequence, larger calcite crystals (i.e., fossils) maintained their integrity. In addition, some ferroan dolomite cements are locally developed within fossils, mostly bryozoans. Replacing carbonate crystals are euhedral dolosparite; bigger crystals are usually zoned. Similar ferroan dolomite has been recorded in other studies of the Simpson Group (e.g., Pitman and Burruss, 1989; Pollastro, 1989); the Sugar Hills section adds a further complexity to the carbonate story, because much of the ferroan dolomite and ankerite have been dedolomitized to nonferroan calcite and iron oxide (Fig. 8).

Two minor replacements seen in the limestones are fibrous chalcedony (which selectively replaces the inner parts of some brachiopod shells) and pyrite (seen as euhedra which also may partially replace shell fabrics). The pyrite is invariably partially or completely oxidized.

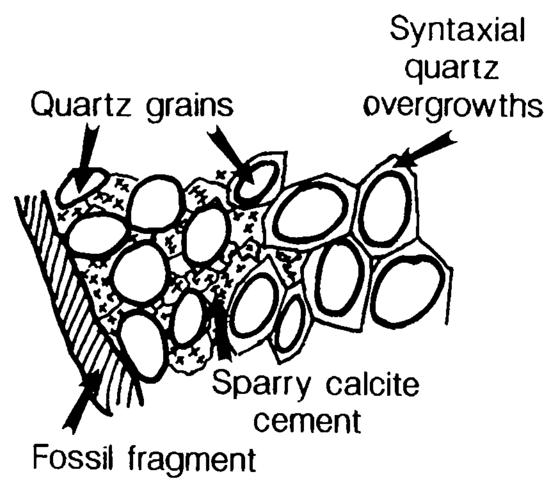

Figure 7. Sketch of complex, early diagenetic relationships in a quartz-rich sandstone which contains a small percentage of shells. Adjacent to the fossil fragments quartz grains are completely cemented by calcite, whereas away from the fossil syntaxial overgrowths have almost destroyed primary porosity. In the central area of the sketch the cement layers suggest that calcite postdates syntaxial quartz. Most probably cementation of quartz and calcite was contemporaneous, but the calcite preferentially precipitated on the existing template (i.e., the fossil). As calcite precipitated at a faster rate than quartz, it eventually crystallized on the syntaxial overgrowths which had formed in areas farther from the fossil.

Stylolites

The limestones are cut by a number of bedding-parallel stylolites; these are mostly nonpersistent and of low amplitude. Their geometry suggests that little carbonate has been removed from the section by pressure solution.

Veins

Two types of veins cut the section. Those which are subperpendicular to bedding are invariably filled by calcite; they are of several ages and show complex infillings of ferroan and nonferroan calcite. A second type of vein is parallel to bedding and is filled by nonferroan calcite, some of which is fibrous with fibre orientation perpendicular to bedding. Quartz grains near such veins are cut by calcite-filled fractures parallel to bedding. These geometrical considerations suggest that the second type of vein formed during uplift of the section.

Porosity

The only significant porosity encountered in the present study occurs in the quartz-rich sandstones of the Mountain Lake Member (Fig. 3). Most of this is reduced primary porosity, and it varies from 0 to 20%. Four factors have contributed to porosity destruction: detrital-clay infiltration, carbonate cementation, syntaxial quartz overgrowths, and authigenic illite formation (minor).

Figure 8. Limestone from the Pooleville Member showing replacement by idiotopic rhombs of ferroan dolomite. Most of the rhombs have been dedolomitized and consist of nonferroan calcite and hematite. Field of view is 4 mm wide.

Where present, detrital clay and carbonate cements (usually associated with detrital carbonate) have effectively destroyed porosity. Hence porosity is only found in those sandstones which are well-washed and are spatially away from accumulations of detrital carbonate. A typical sandstone with porosity contains some syntaxial quartz overgrowths and a little authigenic illite. In essence, the central areas of sandstone beds are porous and the edges are nonporous (Fig. 3).

Perigenesis

Because of its structural position, the Sugar Hills section records a different perigenetic history to adjacent sequences of Bromide in the Anadarko basin. While the latter were buried to variable, though substantial, depths as much as ~30,000 ft, the Sugar Hills were buried to no more than ~4,500 ft (calculated from isochores in Johnson and others, 1988). Furthermore, since uplift during the Pennsylvanian, the Sugar Hills have always been at shallow depth, if not actually exposed at the surface (as they certainly were in the Early Permian) (Fig. 6). Support for this suggestion can be found in the work of Cardott and Lambert (1985) and Cardott (1988). Using vitrinite-reflectance values obtained from the Upper Devonian Woodford Formation, these authors found that the Woodford in the frontal fault zone was characterized by reflectance values of <0.5. Such values are indicative of burial depths of 4,000–5,000 ft and indicate that the section has been heated to no more than 50–65°C. Cardott (1989) suggested that the Woodford in the frontal fault zone achieved its maximum heating prior to Pennsylvanian uplift and deformation. Given that the Bromide is no more than 2,000 ft below the Woodford in the Wichita Mountain area, it can be assumed that the maximum temperature to affect the Sugar Hills was no more than ~70°C, and that this heating took place during the Late Mississippian. Thus an important perigenetic interest of the Sugar Hills section is that it constrains the timing and depth of diagenetic reactions in the Bromide. For example, the widely documented transformation of smectite to illite takes place at temperatures between 60 and 110°C. It has been suggested that this transformation released the ions needed to convert calcite to ferroan dolomite and ankerite, which is commonly found in the Bromide (Pollastro, 1989). In the Sugar Hills, ferroan-carbonate replacement of calcite is very commonly observed; yet no smectite has been detected, and authigenic illite is not abundant. Given the temperature controls noted above, it seems unlikely that the clay transformation was the source of the ions needed for carbonate replacement.

Several recent studies (Houseknecht, 1987, 1988; Pitman and Burrus, 1989; Pollastro, 1989) have examined Bromide cores at depths ranging from 8,000 to 17,000 ft. In general the perigeneses described by these authors greatly resemble that developed in the Sugar Hills sections. Perhaps the most significant difference is the amount of pressure solution which has affected the more deeply buried quartz-rich sandstones. In addition, porosity in the more deeply buried sections is generally secondary, whereas in the Sugar Hills primary porosity as much as 20% is found.

In essence, two periods of diagenetic activity can be recognized: pre- and post-Pennsylvanian uplift. The earliest events involved sediment compaction and very early, partly contemporaneous cementation of detrital carbonate accumulations by nonferroan calcite. This calcite is essentially either drusy or syntaxial around echinoid fragments; there is no evidence of vadose cementation, nor are isopachous rims developed. With time this calcite became increasingly ferroan; perhaps the simplest interpretation is that this phase of carbonate precipitation took place under transgressive conditions during which an increasingly reducing water column evolved.

Early silica mobility was in part contemporaneous with early carbonate cementation (Fig. 6), and this is seen in the partial replacement of the inner parts of some shells by fibrous chalcedony in limestones and the growth of syntaxial overgrowths in quartz-rich sandstones. The former may have resulted from the interaction of alkaline, carbonate-saturated ground waters with locally more-acidic zones resulting from decay associated with individual shells. The source of silica in the quartz-rich sandstones initially may have been from contact with similar ground waters. However, with time it is probable that much of the silica cementation resulted from moderate pressure solution associated with Paleozoic burial.

At some time in the Paleozoic, a second phase of carbonate precipitation took place. This involved the widespread (as much as 20%) replacement of limestone textures by euhedral, idiotopic rhombs of ferroan dolomite and ankerite. Some, but not all, of this replacement is fabric selective; i.e., it is related either to burrows (which presumably functioned as permeability pathways) or to fossils, particularly bryozoans. The source of iron and magnesium for the replacement has been attributed to the conversion of smectite to illite (Pollastro, 1989). Certainly some authigenic illite is present in small amounts in the quartz-rich sandstones. However, as noted previously, this process is unlikely to have been effective within the Bromide itself at the depth of pre-Pennsylvanian burial in the Sugar Hills. An alternative solution might be the migration of these ions from pressure solution of dolomites in the underlying Arbuckle Group.

In summary, the early Paleozoic history of the Sugar Hills sequence involved eodiagenetic compaction and cementation by calcite and, to a lesser extent, secondary silica. This was followed by mesodiagenetic adjustment to increasing burial depth, chiefly recorded by the replacement of calcite by ferroan dolomite and ankerite in limestones, and by pressure solution plus secondary silica precipitation in quartz-rich sandstones.

Following Pennsylvanian orogeny, the Sugar Hills were uplifted to the surface and tectonically emplaced (thrust) above the Anadarko basin, in a series of movements which had ceased by the Early Permian (Fig. 6). Some bedding-parallel fracturing records this uplift, while extensive dedolomitization records surface exposure. Emplacement of the sequence by thrusting above the Anadarko basin meant that the sequence was subjected to the upward movement of warm fluids from the basin. Thus vertical veins cutting the limestones contain baroque dolomite rhombs, while some pores in the quartz-rich sandstones contain degraded hydrocarbon. In this context, it is pertinent to note that the overlying limestones of the Viola Formation are cut by karst fissures which contain hydrocarbon-impregnated speleothems, similar to those described on nearby Bally Mountain (Donovan, 1987).

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GEOLOGY, GEOCHEMISTRY, AND PLATINUM-GROUP-ELEMENT MINERALIZATION OF THE CAMBRIAN GLEN MOUNTAINS LAYERED COMPLEX AND ASSOCIATED ROCKS, SOUTHWESTERN OKLAHOMA

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ABSTRACT.—The Cambrian Glen Mountains Layered Complex (GMLC) is the oldest exposed igneous rock unit of the uplifted part of the southern Oklahoma aulacogen. It underlies an area at least 64 × 32 km, making it comparable in size to other large mafic layered intrusions.

The GMLC, as exposed, is composed of at least three major and several minor cyclic units. Each cyclic unit is, ideally, characterized by a layer of plagioclase–olivine cumulate overlain by plagioclase–olivine–pyroxene, plagioclase–pyroxene, and plagioclase cumulate layers. Plagioclase–olivine, plagioclase–olivine–pyroxene, and plagioclase–pyroxene cumulate parts of cycles are moderately to strongly adcumulate and characterized by an upward decrease in the amount of olivine and upward increase in the amount of pyroxene and oxide. An individual cyclic unit may contain a number of vertically and laterally variable rhythmic layers of these two- and three-phase cumulate rocks. The plagioclase cumulate parts of the cyclic units are orthocumulate with poikilitic ilmenomagnetite and oikocrystic clinopyroxene. Some plagioclase cumulate layers contain accessory sulfide mineralization which forms laterally continuous zones parallel to phase layering and plagioclase lamination.

Whole-rock $\delta^{18}O$ and $\delta^{34}S$ values range from +1.9 to +7.2 per mil and –1.8 to +2.9 per mil, respectively. Eighty-five percent of the samples have $\delta^{18}O$ values between 5.7 and 7.2 per mil, typical of "normal" gabbroic and anorthositic rock types. Evidence for substantial crustal contamination of parent melts is not indicated, based on either S or O isotopic values. Fifteen percent of the analyzed samples are characterized by depletion in $\delta^{18}O$, with $\delta^{18}O$ values between 1.9 and 3.6 per mil. These values occur in plagioclase cumulates adjacent to cross-cutting felsic Cold Springs intrusions, and are thought to be caused by isotopic exchange with ground waters heated during emplacement of the felsic intrusions at shallow crustal levels.

Olivine and orthopyroxene phase chemistry indicates that olivine varies from Fo_{60} to Fo_{74}, and coexisting orthopyroxene from En_{65} to En_{77}. The data, combined with published analyses, suggest an upward Fe enrichment trend that is periodically "reset" to more-Mg-rich compositions due to magma replenishment and/or new influxes of slightly more primitive magma that, in general, correspond to defined cyclic units.

Platinum-group-element mineralization occurs in plagioclase cumulate layers with disseminated (<2%) sulfide mineralization. Locally it is associated with, or adjacent to, discontinuous plagioclase–olivine ± pyroxene cumulate layers within larger plagioclase cumulate units. Sulfides include pyrite, chalcopyrite, pyrrhotite, and pentlandite, in order of abundance.

The Roosevelt Gabbros represent a series of sill- and dike-like bodies that intrude the GMLC. The previously identified "G" zone of the Glen Mountains Layered Complex is included as one of the Roosevelt gabbro series of intrusions, based on detailed drill-core logging and geologic mapping.

INTRODUCTION

The lithostratigraphy described in this paper follows, in general, that suggested or defined by Ham and others (1964) and Powell and others (1980). The principal difference deals with the internal stratigraphy of the Glen Mountains Layered Complex (GMLC). The internal units of the GMLC (the "K", "L", "M", "N", and "G" zones) suggested by Spencer (1961), and further characterized or defined by Powell and others (1980) and Stockton and Giddens (1982), are not used or recognized in this paper, based on detailed mapping by the author (Cooper, 1987,1988).

The GMLC is a Cambrian anorthositic layered intrusion and is the oldest exposed igneous rock unit within the Wichita uplift. Previous work, dealing primarily with the western two-thirds of the GMLC, includes articles and reports by Gilbert (1960), Hiss (1960), Spencer (1961), Karns (1961), Frech (1962), Scofield (1968,1973,1975), Phelps (1975), Alipouraghtapeh (1979), Powell and others (1980), Stockton and Giddens (1982), Stockton (1984), Powell (1986), Cooper and others (1986), and Cooper (1986).

Associated with the GMLC is a series of mafic and felsic intrusive rocks. From oldest to youngest, these include the Roosevelt Gabbros, fine-grained intermediate to mafic dikes, felsic Cold Springs intrusions, and a second series of fine-grained mafic dikes. The Roosevelt Gabbros include several distinct and different hydrous (biotite–amphibole) gabbro intrusions and related dikes which intrude the GMLC. The intrusions have tholeiitic to slightly alkaline characteristics common to "within-plate" mafic rocks (Cameron and others, 1986). Fine-grained intermediate to mafic dikes and sills, some with chilled margins, crosscut the Roosevelt Gabbros and GMLC and are in turn intruded by the felsic Cold Springs intrusions. These intermediate dikes/sills, Otter Creek microdiorite of Powell and others (1980), have been further characterized by Vidrine and Fernandez (1986).

The felsic Cold Springs intrusions are composed of tonalite, granodiorite, and local granite (Vidrine and Fernandez, 1986). Detailed mapping (Cooper, 1987) indicates that the larger intrusions tend to be circular laccolith- to sill-like bodies that are up to 225 ft thick in the center and taper or thin outward, with an overall diameter of 1–2 mi. In addition, the margins or edges of individual intrusions appear to overlap and interfinger in some instances. Field and drill-core relationships indicate that the Cold Springs intrusions were often emplaced along preexisting sills and/or dikes of Roosevelt gabbro and microdiorite.

Fine-grained mafic dikes crosscut all of the igneous rock units described above. The dikes often have a diabasic texture and have been characterized as tholeiitic, based on geochemistry (Gilbert and Hughes, 1986; Cameron and others, 1986).

GLEN MOUNTAINS LAYERED COMPLEX

Stratigraphy and Cyclic Units

The principal exposures of the Glen Mountains Layered Complex (GMLC) can be subdivided into three geographic areas (Fig. 1); these include the

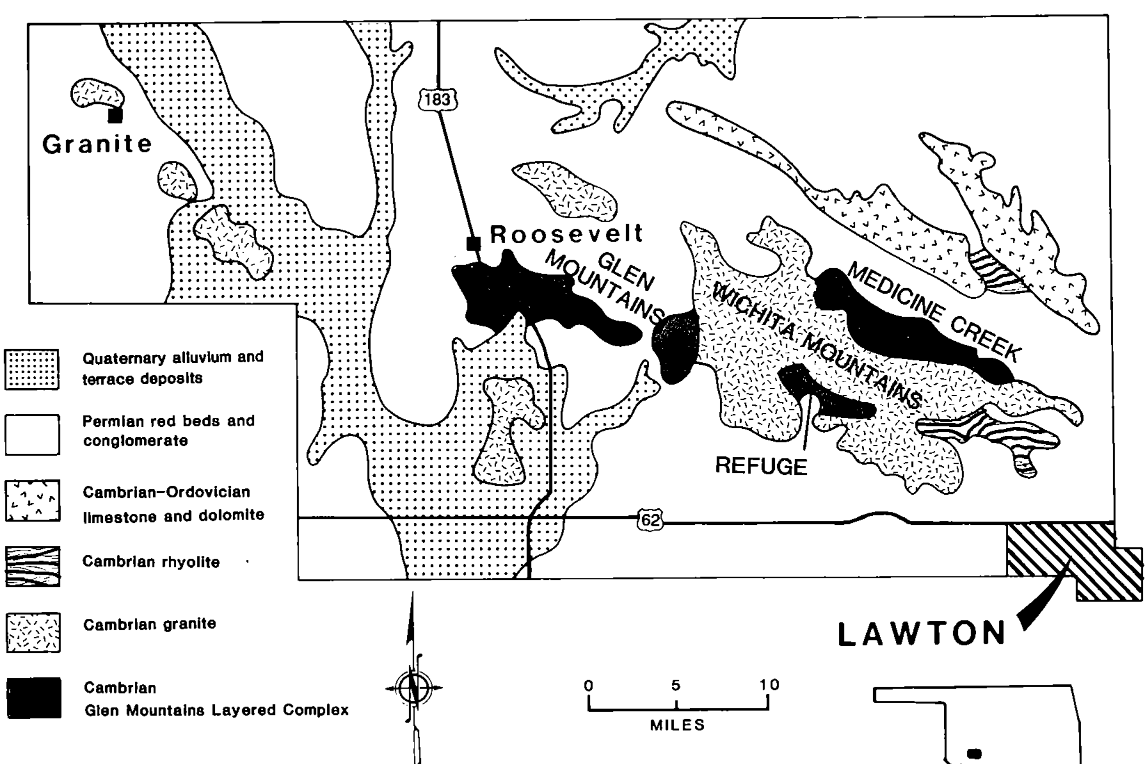

Figure 1. Geographic divisions of the Glen Mountains Layered Complex and generalized geology of the Wichita uplift. Geology modified from Powell and others (1980).

Glen Mountains proper and adjacent exposures near the town of Roosevelt; the Refuge area in the Wichita Mountains National Wildlife Refuge; and the Medicine Creek area near Meers. Only the Glen Mountains proper and some of the exposures near Roosevelt will be dealt with here, because the best, most extensive, and easily accessible exposures of the GMLC occur in this area; a considerable amount of detailed mapping has been done in this area recently.

The reports and work prior to 1986, cited above, emphasized the descriptive nomenclature and stratigraphy suggested by Spencer (1961) to identify and distinguish major rock units. These rock units, the "K", "L", "M", "N", and "G" zones, were further characterized and defined by Powell and others (1980) and Stockton and Giddens (1982). Detailed mapping by the author in Cooper and others (1986) and Cooper (1987,1988) applied the cumulate terminology described by Irvine (1982) to identify and trace major rock units within the GMLC. This resulted in a simplified and more straightforward way to identify and map rock units and interpret the internal geologic relationships within the GMLC. It successfully separated similar-appearing rock units, which previously had been combined, into stratigraphically distinct units (cycles), each of which is composed of a rhythmic sequence of igneous layers. The nature of the cycles and contained igneous layers is further described below.

In the Glen Mountains–Roosevelt area, detailed geologic mapping (Cooper, 1987,1988) indicates that the GMLC is composed of at least three major cyclic units. These cyclic units do not in any way correlate with the previously defined zones of Spencer (1961), Powell and others (1980), and Stockton and Giddens (1982). As defined by the detailed mapping, each major cycle consists of a sequence of igneous layers which form a unit that can be traced laterally for a considerable distance (Fig. 2). Internal to each of the major cycles are minor, incomplete, or abbreviated cycles. They consist of a partial sequence of one or more igneous layers which can be traced as a map unit within a major cycle for short distances. The major cycles are interpreted to represent the crystallized product and fractionation of one or more major influxes of magma. Although the major cycles exhibit vertical and lateral variations, in addition to the minor cycles and layers, some general characteristics and features are common to all of the major and many of the minor cyclic units.

Each major cyclic unit consists of a lower part

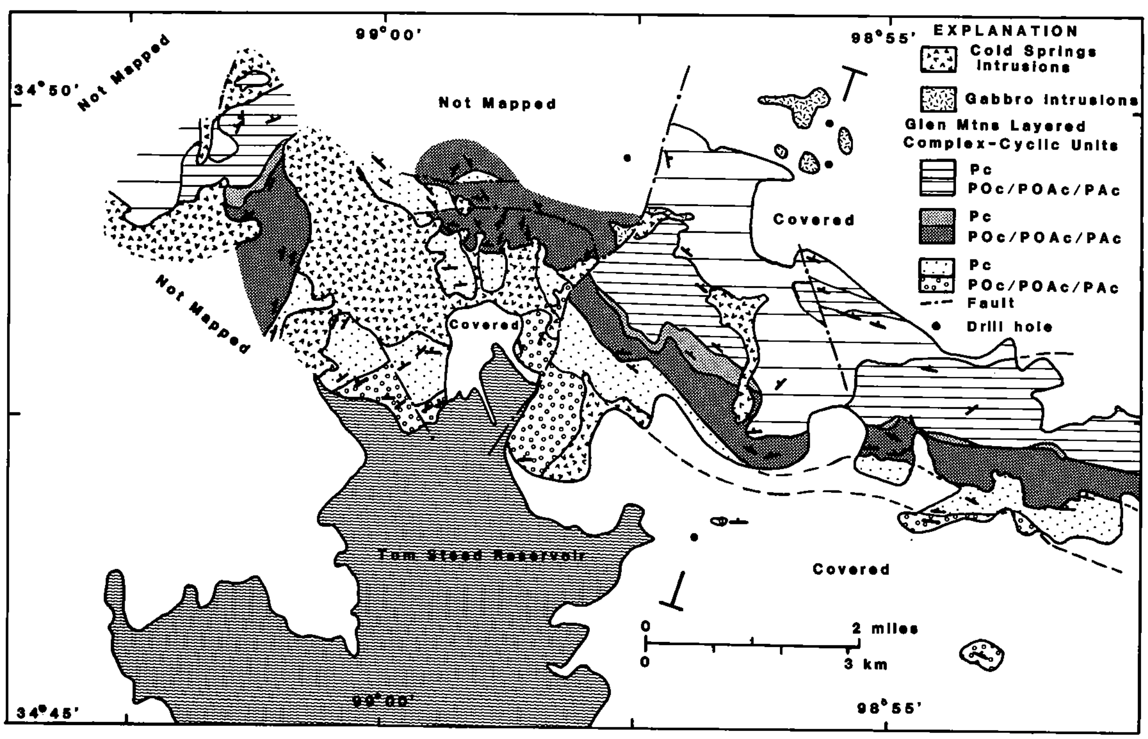

Figure 2. Geology of part of the westernmost Glen Mountains Layered Complex and associated rocks in the Roosevelt and Glen Mountains 7.5′ Quadrangles. Geology modified from Cooper (1987,1988). POc = plagioclase–olivine cumulate; POAc = plagioclase–olivine–pyroxene cumulate; PAc = plagioclase–pyroxene cumulate; Pc = plagioclase cumulate. Line of cross section of Figure 5 indicated (⊢ ⊣).

characterized by an adcumulate texture with variable amounts of cumulus plagioclase, olivine, and pyroxene. The term cumulus refers to the presence of well-defined and outlined crystals which form a framework of crystals and reflects the fractionation of the parent magma (Irvine, 1982). The upper part of each cycle has an orthocumulate to mesocumulate texture. Plagioclase is the only cumulus phase, with pyroxene and ilmenomagnetite as the principal interstitial or intercumulate phases.

The modal and crystal-size variations within an

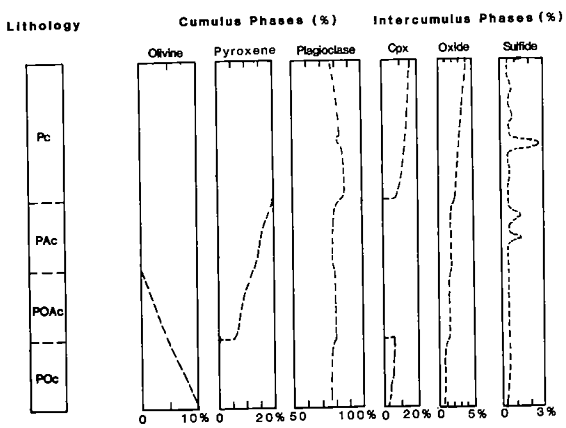

Figure 3. Modal abundance of cumulus and intercumulus phases of an idealized cyclic unit in the Glen Mountains Layered Complex. Symbols as in Figure 2.

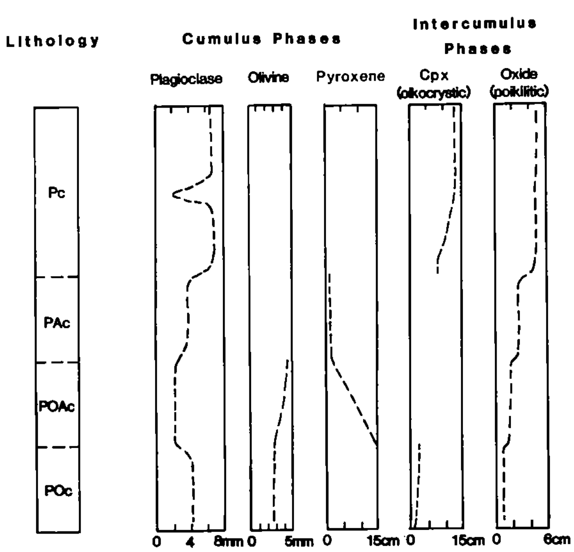

Figure 4. Crystal-size variation of cumulus and intercumulus phases of an idealized cyclic unit in the Glen Mountains Layered Complex. Symbols as in Figure 2.

idealized cycle in the GMLC is illustrated in Figures 3 and 4. In the lower part of the cycle, cumulus olivine decreases in abundance upward while the cumulus pyroxene content increases. In addition, the amount of peritectic orthopyroxene surrounding individual olivine crystals appears to increase with stratigraphic height within the olivine-bearing parts of the cycle. Cumulus plagioclase content fluctuates with no definite trend. As shown in Figure 3, the cumulate rock type changes upward from a plagioclase–olivine cumulate (POc) at the base to a plagioclase–olivine–pyroxene cumulate (POAc), followed by a plagioclase–pyroxene cumulate (PAc). Ilmenomagnetite is the primary intercumulate phase and shows a slight upward increase in abundance in this part of the cycle. Trace amounts of sulfide are present, up to 1% occurring locally in the plagioclase–pyroxene cumulate parts of a cycle. The general variation in crystal size of the cumulus and intercumulus phases is shown in Figure 4. The upper part of the cycle consists of a plagioclase cumulate with interstitial clinopyroxene and oxide. Texturally this part of a cycle is an orthocumulate to mesocumulate with well-defined plagioclase crystals that form the framework of the rock. Clinopyroxene is the dominant interstitial phase, with 1–4% ilmenomagnetite being present. In addition, trace to 1% sulfide is present, with up to 3% occurring locally (Figs. 3,4).

Detailed geologic mapping indicates that internal to the major cycles shown in Figure 2 there are abbreviated or incomplete cycles (Cooper, 1987,1988). Thus, several repetitions (rhythmic layers) of the adcumulate plagioclase–olivine, plagioclase–olivine–pyroxene, and plagioclase–pyroxene cumulates may be present within any given cycle. This is more common in cycles which have a greater thickness of the two- and three-phase cumulate rocks. Thin, discontinuous units of plagioclase cumulate also occur locally within a sequence composed dominantly of two- and three-phase cumulates. In addition, thin, discontinuous layers of plagioclase–olivine ± pyroxene occur locally in the dominantly plagioclase cumulate parts of cycles.

A generalized cross section of the GMLC in the eastern part of Figure 2 incorporates the surface exposures and the results of diamond drilling by Placer, Inc. and Nerco Minerals Co. on the north and south sides of the GMLC exposures, respectively (Fig. 5). The principal geologic features include the cyclic units of the GMLC that are exposed at the surface; additional cyclic units occur to the south beneath the Permian cover. In addition, some of the internal relationships in the GMLC are shown. A discontinuous Pc layer within a POc/POAc/PAc sequence (Fig. 2) is also shown as being discontinuous in a downdip direction (1,000 ft vertically below the Cold Springs intru-

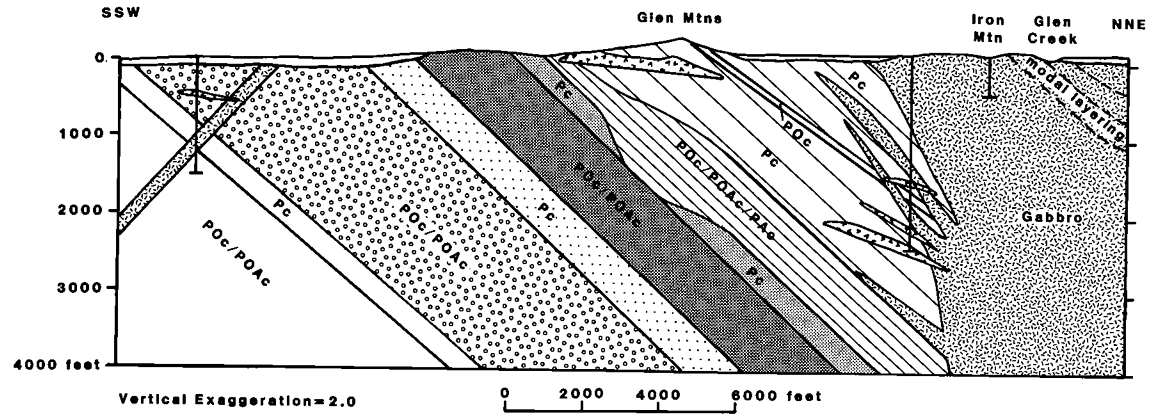

Figure 5. Generalized cross section of the Glen Mountains Layered Complex. Location of cross section shown in Figure 2; symbols as in Figure 2.

sion in the central part of Fig. 5). The channel feature referred to below (Fig. 6) is diagrammatically illustrated immediately below the same Cold Springs intrusion in the central part of Figure 5. However, on the north side, the GMLC is cut by a cylindrical intrusion of Roosevelt gabbro. The intrusion is informally referred to here as the Iron Mountain intrusion, after the local name for a small hill which contains the principal surface exposure of the intrusion. This area of exposure has previously been referred to as the "G" zone of the GMLC (Powell and others, 1980), or assigned to the "L" zone of the GMLC (Powell, 1986).

Detailed surface mapping and detailed logging of the drill core in the Iron Mountain area suggest that it represents a separate and distinct, crosscutting intrusion. Prominent olivine–plagioclase modal layering, developed on the north side of the principal surface exposure, is presently interpreted to have developed within the Iron Mountain gabbro intrusion during crystallization and is shown schematically in Figure 5. Other interpretations of the well-developed modal layering are possible, based on the surface mapping; one interpretation is that it represents the base or near-base of a major cyclic unit within the GMLC, along which part of the Iron Mountain gabbro intrusion was emplaced, as represented by the main portion of the surface exposure. Between the surface exposures (Iron Mountain and vicinity) and the GMLC to the south is a near-vertical zone of fine- to medium-grained, oxide-bearing "diabase" encounterd by one of the Placer, Inc. drill holes. It probably represents the chilled margin of the Iron Mountain intrusion. The appearance of the "diabase" is somewhat different from the "microgabbro" (Roosevelt Gabbro-related) dikes and sills described by Gilbert (1960), Spencer (1961), Powell and Gilbert (1982), and Gilbert and Hughes (1986). It appears to be finer-grained and to contain more oxide. Additional petrographic and geochemical work is needed before the "diabase" can be fully characterized. Offshoots of this intrusion occur as dike- and sill-like intrusions within the GMLC. The Glen Creek Gabbro, which has been described by Powell and others (1980) and Powell (1986), represents a later Roosevelt Gabbro intrusion that intrudes the GMLC, and probably the Iron Mountain intrusion. A crosscutting gabbroic dike also occurs in the Nerco Mineral Co. drill hole on the south side of the GMLC exposures.

Figure 6 is a schematic long section which illustrates two of the major internal features of the GMLC in the study area. The first is a large channel-like feature which appears to cut down and into stratigraphically lower cycles. The "channel" is filled by plagioclase cumulate characteristic of the upper parts of individual cycles, and when traced laterally away from the channel feature becomes the upper part of a cycle. One of the principal features of the "channel" is the presence of a zone of disseminated sulfide mineralization which occurs just above an internal contact within a cycle, but appears to carry across the "channel" at approximately the same stratigraphic level. The second major internal feature illustrated in the long section is the lateral variations that occur within some of the cyclic units. Individual plagioclase cumulate layers, as well as thin layers of adcumulate plagioclase ± olivine ± pyroxene cumulates, pinch out laterally. In Figure 6 a Pc unit present west of the "channel" is not present, or only present locally, east of the "channel" (see also Fig. 2). A minor POAc unit within the Pc layer of the lowermost major cycle also pinches out, as shown in Figure 6. In general the small-scale internal units appear to pinch out from west to east. This suggests either lateral changes in fractionation processes or a thickening of the GMLC as a whole to the west.

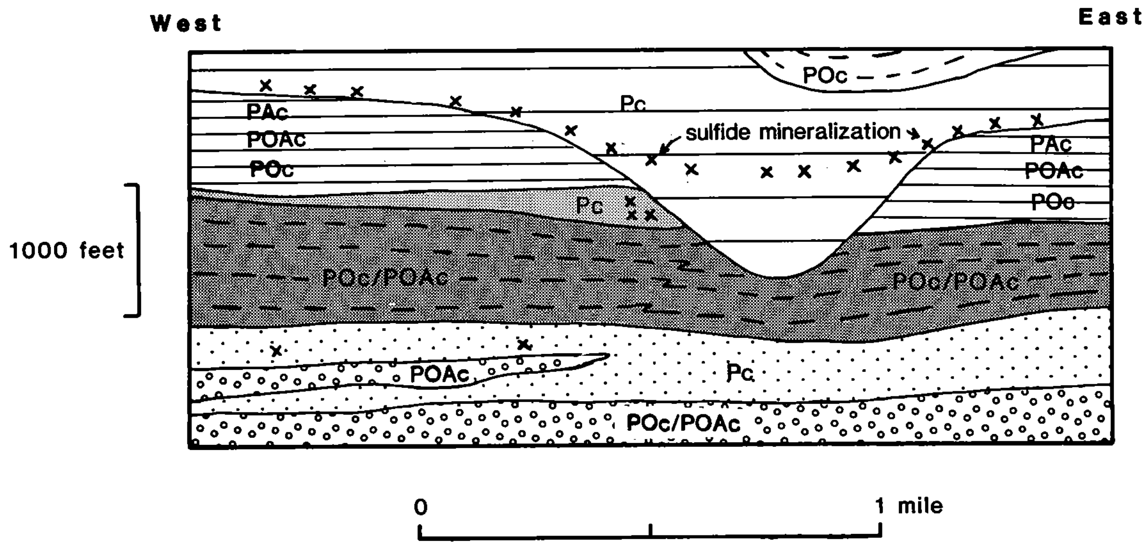

Figure 6. Generalized long section of the exposed part of the Glen Mountains Layered Complex in the eastern half of Figure 2. Symbols as in Figure 2.

Sulfur- and Oxygen-Isotope Data

Whole-rock sulfur- and oxygen-isotope determinations have been completed on selected samples of the GMLC. The upper diagram in Figure 7 illustrates the restricted range of sulfur-isotope values. All of the $\delta^{34}S$ values fall between 0 and +3, with one exception. The values compare favorably with $\delta^{34}S$ values from the JM reef and Picket Pin zone in the Banded series of the Stillwater complex. The sulfide present in both of these zones is interpreted to be magmatic sulfide, and the $\delta^{34}S$ values to be characteristic of magmatic sulfur (Zientek and Ripley, 1989). Oxygen-isotope values from the GMLC fall into two groups. One group of 14 samples has $\delta^{18}O$ values between +5.7 and +7.2, while a second group of 5 samples has $\delta^{18}O$ values between +1.9 and +3.6. The first group compares favorably with data from the Stillwater, Bushveld, and Kiglapait layered intrusions (Dunn, 1986; Keyser, 1986). The second group represents 5 samples collected from GMLC exposures immediately adjacent to felsic Cold Springs intrusions. This suggests that the oxygen-isotope system of the GMLC was locally disturbed by emplacement of the Cold Springs intrusions at shallow crustal levels, which resulted in isotopic exchange with ground water. Both the $\delta^{34}S$ and first group of $\delta^{18}O$ values (+5.7 to +7.2) are consistent with the GMLC being derived from a mantle source.

Olivine and Orthopyroxene Geochemistry

The compositional variation of olivine and orthopyroxene in a composite stratigraphic section in the eastern part of the study area is shown in Figure 8. It includes the three major as well as the minor cycles, and discontinuous igneous layers that were actually mapped. Olivine and orthopyroxene compositional data, derived from the literature and in the same area, which could be reasonably placed in their correct stratigraphic context, are also represented. Two conclusions are apparent. First, most of the available compositions from the literature were derived from the two lowermost cyclic units. Second, the olivine and orthopyroxene compositions determined during the present study, when combined with the literature values, show a complex relationship of forsterite and enstatite compositions, not only between but also within the major cyclic units. Olivine exhibits a possible magnesium-enrichment trend upward in the olivine-bearing part of the lowermost major cycle. Olivine in the second major cycle appears to have a relatively uniform forsterite composition. However, olivine in the third major cycle exhibits a magnesium-enrichment trend in the lower half of the olivine-bearing part, followed by a strong iron-enrichment trend in the upper half. This may indicate that more than one major magma pulse is represented within this very thick major cycle. The composition of coexisting and peritectic orthopyroxene in equilibrium with olivine in the uppermost cycle shows an even more complex stratigraphic variation, but the overall trend appears to be one of iron enrichment. Both sets of data need to be augmented by additional phase-chemistry analyses to fully characterize and interpret the stratigraphic trends.

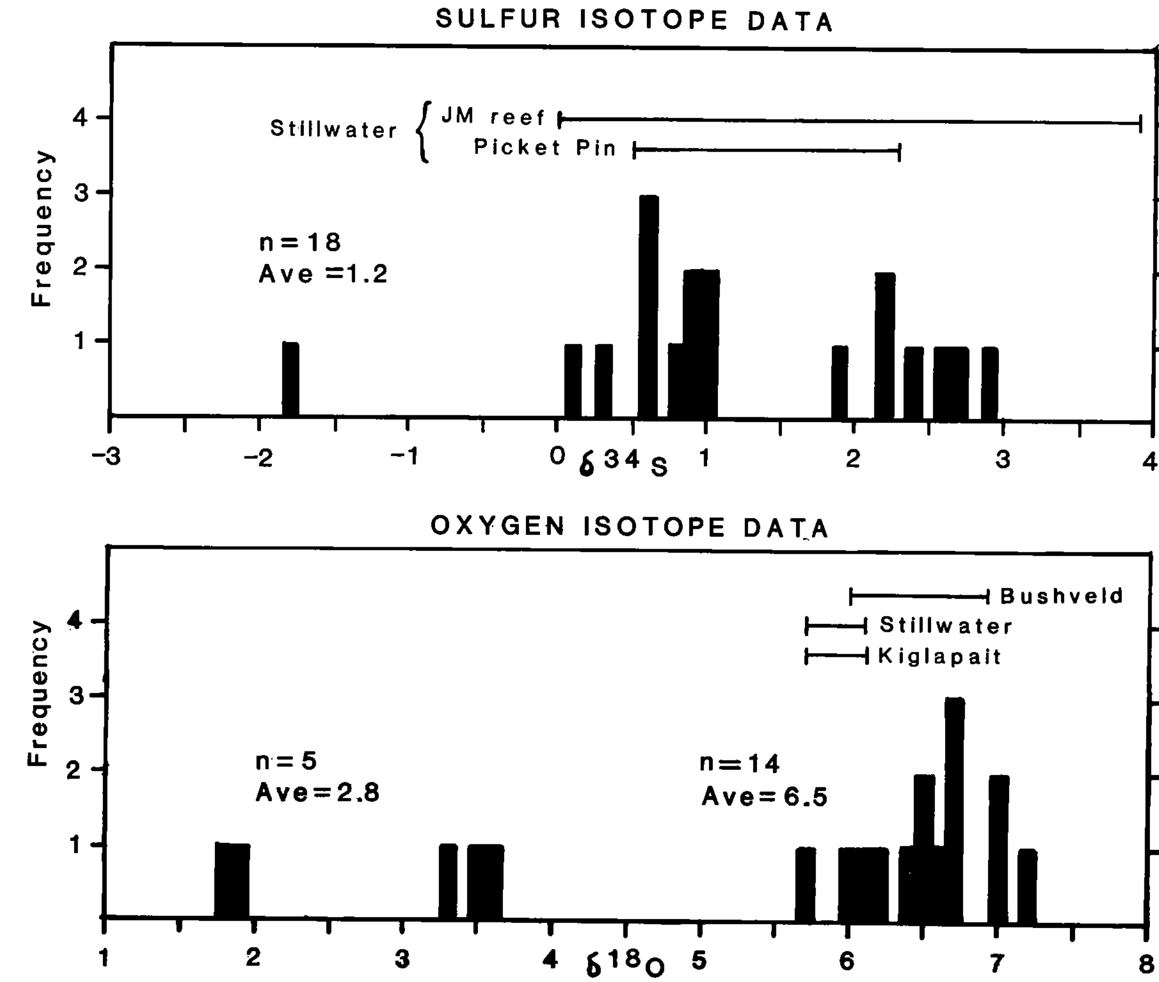

Figure 7. Distribution of sulfur- and oxygen-isotope values for samples of the Glen Mountains Layered Complex.

Platinum-Group-Element Mineralization

Low levels of platinum-group-element mineralization are locally associated with disseminated sulfide mineralization in two principal settings within the GMLC. The two setting are illustrated in Figure 6 and include (1) stratiform sulfide mineralization near the base of orthocumulate textured plagioclase cumulates in the upper part of a cycle, and (2) local sulfide mineralization near and adjacent to thin, discontinuous plagioclase–olivine ± pyroxene cumulates within a dominantly plagioclase cumulate part of a cycle. The common factor in both settings is that the mineralization occurs in the plagioclase cumulates.

Table 1 summarizes the available published anomalous platinum-group-element values for the GMLC (Cooper, 1986), according to sample type and setting of the sulfide mineralization. The whole-rock platinum to platinum + palladium ratios of the sulfide mineralization in the first setting are comparable to ratio values from the JM Reef and Picket Pin occurrences in the Stillwater Complex, Montana (Page and others, 1985). Extremely high whole-rock platinum to platinum + palladium ratio values are associated with sulfide mineralization in the second setting. The significance of the difference in the ratio values between the settings, and its importance, if any, in regard to the mineral potential of the GMLC, remain to be determined. Stream-sediment samples with anomalous platinum and palladium values are also listed in Table 1. The ratio values of platinum to platinum + palladium in these samples are comparable to the whole-rock ratio values and settings.

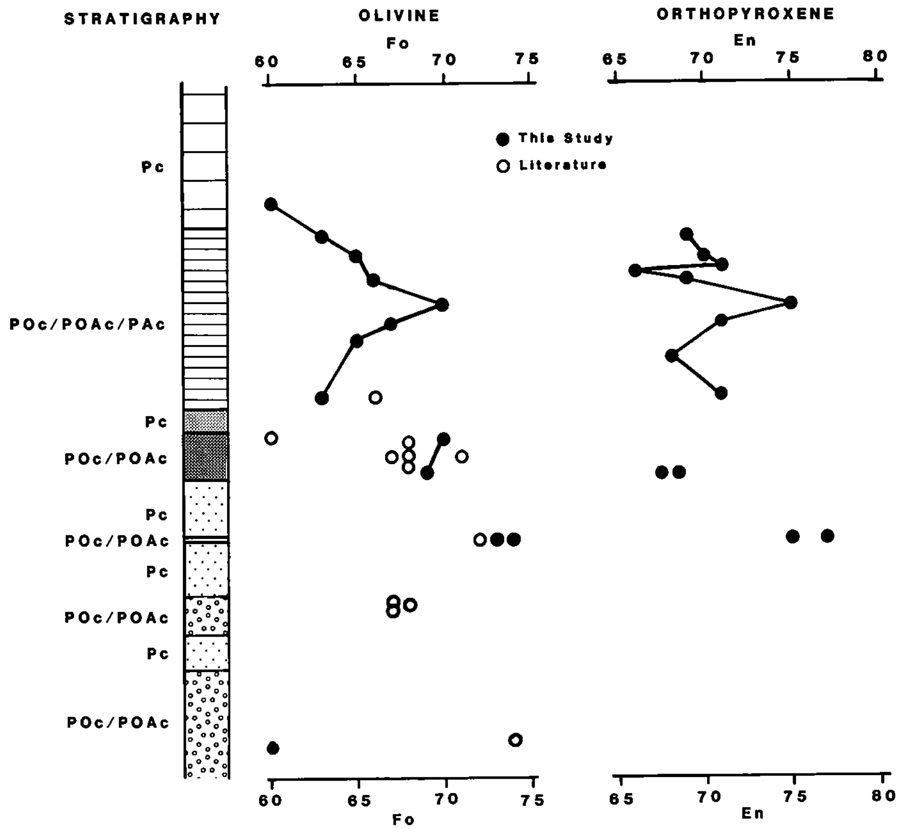

Figure 8. Olivine and orthopyroxene phase chemistry corresponding to the stratigraphic section of the exposed part of the Glen Mountains Layered Complex shown in Figure 5 and the eastern half of Figure 2.

ROOSEVELT GABBROS

The Roosevelt Gabbros include a variety of hydrous (biotite–amphibole), olivine-bearing gabbros scattered throughout the Wichita uplift (Powell, 1986; Cameron and others, 1986). They occur in several distinct settings within the westernmost GMLC. These include (1) large, discrete, cyclindrical intrusions such as the Iron Mountain gabbro intrusion with a well-developed chill margin and internal modal layering; (2) sill-like intrusions as much as 200 ft thick, such as the Glen Creek Gabbro, which may represent offshoots related to larger bodies such as the Iron Creek gabbro; (3) thin, dike-like intrusions <10 ft wide, scattered throughout the study area. Thin, discontinuous wedges also occur locally along contacts between the Cold Springs intrusions and the GMLC. These wedges are interpreted to represent the remnants of small gabbro sills and/or dikes that were preserved when the Cold Springs intrusions were emplaced along essentially the same avenues as the gabbro sills and dikes.

LATE DIKES/SILLS AND COLD SPRINGS INTRUSIONS

Detailed surface mapping and logging of the Placer, Inc. drill core in the study area suggest the following post-Roosevelt Gabbro sequence of igneous events:

1) Emplacement of intermediate (microdiorite) and possibly more-mafic sills and/or dikes throughout the study area.

2) Emplacement of the felsic Cold Springs intrusions as generally sill-like, but also dike-like, bodies at shallow crustal levels along many of the same avenues as the gabbro and intermediate–mafic sills and dikes. One line of evidence for a definite age distinction between the post-gabbro intermediate–mafic intrusions and the Cold Springs intrusions was found in one of the Placer, Inc. drill holes; in this instance an intermediate–mafic intrusion with definite chill margins and diabasic interior is cut by felsic Cold Springs material. A shallow level of emplacement for the Cold Springs intrusions is suggested by the whole-rock

TABLE 1.—PLATINUM AND PALLADIUM ANALYSES OF THE GLEN MOUNTAINS LAYERED COMPLEX ACCORDING TO GEOLOGIC SETTING (DATA FROM COOPER, 1986)

Setting sample type	Pt (ppb)	Pd (ppb)	$\frac{Pt}{Pt+Pd}$
	480	1,400	0.255
Stratiform sulfide in plagioclase cumulate	145	335	0.302
	90	205	0.305
	55	45	0.55
Sulfide associated with POc/POAc layer in plagioclase cumulate	270	5	0.982
	3,325	15	0.995
Stream sediment	260	685	0.275
	255	670	0.276
	390	535	0.422
	60	5	0.938

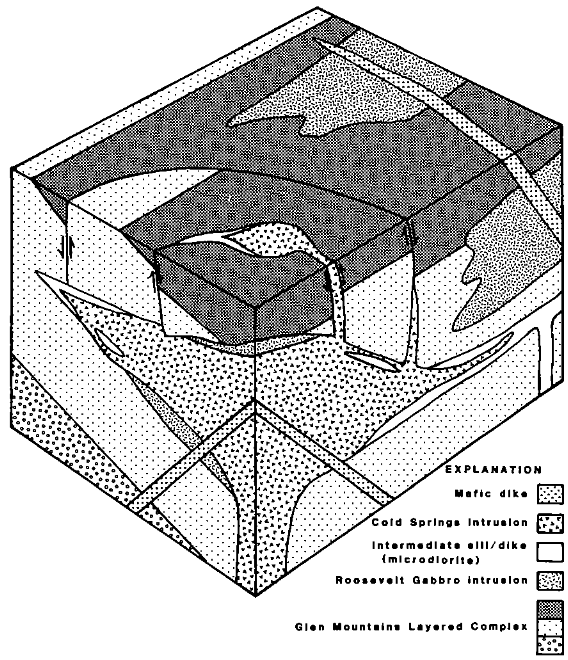

Figure 9. Schematic block diagram illustrating igneous and geologic relationships of post-Glen Mountains Layered Complex igneous rock units.

oxygen-isotope data discussed above. The laccolith- to sill-like Cold Springs intrusions had a doming effect on the overlying GMLC. This effect is readily seen and apparent in the study area (Fig. 2). The GMLC, in the western part of the area, strikes NE and dips NW. In the north-central part of the area, the GMLC strikes W to NW and dips N and NE. The GMLC in the eastern half of the study area generally strikes NW and dips NE. In addition to the overall doming effect, the emplacement of the Cold Springs intrusions produced concentric and radial faults in the overlying GMLC. The more prominent of these faults are shown in Figure 2. In some instances, Cold Springs intrusive material is present along these faults. That the Cold Springs intrusions and the concentric and radial faulting complicate interpretation of the internal stratigraphy and features of the GMLC should be obvious.

3) Emplacement of additional mafic dikes which crosscut the Cold Springs intrusions and all other igneous rock units in the study area.

The entire sequence of post-GMLC igneous events is illustrated and summarized in Figure 9.

SUMMARY

The Glen Mountains Layered Complex is a mantle-derived, anorthositic layered intrusion composed of a sequence of cyclic units, as indicated by detailed field mapping, logging of diamond drill core, sulfur- and oxygen-isotope data, and olivine and orthopyroxene compositional data. The potential of the Glen Mountains Layered Complex to host economic quantitites of platinum-group-element mineralization is difficult to evaluate, because of limited surface exposures and extensive Permian cover, and is further complicated by emplacement of the Roosevelt Gabbros and Cold Springs intrusions. Both types of intrusions have significantly modified and affected the original disposition of the Glen Mountains Layered Complex.

ACKNOWLEDGMENTS

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CAMBRIAN BASEMENT ROCKS AND THE SETTING FOR DEPOSITION OF LATE CAMBRIAN SEDIMENTS IN WESTERN OKLAHOMA

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ABSTRACT.—The southern Oklahoma aulacogen was formed during an episode of Cambrian tectonism and igneous activity that was unique within the Midcontinent. The event is represented directly in the geological record by a bimodal suite of Cambrian igneous rocks which crop out in the Wichita Mountains, and indirectly by Late Cambrian to Devonian strata deposited in a basin formed by thermal subsidence following rifting. The distribution, composition, and interrelationships of igneous rocks are used to help constrain the tectonic state of the southern Midcontinent prior to onset of Paleozoic sedimentation.

Our interpretation hinges on four principal points:

1) The age of the igneous rocks is the age of the rifting. Radioactive dating has yielded ages of 550–525 m.y. for the igneous rocks, based upon earlier U–Th–Pb work on granites/rhyolites in the early 1960s, and later Rb–Sr, Sm–Nd analyses of the gabbros in the 1980s. Note that the Late Cambrian is taken at about 525–505 m.y., so the pre-Reagan unconformity was not of long duration.

2) The igneous units were emplaced in a continental intraplate setting. The geochemistry of the granites is A-type, and basaltic units are transitional between tholeiitic and alkaline in composition, both characteristic of continental igneous activity.

3) Igneous units filled the extending rift, and relief was generally positive. The intrusive igneous rocks are high-temperature, low-pressure types; some granites were surface-breaking; rhyolites were subaerial. With the possible exception of the Meers quartzite, there are no recognized sedimentary units which were coeval with igneous activity.

4) As much as 20 km of extension may have occurred in the crust. The paucity of preserved synorogenic sedimentary rocks prompts the interpretation that thinning and extension of the crust was balanced by the intrusion of igneous rocks. The magnitude of extension is calculated from estimates of the volume of intrusive rocks in the upper crust.

A large spine or fin of heated crust (and lithosphere) existed at the close of the Cambrian. Subsequent cooling of the igneous rocks resulted in thermal subsidence of the rift and surrounding crust, marking the earliest stage of the ancestral Anadarko basin. Maximum subsidence was centered over the rift along a WNW trend.

INTRODUCTION

This paper focuses on the Late Proterozoic–Cambrian framework of southwestern Oklahoma, as this constitutes the basement onto which the Cambrian–Ordovician sedimentary section was built. This region is anomalous for the southern Midcontinent, and probably largely so because of the character of its basement, as first extensively documented by Ham and others (1964). However, much of the structural work directed to the basins (e.g., Anadarko, Ardmore, Hollis) and the uplifts (e.g., Wichita, Criner, Arbuckle) has concentrated on the later Paleozoic history, which appears to have been compressional or transpressional (e.g., McConnell, 1989).

What appears to give this region a special character is the Southern Oklahoma aulacogen, the NW-trending Cambrian rift that cuts into the continental interior from the southern margin of the North American plate (Fig. 1). This rift is important (1) because it represented the emplacement of dense, hot rocks into the continental lithosphere, during an episode of extension with subsequent thermal cooling resulting in the formation of the ancestral Anadarko (= Oklahoma)

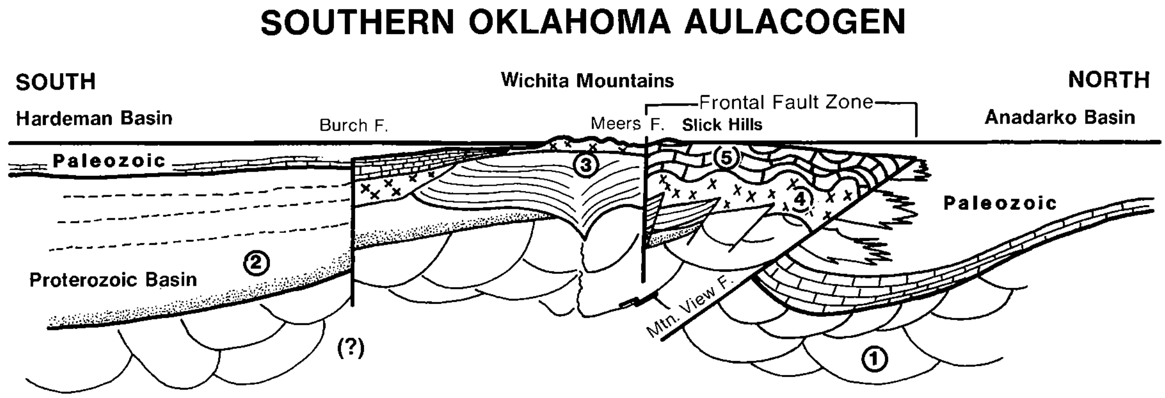

Figure 1. South-north schematic cross section across the Wichita uplift (after Gilbert and Powell, 1988). The uplifted central block, the Wichita uplift, is the floor of the southern Oklahoma aulacogen. This section is drawn through the Lawton segment of the frontal fault zone of McConnell (1989). It is mostly unspecific about the nature of the Pennsylvanian faults, other than that they are generally high-angle (except for the Mountain View fault, which clearly dips 30–40°) and compressional. In detail, the nature of the faults changes along strike. The emphasis of the diagram is the distribution of petrologic terranes: (1) = granite–rhyolite terrane of 1.35–1.40 b.y., plating the southern Midcontinent basement (Denison and others, 1984); (2) = Proterozoic basin of Brewer (1982) and Brewer and others (1981,1983); (3) = Cambrian gabbroic–anorthositic floor of the SOA with its smaller gabbroic plutons (Powell and others, 1980); (4) = Cambrian silicic cap of sheet granites and rhyolite/tuffs; (5) = lower Paleozoic carbonate sequences.

basin (Johnson, 1989; Perry, 1989; Gilbert, in press), and (2) because the structural heterogeneities which the rift and its igneous filling imparted to the crust influenced the location of the late Paleozoic basins and uplifts. Consequently, we will concentrate here on the formation of the aulacogen and outline what is known of the structural environment that existed prior to deposition of the Timbered Hills and Arbuckle Groups, the basal Cambrian–Ordovician sequence in Oklahoma.

CAMBRIAN RIFTING AND FORMATION OF THE SOUTHERN OKLAHOMA AULACOGEN

Definition of the Southern Oklahoma Aulacogen

We define the southern Oklahoma aulacogen (SOA) as the rift of apparent Cambrian age which now trends WNW from about the Dallas area to the Amarillo area. Following Perry (1989), it seems best to restrict the usage of SOA to the extensional regime of the rift and its structures and rocks, rather than to include subsequent Paleozoic structural developments. This means, for example, that the present Anadarko basin is not the same structure as the SOA, or directly a part of the SOA. Obviously, the later Paleozoic structures dominate the tectonic fabric of the region, and inferences about the rifting must be made by looking through a screen of later events. While a strong case can be made for the influence of the SOA on the formation of later basins and uplifts and upon sedimentation patterns, it is helpful to keep the definition of the SOA focused on the rifting stage and its concomitant lithospheric effects.

We know that the SOA existed from the work of Ham and others (1964), and from subsequent refinements, as exemplified by Powell and others (1980) on lithostratigraphy, Keller and others (1983) and Coffman and others (1986) on geophysical interpretations, Denison and others (1984) on basement relations in the Midcontinent and Texas, and Lambert and others (1988) on constraining the age of the SOA to early to middle Cambrian. McConnell and Gilbert (1986,1990) used this previous work to generate a more detailed interpretation of what went on during this tectonic episode. They concluded that there was a close tie between magmatism and tectonism during rifting.

The present extent of the SOA is recognized in five ways: (1) the profound geophysical signature, outlined by prominent gravity (e.g., Lyons and O'Hara, 1982) and magnetic (e.g., Zietz, 1982) anomalies; (2) the distribution of Cambrian basement (Ham and others, 1964; Lambert and others, 1988); (3) the distribution of mafic basement rocks (Ham and others, 1964; Powell and others, 1980);

(4) the distribution of faults and uplift structure (Ham and others, 1964; Brewer, 1982; McConnell, 1989); and (5) the thickening of Cambrian–Ordovician sedimentary units toward the uplifted block. All these features are centered around the Wichita–Amarillo–Criner uplift. Consequently, the present extent of the SOA is essentially coincident with this uplifted block. We interpret this to indicate that the Pennsylvanian compressional to transpressional fault system bounding the uplift is in part a reactivated Cambrian extensional system which defines the major boundaries of the rifting. It further means that the mafic basement of Cambrian age was the principal infilling of the rift as extension proceeded (Gilbert, 1983; McConnell and Gilbert, 1986). Related silicic volcanics appear to have overlapped the boundaries of the rift (Ham and others, 1964), and thus their extremeties do not directly outline the major extensional zones of the SOA.

Figure 2 is a geologic map of southwestern Oklahoma with the Wichita uplift outlined by the subsurface fault system. The igneous floor of the SOA crops out in the Wichita Mountains sufficiently to allow meaningful observations on the relation of magmatism to tectonism. The positions of the deeper parts of the Anadarko basin are marked approximately by the distribution of some of the Permian units, the deepest part lying at the northern margin of the frontal fault zone. A change of structural orientation (the Lawton bight) along the uplifted block is noticeable. This appears to be due to Pennsylvanian overthrusting in the Lawton segment of McConnell (1989). Coffman and others (1986) did a detailed 2½-D gravity model along the line A–A′ shown in Figure 2. There is a pronounced gravity bulge in the profile across the frontal fault zone that can be explained by the presence and character of the overthrust block.

Figure 3 is a basement map of the Wichita uplift taken from Coffman and others (1986), but based on Ham and others (1964). The strip of Raggedy Mountains Gabbro Group along the northern margin of the uplift resulted from block rotation in the Pennsylvanian, such that more erosion occurred on the north, stripping off the Carlton Rhyolite and Wichita Granites. This exposed the main fill in the rift, namely, the gabbro group. The gravity signature indicates that the silicic units are only a thin covering toward the south, and that the gabbro extends to the southern boundary of the uplift. The thickest sections of rhyolite are found in less-uplifted parts on the north, within the frontal fault zone. Consequently, McConnell and Gilbert (1990) argue that this suggests an original half-graben configuration for the rift, the most Cambrian extension having taken place on the northern margin of the aulacogen.

Characterizing the Cambrian Extensional Environment

This rift was filled primarily with igneous rocks, and not sediments, as it "opened." Okaya and Thompson (1986) presented arguments to show that rifting often may include intrusion of material of crustal density from the mantle. While their analysis was centered around the deeper, more ductile parts of the crust in the Basin and Range province, we find it also applicable to higher crustal levels. Except for possibly the Meers Quartzite, a very minor stratigraphic unit, no sedimentary units coeval with the igneous fill are known. All evidence points to dominantly subaerial erosional processes occurring periodically throughout the accumulation of the igneous pile. Thus, episodic uplift must have continually occurred. This is particularly clear in the case of the Carlton Rhyolites. White and McKenzie (1989) noted that widespread igneous activity is characteristic of rifting. Their calculations show that where the mantle is significantly hotter than normal, as expected under rifts, crustal stretching can actually lead to uplift. We believe that the SOA behaved in this manner.

The data available to analyze extension are of two types: structural and petrological. Figure 4 illustrates the spatial arrangement of the major rock units and their layering, but without specification as to nature and placement of faults. What can be documented in the field is that attitude discordance exists between and among successively younger igneous units (e.g., McConnell and Gilbert, 1990). Older units, while not always more steeply inclined, almost always have different attitudes than the intruded (younger) unit. Fortunately, the two major intrusive units, the Glen Mountains Layered Complex (GMLC) and the Roosevelt Gabbro (RG), each have distinctive primary igneous fabrics primarily due to crystal-settling processes. These processes typically result in subhorizontal layering when originally formed. The GMLC is particularly well layered on several scales, from plagioclase lamination, well displayed in thin section and at hand specimen scale, through modal layers of different mineralogy (such as olivine + plagioclase, olivine + clinopyroxene + plagioclase, plagioclase) at the outcrop tens-of-meters scale, to gross cyclic units at the map scale (Powell and others, 1980; Cooper, this volume). A reasonable interpretation of these discordances is that normal faulting accomplished extension of the rift as igneous intrusions went on, successively rotating small blocks within the igneous floor.

The contact between the Wichita Granites and the substrate gabbros is both an unconformity and an intrusive contact. The truncated nature of both

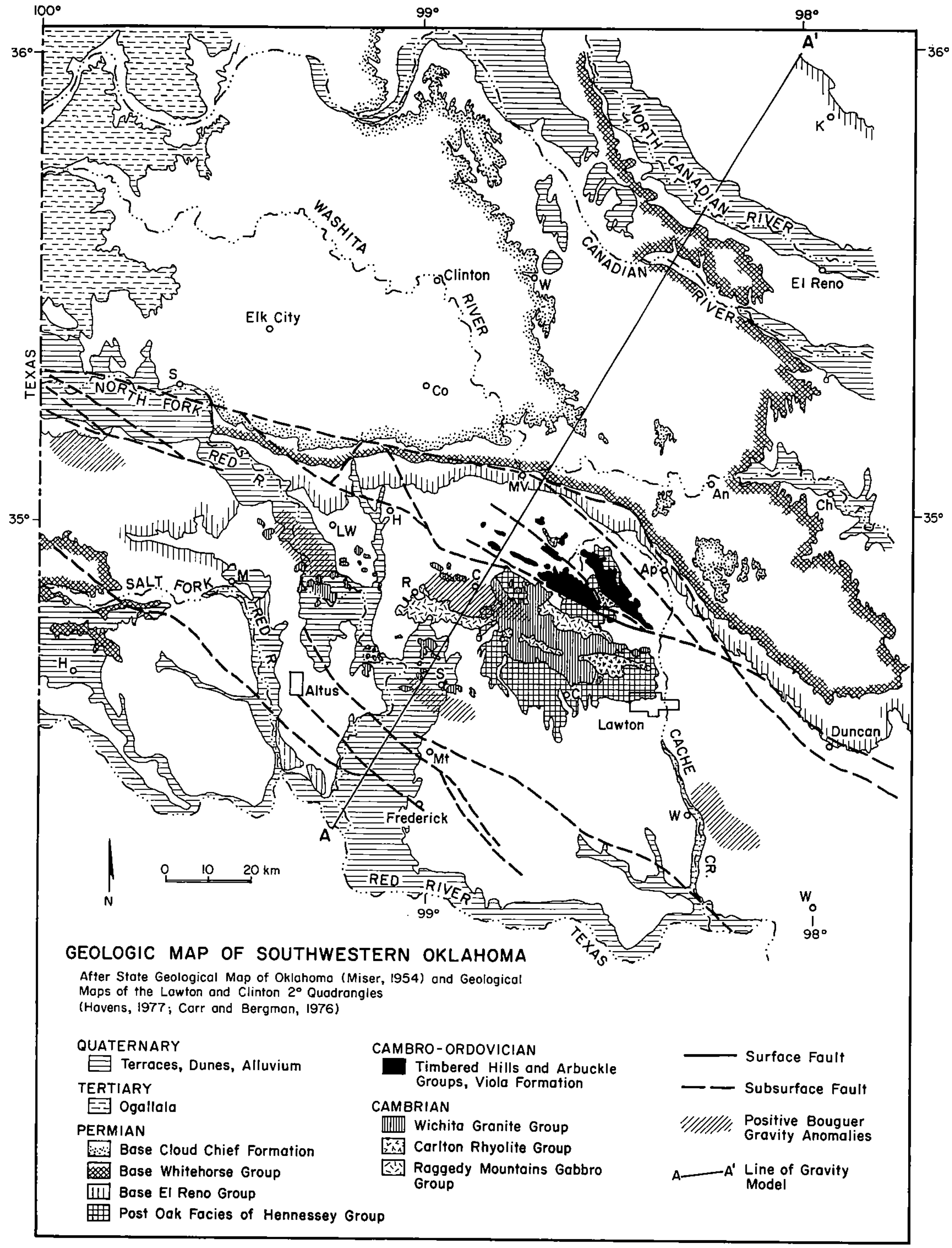

Figure 2. Map of southwestern Oklahoma from Coffman and others (1986). The subsurface Pennsylvanian faults bounding the Wichita uplift are taken to be the approximate outline of the SOA and the zones of reactivated Cambrian faults. The outcrop areas of Cambrian igneous rocks represent a small part of the material which filled in the rift as extension progressed. The Lawton segment of the frontal fault zone (McConnell, 1989) is clearly evident, as is the Lawton bight. The orientation of the structures changes, from 280–290° in the west to ~300° in the southeast; this structural change is well marked in the Permian outcrops.

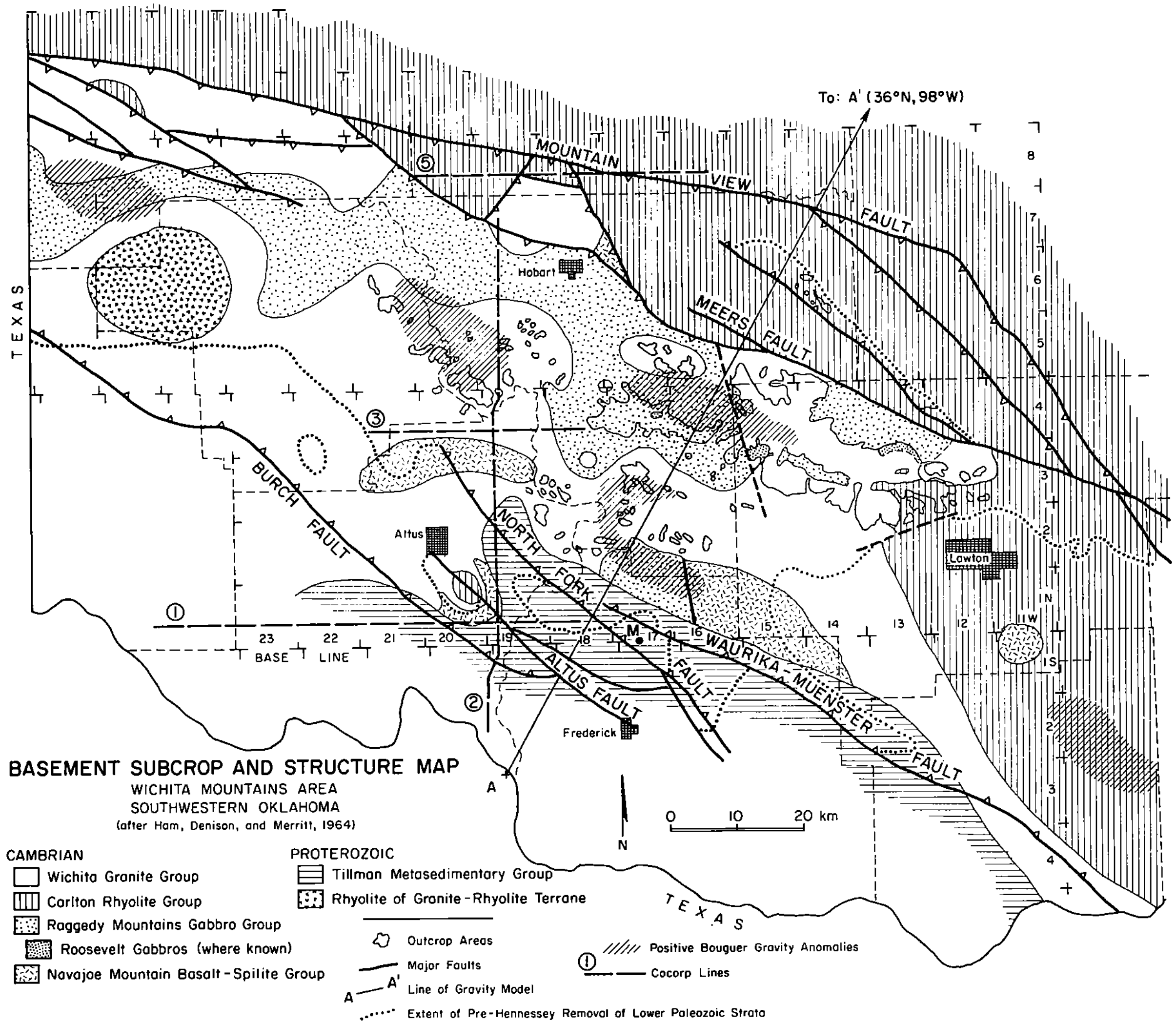

Figure 3. Map of basement and structure taken from Coffman and others (1986). Presumably the gabbroic substrate extends under the thin sheet granites of the Wichita Granite Group from the northern side, as portrayed in the structural and gravity model of Coffman and others (1986). The width of the uplift varies from 45–50 km to nearly 85 km. Some of the width is clearly due to Pennsylvanian overthrusting. We have used an initial estimate of 50 km for the size of the main part of the rifted zone of the SOA.

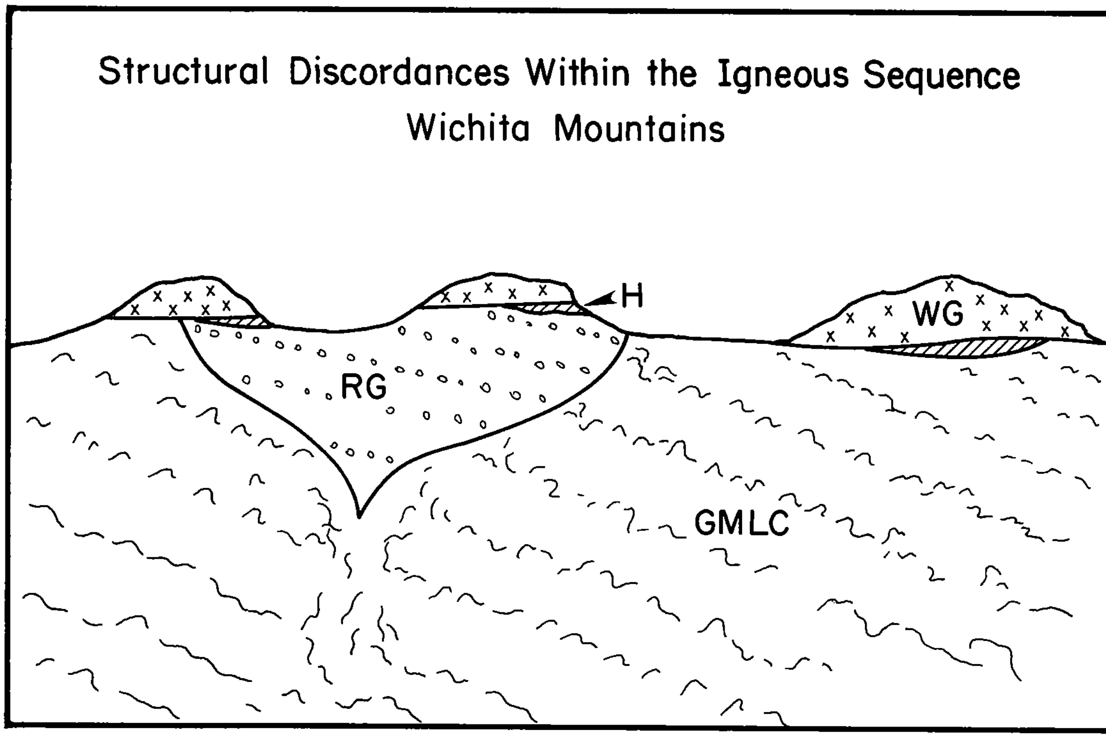

Figure 4. Diagrammatic view of the discordances between the Glen Mountain Layered Complex (GMLC), the younger intrusive Roosevelt Gabbros (RG), and the still younger Wichita Granites (WG). H is a set of hybrid rocks which developed along the gabbro/granite contact. The GMLC and the RG each have internal primary igneous layering, which is interpreted to have developed subhorizontally during the formation of each of the units, thus providing structural markers. The granite/gabbro contact is first an erosional unconformity and then later an intrusive bottom contact for the sheet granites, which contact is also taken, on a region scale, to be subhorizontal and a structural marker.

the GMLC and the RG, and the fact that a single mapped granite overlies both in one area, are strong arguments for the existence of an unconformity. Consistent also with this is the occurrence of hybrid igneous rocks along the gabbro/granite contact, which appear to have been regolith on the gabbro, melted and mobilized locally by the intrusion of the overlying granite. The relief on this unconformity appears to have been relatively small. Thus, on the regional scale of Figure 3, this granite/gabbro contact was nearly horizontal, and its present attitude can also be used as a marker to recognize later deformation.

The radiometric ages of all the exposed igneous units are now known to be Cambrian (Lambert and others, 1988), approximately in the range 550–525 m.y. Although some of the units have not been dated with high precision, it is clear that the biomodal suite relates overall to a single geologic event. That event is taken to be the formation of the SOA. The erosional surfaces and erosional effects inferred from the geology and petrology are part of the active rifting history of the SOA.

The shapes of the granites are consistent with emplacement in an extensional regime. None has a discernible gravity signature that might indicate a substantial depth component. Thicknesses of granite sills as determined from a few wells into basement appear to be <500 m (Ham and others, 1964). Individual granites extend laterally for tens of kilometers and suggest that the form of the granites can be described as sheet-like. Such forms are to be expected in rifts, as opposed to more-bulbous shapes typically found in compressional zones, such as the Sierra Nevada.

Petrologic data all are consistent with shallow emplacement for the granites. Granophyric texture is common. Such a texture can represent rapid quenching, which would occur in a near-surface environment. The granites have A-type chemistry, typical of anorogenic settings (Myers and others, 1981; Gilbert, 1983). Rhyolite dikes which cut some of the granites indicate that erosion due to active uplift and/or block rotation due to normal faulting in the rift had gone on since the original emplacement of the granite. The overburden had to be decreasing with time; otherwise, the rhyolite dikes would exhibit granitic texture.

Finally, we suggest that the strong plagioclase lamination common in some horizons of the GMLC may be due to paleoseismic events associated with extensional faulting. Plagioclase layering is to be expected in layered mafic bodies like the GMLC through either float or sink mechanisms in the magma chamber. However, the extreme compaction and exceptional lamination of some zones probably require some additional physical process, and seismic shaking is one reasonable possibility.

Estimating the Amount of Extension

Several approaches are available to estimating the amount of extension in a rifting environment, including (1) analysis of fault geometry (Wernicke and Burchfiel, 1982), (2) determination of the amount of crustal thinning (McKenzie, 1978, 1984), and (3) mass-balance calculations (Cordell, 1982). Characterization of fault geometry is difficult because distinctive marker horizons have not yet been indentified in the rhyolites, granites, or gabbros (except for the granite/gabbro contact), and because Pennsylvanian tectonism has resulted in structural overprinting. Furthermore, magmatic additions to the rift can take up variable amounts of extension and will also affect the depth of the brittle–ductile transition (Tullis and Yund, 1977) through heating of the extended zone. The position of this transition in the crust can substantially affect depth and style of faulting. McConnell and Gilbert (1990) noted that an essentially simple shear model could be used to explain the stretching of the brittle upper crust. This could lead to a relatively low estimate of macroscopic extensional strain, <10% (5 km), with no allowance for magmatic additions.

McConnell and Gilbert (1986) also used Cordell's (1982) method of mass-budget calculations to estimate extension. This method requires knowledge of the amount of igneous material added to the rift throughout the lithospheric column. Estimates for these amounts were available from Ham and others (1964), Gilbert (1983), and Coffman and others (1986). This method yielded estimates of ~40% strain (17–21 km). Hildenbrand (1985) estimated comparable amounts for the Reelfoot rift. These two estimates are the only ones available so far, and probably represent the range of possible valves until substantial new data are at hand. It is not likely that the Cambrian faulting style can be more nearly specified; however, improved mass estimates can probably be made when 3-D gravity models are employed, and as new deep seismic data for this region are made available (e.g., Chang and others, 1989).

CONSEQUENCES

Figure 5 summarizes the upper crust development (down to ~10 km) during the time of rifting in the mid-Cambrian. The SOA is shown cutting across the northern edge of a deep Proterozoic basin (Brewer, 1982); that basin now underlies the Hollis–Hardeman basin. Rifting started with intrusion of basaltic liquid at shallow crustal depths, which resulted, upon crystallization, in the Glen Mountains Layered Complex. Some extrusion also occurred to form the Navajoe Mountain Basalt–Spilite Group. This latter unit is not known to be volumetrically significant. Continued extension resulted in faulting and rotation of blocks of the

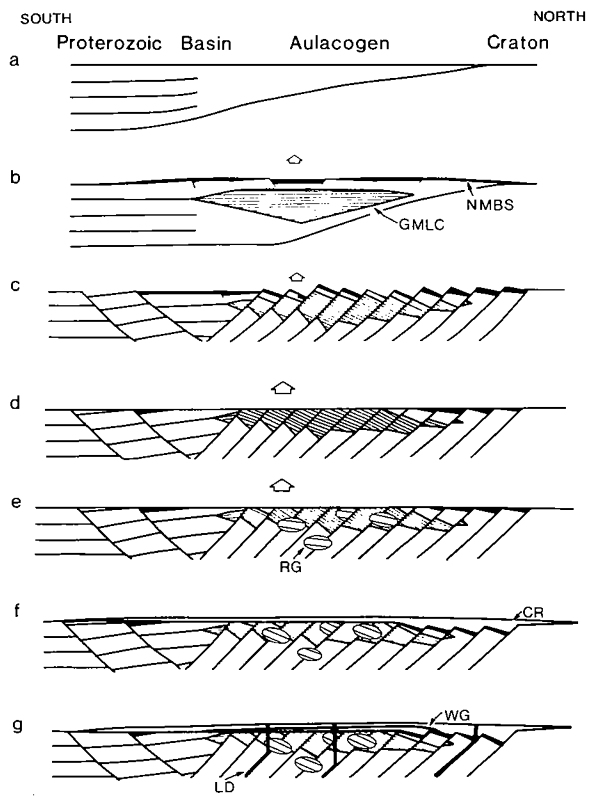

Figure 5. Schematic showing the relation between magmatism and tectonism from McConnell and Gilbert (1986,1990). NMBS = Navajoe Mountain Basalt–Spilite Group (Ham and others, 1964), which is taken to be the extrusive equivalent of the GMLC (Gilbert, 1983); CR = Carlton Rhyolite (Ham and others, 1964), which overlaps the structural margin of the rift; and LD = Late Diabase (Ham and others, 1964; Powell and others, 1980; Gilbert, 1983); other abbreviations as in Figure 4. Progressive rotation due to normal faulting is shown for the GMLC, RG, and WG/CR. Pathways for extrusion of the CR are taken to be the normal faults. LD may also in part have used these faults as conduits.

GMLC. Continued uplift of this extending terrane occurred, so that erosion kept removing overlying material and the RG plutons were also, in turn, rotated by continued faulting. Erosion was sufficient to have laid bare both gabbroic units when silicic liquids arrived at the surface, first in the form of the Carlton Rhyolites, and then shortly thereafter and coevally in the form of the Wichita Granites. Late diabase dikes and plugs cut all other igneous units.

Ultimately, the rift with its fill of igneous rocks near the surface, and with additional coeval igneous material emplaced at mid-crustal depths (Coffman and others, 1986; McConnell and Gilbert 1990), represented a reverse keel of hot, mantle-derived material transecting the crust. By the Late Cambrian, igneous activity had substantially diminished. The rifting process apparently was over, and the crustal column started to cool. The inverted keel and its lithospheric root began to subside; this subsidence affected a broad surrounding region and led to the ancestral Anadarko (or Oklahoma) basin, as shown by thickened Cambrian–Ordovician units over the keel. More important, a profound structural heterogeneity had been imparted to the crust; this heterogeneity influenced the tectonic evolution of the southern Midcontinent during the Ouachita collision.

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A NEW ISLAND IN THE SOUTHERN OKLAHOMA ARCHIPELAGO

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ABSTRACT.—The unconformity between the Carlton Rhyolite and overlying Cambrian sediments is well exposed in the Slick Hills of southwestern Oklahoma. This unconformity has previously been interpreted by R. N. D. as a hilly landscape, with relief of ~100 m, that was transgressed slowly during the late Cambrian (Franconian) to form a group of islands—the southern Oklahoma archipelago. Here we report the discovery of a new site which more than doubles the estimate of the relief of the Franconian land surface and displays an unconformable contact high in the Fort Sill Formation (basal division of the Arbuckle Group). Perhaps the most interesting aspect of the discovery is that facies developed immediately adjacent to the unconformity are identical to those found at similar sites several hundred feet below and bear little resemblance to those seen in the enveloping Fort Sill; i.e., the islands of the southern Oklahoma archipelago were characterized by a distinctive and diachronous shoreline facies. Important elements of this shoreline facies include mechanical bafflestones, glauconitic hardgrounds, pelmatozoan-rich grainstones, and ferruginous stromatolites.

INTRODUCTION

In a recent review of Paleozoic rocks of the Midcontinent, Denison (in Johnson and others, 1988) noted the diachronous character of the unconformity between the igneous/metamorphic craton and an overlying Upper Cambrian/Lower Ordovician sedimentary sequence that records a major marine transgression. In general terms, this transgression began in the Late Cambrian (Franconian) in southern Oklahoma, but it is clear that the transgressed surface had considerable relief in certain areas, both within the southern Oklahoma aulacogen and on the adjacent craton. By way of illustration of this point, subsurface exploration has yielded clear evidence that a range of basement hills in the Tulsa, Oklahoma, area remained sediment free until after the deposition of the Arbuckle Group (as much as 2,300 m thick in the aulacogen axis). Similarly, Ham and others (1964) recognized hills of Precambrian granite as much as 200 m high in the Tishomingo area of southern Oklahoma (outside the original aulacogen area).

The finest exposures of this land surface are those found in the eastern Slick Hills of southwestern Oklahoma, where a substantial relief of hills as high as 100 m, with slopes up to 10°, has been described (Donovan, 1986; Donovan and others, 1986). In this area, which is within the area of the original aulacogen, the basement rock is the Carlton Rhyolite.

One of the interesting conclusions to be drawn from this overlapping geometry is that the aulacogen area (as defined by the Carlton Rhyolite) underwent rather rapid cooling and subsidence prior to the transgression (given the available radiometric dates, there is at the most 10 million years between the end of igneous activity and initial sedimentation). In other words, there is no evidence that the Carlton Rhyolite as a whole was elevated relative to the adjacent craton at the time of transgression.

The geometry of Cambrian deposits above the Franconian land surface is remarkably complex, involving an initial siliciclastic phase (the Reagan Formation) and a subsequent period of carbonate grainstone accumulation (the Honey Creek Formation), following which the area gradually evolved into a full-blown carbonate platform (as recorded by the Fort Sill Formation, lowest unit in the Arbuckle Group). The sediments throughout are rich in iron, containing such unusual features as hardpan hematite cements, ferruginous ooids, abundant glauconite, and localized replacement of calcite by ankerite (Cloyd and others, 1986). The oldest deposits have been interpreted as alluvium, whereas most of the topmost Reagan and the entire Honey Creek appear to have been constructed in a tide-dominated, shallow-marine environment. Thus, with time, the land surface was inundated; hills became islands, and an archipelago (herein referred to as the southern Oklahoma archipelago—new name) evolved.

Donovan (1986) has suggested that the initiation of carbonate production was brought about through intensive colonization by pelmatozoans of the clear-water sites around islands in the archipelago and the rocky shorelines of the adjacent craton. Upon death, these pelmatozoans fragmented and produced great amounts of coarse carbonate sand which was slowly worked around and between islands. If this is so, it follows that the passage from Reagan to Honeycreek is diachronous.

While most of the Honey Creek comprises coarse carbonate grains, at certain sites an exceptional facies is found adjacent to the unconformity. This facies is characterized by carbonate-mud accumulations, comprising small brachiopod-dominated buildups, ferruginous stromatolites, and highly fossiliferous mud banks (Donovan and others, 1986). These authors suggested that this mud-dominated facies was deposited in sheltered sites to the leeward of islands. Conversely, in their model, exposed windward shorelines are characterized by accumulations of very coarse pelmatozoan-rich grainstones.

In this short paper we give an initial description of a newly discovered island margin site which indicates that the Cambrian land surface within the aulacogen was not inundated until most of the Fort Sill Formation had been deposited (Fig. 1). This site is located at the southeastern extremity of the Slick Hills in the Richard's Spur area (Comanche County, sec. 27, T. 4 N., R. 12 W.).

DESCRIPTION OF SITE

The Carlton Rhyolite exposure occupies a very small area of ~100 m². To the west the rhyolite is covered by Permian Post Oak Conglomerate, whereas to the east it passes beneath limestones of the Fort Sill Formation that dip NE at 50–60°. Distinctive island-tied facies in the Fort Sill can be found as much as 5 m on either side of, and as much as 2 m above the rhyolite. As the facies occurs above the present level of exposure of the rhyolite, it follows that the exposure appears to be that of a marginal site, rather than the tip of the island. The facies passes laterally and upward into limestones "typical" of the upper part of the Fort Sill Formation. These consist of thick beds of lime mudstone and wackestone that are interbedded with algal boundstones. Stromatolites associated with the latter are distinctively squat, with heads averaging 11 cm in diameter. Limestones of this general type persist at this horizon throughout the Slick Hills (in the northwest they have been dolomitized).

THE SHORELINE FACIES

In general, the shoreline facies is characterized by extraordinary diversity and richness of fossil content (this is in stark contrast to the general monotony of the "typical" Fort Sill). Individual beds are rarely >2.5 cm thick, hardgrounds with relief are common, and there is abundant evidence of reworking in the section. As a result individual beds can rarely be traced for >1 m. The most important subfacies are:

1) Rhyolite breccia.—This subfacies comprises fragments of devitrified rhyolite that are mixed with a variety of broken fossils and intraclasts, and cemented by drusy sparite. It is located immediately adjacent to the Carlton Rhyolite.

2) Pelmatozoan-rich grainstones.—These are clastic deposits in which ~80% of the grains are pelmatozoan fragments; thick-shelled trilobites, broken brachiopods, lime mud intraclasts, quartz silt, and glauconite peloids comprise the remainder of the grains. The dominating cement is syntaxial overgrowths of calcite.

3) Pelmatozoan-rich packstones and wackestones.—Detrital components in rocks of this type are found in approximately the same proportions as in the grainstones. However, the lithology is

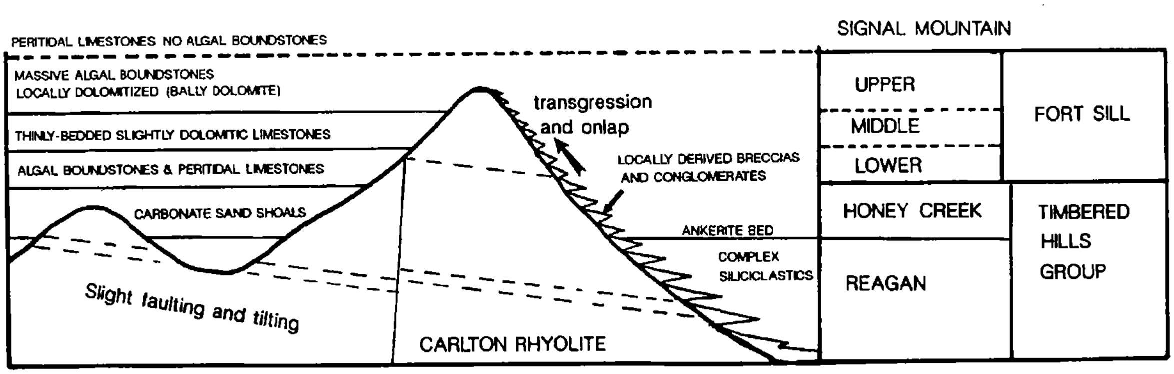

Figure 1. The stratigraphic relationships seen at the base of the lower Paleozoic section in the Slick Hills.

more poorly sorted and in some cases the pelmatozoans are less fragmented.

4) Fossiliferous mudstones.—This subfacies is the most variable of the six recognized here. Fossils are poorly sorted and have suffered little abrasion; they include thin-shelled trilobites, a variety of brachiopods (especially orthids), various pelmatozoans (some entire), calcareous sponges, and algae (cf. *Renalcis*) (Fig. 2). Intraclasts of lime mud are abundant in some beds. The facies exhibits numerous hardgrounds which show relief of as much as 2.5 cm and may have a veneer of glauconite on their surfaces.

5) Mechanical bafflestones.—Deposits of this kind constitute an unusual kind of carbonate buildup in which the mechanical stacking of brachiopod valves in vertical or near-vertical positions by gentle wave action (Donovan and others, 1986) has resulted in a fabric that traps mud and small clastic particles (including quartz silt and glauconite). Bafflestones of this type formed small buildups that were stable enough to support the growth of pelmatozoans.

6) Ferruginous stromatolites.—This subfacies is visually dramatic by virtue of its color—a vivid red. Stromatolites of this type consist of very fine-grained hematite arranged in thin laminae of great delicacy. Typically these stromatolites have formed on a substrate of lime mud (often a hardground); less commonly they have formed on or around a fossil fragment (Fig. 3). The initial deposits are generally a flat mat no more than 1 cm thick (this mat may mimic the shape of the underlying fossil). In the thicker mats the flat-mat stage is followed by differentiation into hemispheroids which may or may not show lateral linkage. The maximum thickness of mat is ~4 cm; the stromatolites are generally overlain by clastic carbonates (mudstones, packstones, etc). The ferruginous mats evidently were stable entities, for they formed the substrate on which large numbers of pelmatozoans grew.

DISCUSSION

Applying the model proposed by Donovan and others (1986), we suggest that the entire shoreline facies was deposited in a low-energy coastal setting adjacent to a small island composed of Carlton Rhyolite. Because of the small area of exposure it is not possible to determine the precise geography of the island. However, it is clear from the thinly bedded character of the sediments that they do not resemble the "exposed coastline" facies of Donovan and others (1986). It seems most likely that the setting was a sheltered one, perhaps a bay on the lee side of an island.

Within this context, the rhyolite breccia (subfacies 1) is interpreted as a veneer of beach gravel; the pelmatozoan grainstones (subfacies 2) are seen as small wave-washed sand shoals, while the various packstones and wackestones (subfacies 3) were deposited either below wave base or in greatly sheltered sites with negligible wave action. Similarly, we suggest that the various types of mudstone (subfacies 4) record sedimentation in very quiet settings where various delicate organisms could exist. In this environment the dominant sedimentary process was the deposition of large amounts of lime mud. Whenever mud production ceased, hardground formation ensued. Because the Franconian seas were more iron-rich than those of today, many of the hardgrounds developed a veneer of glauconite. The presence of

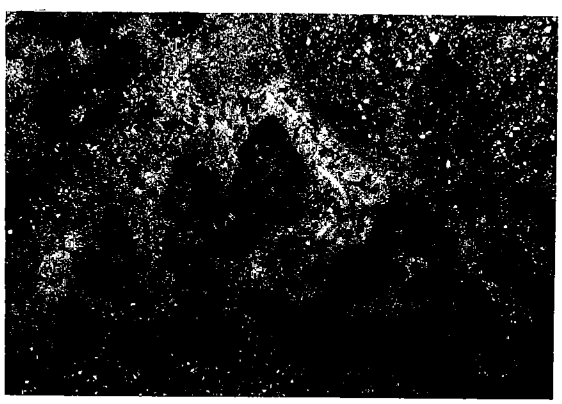

Figure 2. Lime mudstones containing algae cf. *Renalcis*. The latter are surrounded by a crust of fibrous calcite cement. Cross-polarized light; field of view is 4 mm wide.

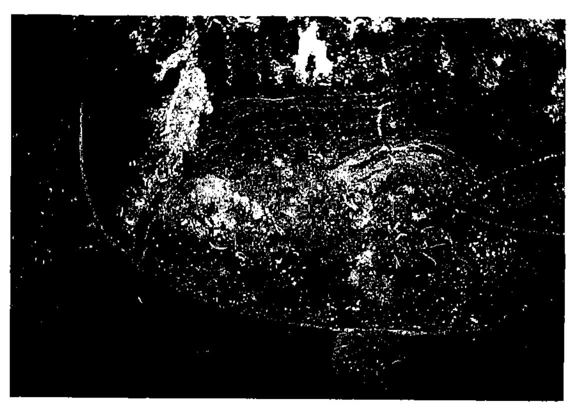

Figure 3. Thin-shelled brachiopod valve that is partly filled with fossiliferous mudstone and has then served as a site for ferruginous mat growth. Cross-polarized light; field of view is 4 mm wide.

some ripped-up mudstone clasts suggests that this environment was disturbed by storms from time to time.

The mechanical bafflestones (subfacies 5) are interpreted as shell banks produced by waves of moderate energy (Sanderson and Donovan, 1974). These formed stable accumulations that trapped sediment and acted as a substrate both for the growth of pelmatozoans and also for the growth of iron-fixing algae/cyanobacteria (subfacies 6). The ferruginous stromatolites form the most conspicuous element of an unusual facies that is spatially tied to the rhyolitic basement (and hence is diachronous) and is characterized by generally low-energy motifs, an abundance of iron (derived from weathering of the rhyolite), and a distinctive fauna that includes pelmatozoans, brachiopods, trilobites, sponges, and algae.

ACKNOWLEDGMENTS

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THE ENIGMA OF TECTONIC DOLOMITIZATION AND FRACTURING IN ARBUCKLE EXPLORATION

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ABSTRACT.—The "anticlinal theory" was advanced in 1861. Currently, this is the concept followed by most explorationists. Structural closure and high structural position is still considered the "in thing." The majority of petroleum geologists are unaware of the "break theory" of Evans (1866) in connection with oil production in the Lima field, Indiana, Trenton–Black River dolomites.

Previous studies have shown that fractures are localized on structures in areas of maximum stress formed by tectonic movements. Fracture patterns are also controlled by bed thickness, brittleness, and stratigraphic variation. Further investigations have shown that tectonic dolomitization may occur in the fractured area and may conform to iso-fracture patterns, or second-derivative interpretations. This concept of the origin of tectonic dolomite is complicated by the formation of dolomite by depositional processes and also by metasomatic tectonic processes. "Minerals and suites of minerals occur in their environment of stability."

Geologists and engineers should examine well-cuttings at the well-site and study drilling characteristics, in order to properly explore for tectonic dolomite in fractured reservoirs. Fractured reservoirs in tectonic dolomite are characterized by (1) large vertical and stratigraphic differences in producing zones in offset wells, (2) variable oil and gas characteristics in adjacent locations, (3) lost circulation, (4) blowouts, and (5) variable bottom-hole pressures. Good potential producing zones have been ruined when standard operating procedures were followed. Many major oil fields have been discovered due to blowouts, unfortunately, and then later they were developed by modification of drilling and completion techniques.

A fracture or metasomatic channel 1 ft in vertical extent and 1/16 in. wide may produce 100,000 barrels of oil per day if carefully completed in the relevant structural and stratigraphic setting. Privileged information indicates that many of the high-volume wells worldwide produce from only 5 to 6 ft of vertical section. How many fields have been missed due to lack of experience?

Electric logs are excellent tools, but are not a panacea. Good wells have been and are being produced from rocks with log porosities <6%. Ankeritic dolomite, sulfide veins, water loss, and many other specific conditions can only be overcome by direct studies of samples.

INTRODUCTION

Much of my early work on fractured reservoirs was done for a thesis supported by Stanolind Oil and Gas Co. in 1950–51, held confidential for nine years, and published by the AAPG in December 1960. In research, "Chance only helps those whose minds are well prepared for it" (R. Taton).

During 1954–59, I recognized the relationship between dolomitization and fracture localization while preparing an iso-fracture map of the Goose Egg Dome structure in Wyoming and studying the dolomite production of the Deep River field in Michigan. Worldwide consulting, from 1959 to present, in Mexico, Canada, Libya, Turkey, Iran, Brazil, China, and the United States, has helped me to refine my concepts. The Sanish field was discovered in 1954 from a fractured euxinic stratigraphic-structural complex. The first producing well, completed in tectonic dolomite, was drilled on purpose by Stanolind Oil and Gas Co. in 1957. Based on some of these initial concepts, the importance of synclinal negative-departure fractures and associated tectonic dolomite was emphasized by study of the Scipio-Albion trend from 1960 to 1968. This dilation hypothesis is reinforced by the synclinal positions of many of the major fracture producing fields of the world, including the shallow Florence field of Colorado, discovered in 1870.

FRACTURING

The frequency of fracturing is controlled by both sedimentary and structural fabric (Harris and others, 1960; Harris, 1984; Murray, 1968,1973). The depositional factors which control the ultimate fracturing of sedimentary rocks include thickness, competency or brittleness of beds, ductility, and sequence of sedimentation. Many weathering features, such as mud cracks, gravity fractures, and slump features have confused the casual observer. These features are generally discernible by their lack of symmetry, repetition, and influence on geomorphology.

Modification of fabric by either subsurface or surface solution is generally controlled by a preexisting permeability fabric, either of depositional or tectonic origin. Often the controlling fabric (Fig. 1) is obscured by the grandiose cavern, and by mineralogical modification by solution and mineral replacement (Bretz, 1956).

Once the sedimentary and weathering influences have been determined, it is possible to correct, quantify, and qualify the degree of fracturing due to regional and local tectonic controls. These resultant data can be contoured into an iso-fracture map. The trends of the major and minor fracture sets can be traced to structure of the fracture patterns.

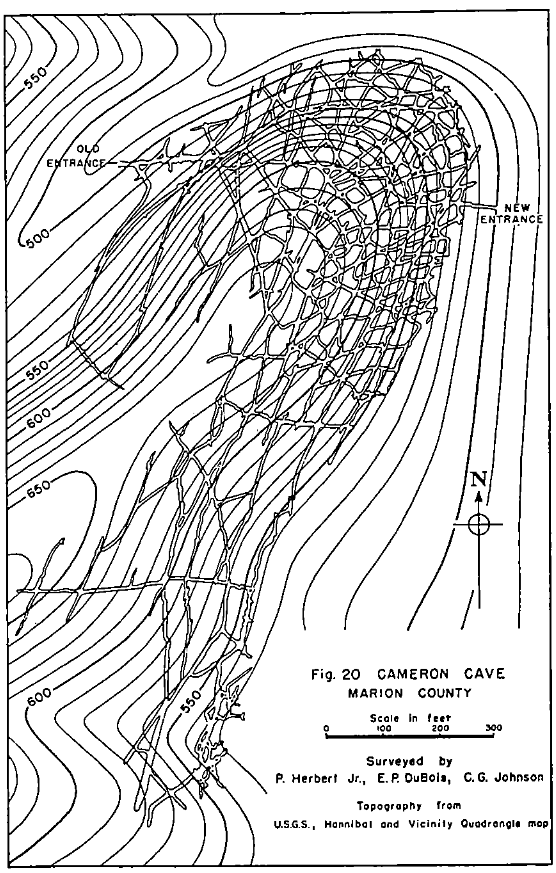

Figure 1. Topographic contours superposed on top of a map of Cameron Cave in Marion County, Missouri (reproduced from Bretz, 1956). Three sets of fractures control the solution porosity. Dependent upon local structure, at least two interpretations are possible.

Together, the fracture pattern map and the iso-fracture map, on either a regional or local basis, illustrates that maximum fracturing occurs in those areas of maximum change in dip and/or strike. These are the areas in which maximum stress occurs (Fig. 2).

It has been found that a second-derivative interpretation of structure *at the level of fracture study* duplicates the areas of maximum curvature and deformation and can be used for exploration.

TECTONIC DOLOMITIZATION

In 1954, following initial completion of the fracture studies and transfer to the Tulsa office, I was assigned to a division of carbonate studies by Stanolind, which led to further studies of the similarities between dolomite deposits and areas of intense fracturing (Harris, 1966,1970; Mickey, 1982; Sargent, 1969). Work by Dr. William E. Ham on some of the dolomites in the Arbuckle Mountains was studied, and the conclusion was reached that the surface dolomitization predated the faulting with which it had previously been associated. In addition, aerial photos and surface work showed that the dolomite occurs on structures in areas with sharp changes of dip and/or strike.

The first well drilled for tectonic dolomite was drilled in 1957, 160 acres downdip from a dry hole which was drilled in a crestal position on a seismic feature. This well produced from two zones of tectonic dolomite within the Devonian.

Upon my resignation from Stanolind Oil and Gas Co. (Pan American Petroleum Corp.) in late 1959, that company allowed publication of the aforementioned fracture thesis, and subsequently in 1967–68, aided in the preparation of SEM slides for use in the continuing-education program of AAPG.

Work with Lon B. Turk and various clients in Michigan, as well as worldwide studies, have shown that many of the dolomite producing fields which have been classified as reefs, banks, and carbonate bars are actually metasomatic replacement of calcite by tectonic dolomite and other minerals. In some cases, these occurrences are a secondary influence on oil entrapment, but often are of primary importance in the oil-field fabric.

During the early work on fracturing, it was believed that fractures were only important in areas of positive structural departure. However, work on tectonic dolomite showed that often maximum fracturing and dolomitization occur in areas of negative departure. Tectonic dolomite can occupy extensive synclinal fold areas (e.g., Scipio-Albion), which can also include fracture-producing fields, such as the Florence field.

The influence of deposition on the occurrence of tectonic dolomitization is less obvious, but of

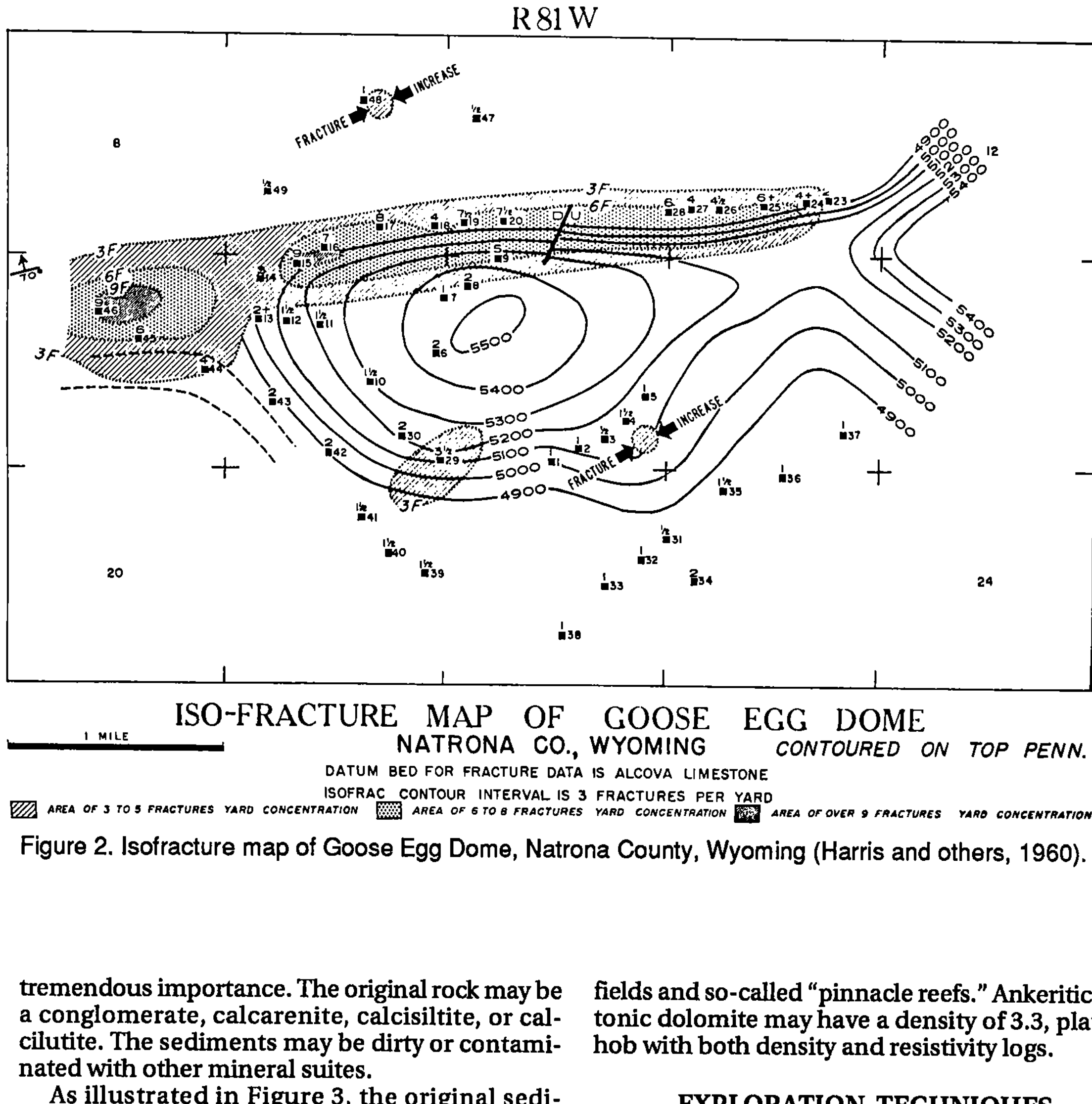

Figure 2. Isofracture map of Goose Egg Dome, Natrona County, Wyoming (Harris and others, 1960).

tremendous importance. The original rock may be a conglomerate, calcarenite, calcisiltite, or calcilutite. The sediments may be dirty or contaminated with other mineral suites.

As illustrated in Figure 3, the original sedimentary rock, subjected to tectonism, undergoes fracturing, diminution of grain size, dilation, and eventually plastic flow. Connate or hydrothermal fluid may furnish a source of magnesium and create the more-stable double salt dolomite. Lathlike plate-glass anhydrite may be a byproduct of this metasomatic change. If no magnesium source is present, the ultimate result may be a carbonatite flowage or plug (diapir). Dependent upon tectonic fabric, extent and number of orogenic movements, tectonic dolomitization may be of local or regional development. Obviously, the older sediments have a greater potential of being so altered by repeated structural movement.

Valuable ore minerals may be related to these tectonic dolomites, and during mining of lead, silver, zinc, and fluorite deposits in rocks as old as Precambrian, some oil flows were discovered. Also, similar minerals occur in many carbonate oil fields and so-called "pinnacle reefs." Ankeritic tectonic dolomite may have a density of 3.3, playing hob with both density and resistivity logs.

EXPLORATION TECHNIQUES

The complexity of fracture and tectonic dolomite reservoirs requires a new and concise modification of exploration techniques, involving geophysical interpretation, structural objectives, drilling and testing techniques, completion methods, production procedures, and logging and interpretation as well as reserve calculations (Harris, 1960,1968,1972,1969,1970,1984; McQuillan, 1973; Murray, 1968,1973). The references above are based upon experiences, and should be consulted. Suffice to say that standard operating procedures must be modified to economically explore and develop these enigmatic tectonic dolomite traps.

RECOMMENDATIONS

1. Use the "sermons in stone" that are there to read in samples and cores.

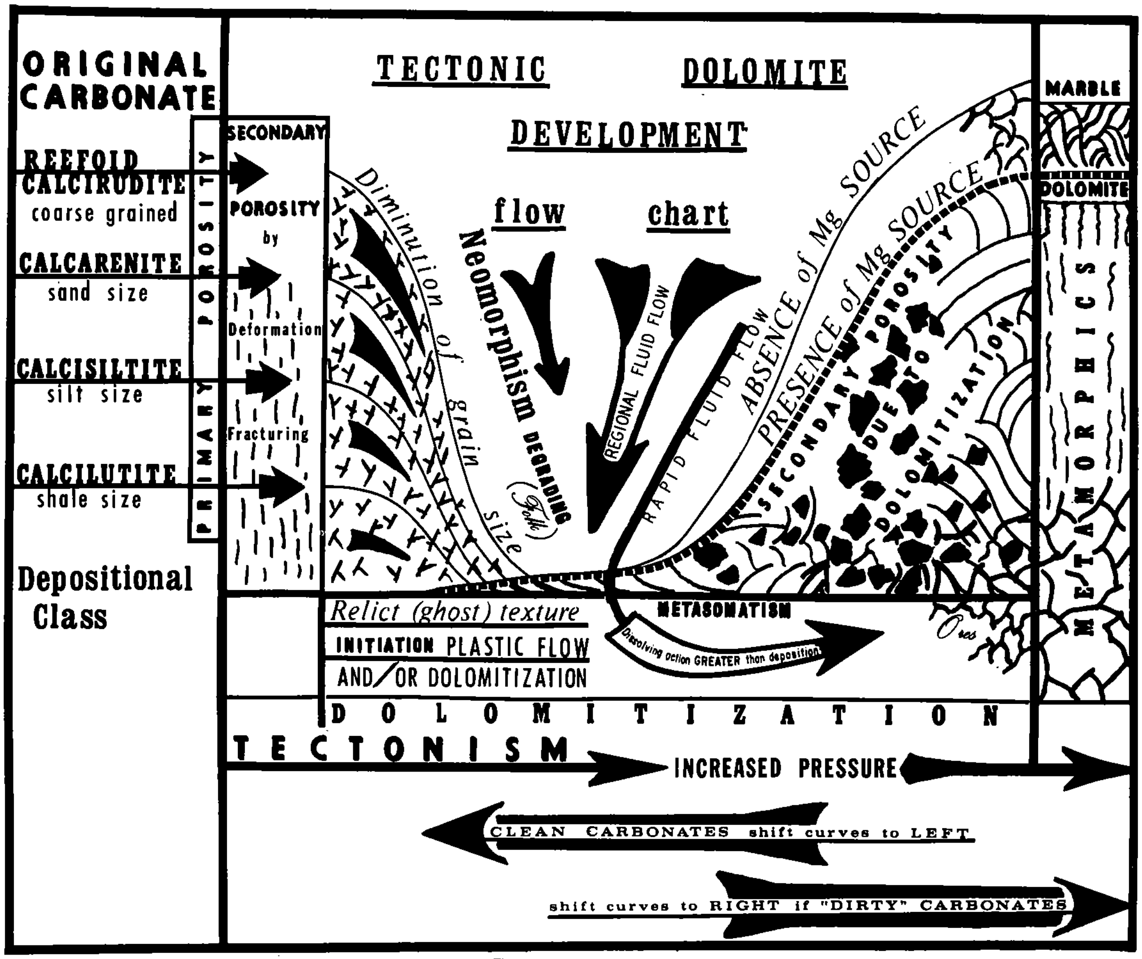

Figure 3. Flow chart showing tectonic influence on fabric alteration and ultimate result of metasomatic replacement or nonreplacement of original sediment types.

2. Use the "God-given" computer between your shoulder blades.
3. Treat every instrument as a tool and not a panacea.
4. You make your own luck.

ACKNOWLEDGMENTS

Credit is given to the many organizations, clients, and companies who, for more than 40 years, have supported my investigations of fractures, carbonate rocks, and tectonic dolomitization. A full list of publications is included, from which I have prepared slides and figures. Some of the slides have never been published, but credit is given where credit is due. I wish to thank Stanolind Oil and Gas Co. (AMOCO), Tulsa Geological Society, American Association of Petroleum Geologists, Dr. Garvin L. Taylor, Dr. Jack Walper, Dr. A. N. Murray, Lon B. Turk, Erik A. Nelson, and Robert Behrman, and especially Alan P. Emmendorfer, who made the excellent oral presentation due to the author's sickness. Robert O. Fay reviewed and edited the manuscript.

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SILICA IN THE UPPER ARBUCKLE GROUP, SLICK HILLS, SOUTHWESTERN OKLAHOMA

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INTRODUCTION

The dominant structural elements in southwestern Oklahoma are the Wichita uplift and the Anadarko basin. The hinge line between these elements is the frontal fault zone, a region of complicated faulting and folding with a combined stratigraphic relief locally in excess of 45,000 ft. The Slick Hills of southwestern Oklahoma, the exposed part of the fault zone, display thick sections of the Arbuckle Group; only the top part of the West Spring Creek Formation is not exposed (Figs. 1,2). The Hills are divided into two roughly equal areas by the Blue Creek Canyon fault; as the western Slick Hills are an area of great structural complexity, most of our observations are based on work in the relatively simple, generally homoclinal sequences of the eastern Slick Hills.

Our purpose in this paper is to describe and interpret the occurrence of silica in the upper part

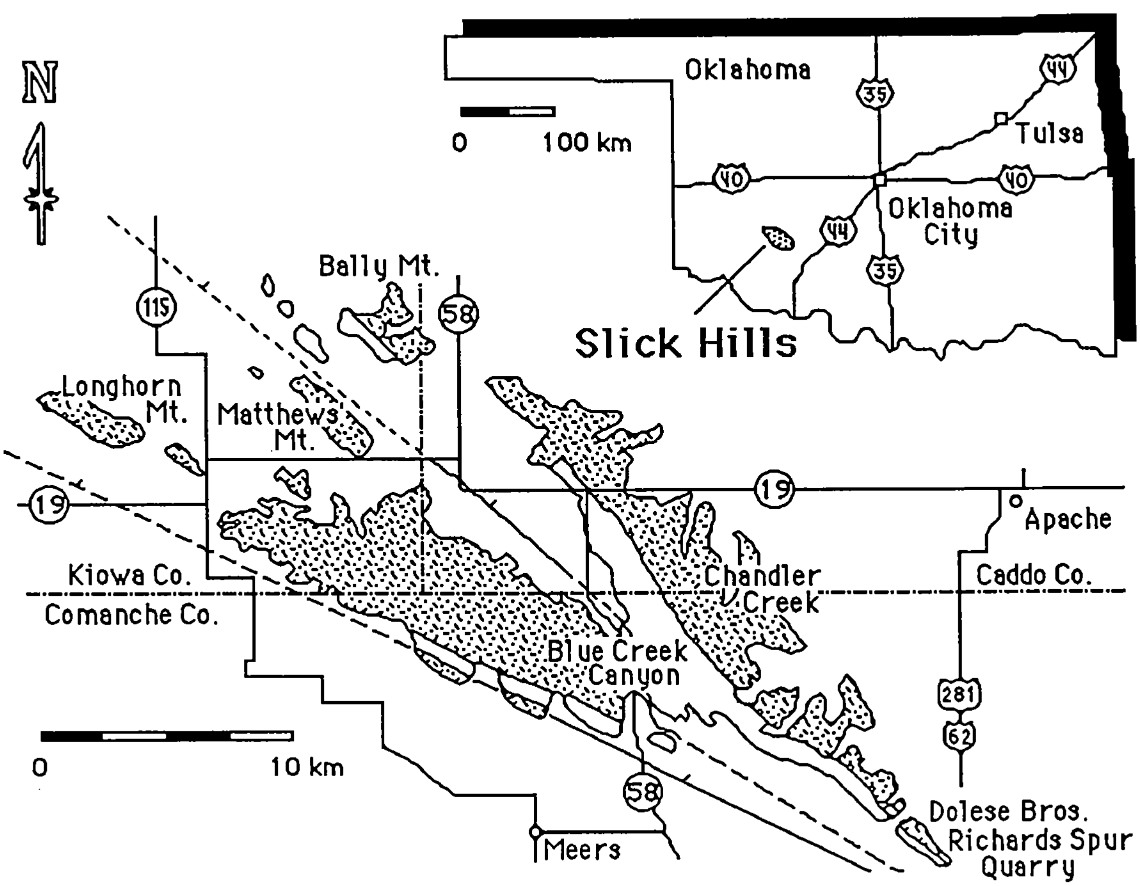
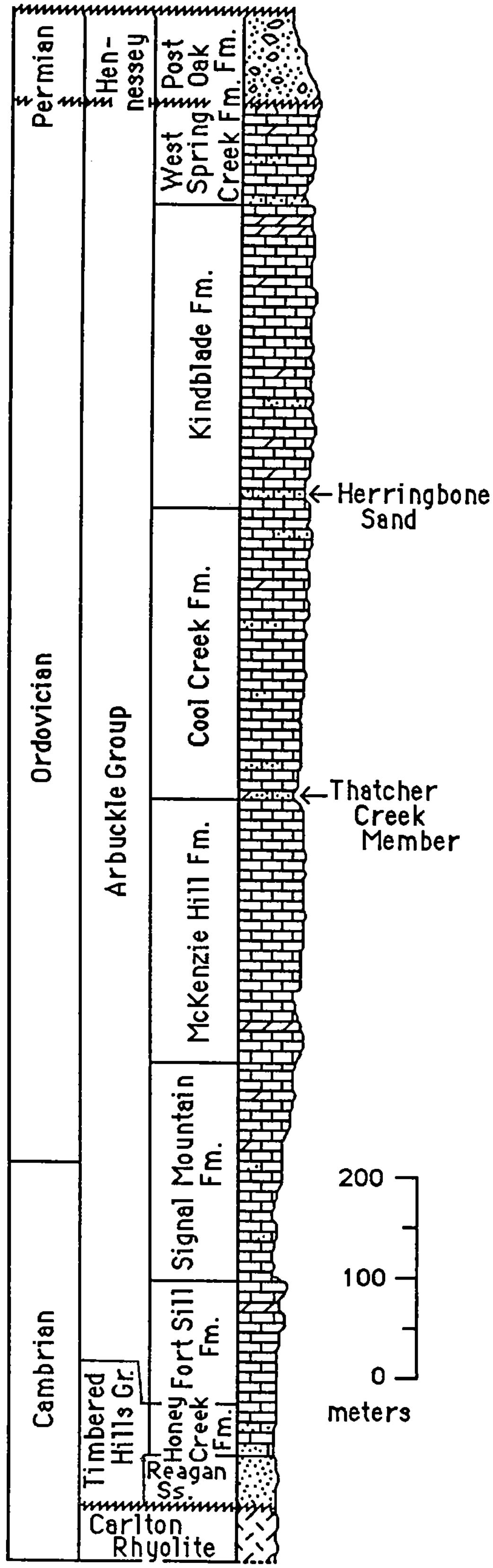


Figure 1. Location map of the Slick Hills, southwestern Oklahoma.



of the Arbuckle Group (specifically the upper McKenzie Hill, Cool Creek, Kindblade, and available parts of the West Spring Creek Formations). We use the term *silica* inclusively to cover both detrital quartz and various forms of secondary silica ("chert").

DETRITAL QUARTZ

Stratigraphic Occurrence

There are few useful lithostratigraphic markers within the upper Arbuckle Group, to the extent that it is difficult to justify formational subdivision of strata (Fig. 2). However, the base of the Cool Creek Formation is clearly demarcated by a distinctive quartz-rich calcarenite, which we have previously described as the Thatcher Creek Member (Ragland and Donovan, 1985). The areal extent of the Thatcher Creek Member is remarkable; it can be traced from the Slick Hills to the Arbuckle Mountains and appears to coincide with a significant change in conodont biostratigraphy (R. L. Ethington, personal communication, 1989). The stratigraphic significance of this member is that it is the first consistently mappable quartz-bearing unit above the Honey Creek Limestone (Timbered Hills Group). The genetic significance of the member is that it heralds the frequent occurrence of similar quartz-rich calcarenites in the overlying Arbuckle formations, particularly the Cool Creek and West Spring Creek Formations.

Before discussing the generalities of quartz-sand distribution, we will describe two occurrences in detail: the Thatcher Creek Member, and a unit just above the base of the Kindblade Formation which we informally refer to as the Herringbone sandstone. In both cases we have correlated these units over large areas of the Slick Hills.

Many of the quartz sands in the Cool Creek and Kindblade Formations do not form thick units that can be easily recognized in the field. Siliciclastic detritus is usually dispersed in small percentages as an admixture with carbonate grains in micritic matrices. True quartz sandstone beds are usually <0.25 m thick and are often laterally discontinuous (albeit such units are more persistent in the incompletely exposed West Spring Creek Formation). Two distinctive quartz-sand units are exceptions to this generality. The Thatcher Creek Member is a complex sandstone/carbonate unit at the base of the Cool Creek Formation (Ragland and Donovan, 1985) and the Herringbone sandstone is a quartz-sand/carbonate unit occurring near the base of the Kindblade Formation.

Figure 2 (left). General stratigraphic column of outcrops in the Slick Hills.

Thatcher Creek Member

Petrography.—Although the Thatcher Creek Member is one of the easiest units to correlate across the area, the unit is neither consistent in thickness nor composition. Thickness ranges from 4 m in the southern exposures to <1 m in the northeastern-most exposure (Fig. 3). The percentage of quartz-sand grains decreases from the southeastern parts of the Slick Hills, where quartz detritus dominates, to northwestern sites where calcimicrite and dolomicrite comprise the majority of the member. Intraclasts, ooids, and peloids are abundantly intermixed with the siliciclastic grains; calcite and dolomite comprise the matrices and most of the cements. Medium- and small-scale cross bedding, ripple marks, and/or parallel stratification are found at all of the outcrops. Disrupted bedding, vertical burrowing, and horizontal trace fossils are commonly seen. Mud cracks are found in some of the more micritic facies of the member.

Mineralogically, the Thatcher Creek sandstones contain a high enough percentage of feldspars and/or rock fragments such that more than half the samples examined can be classified as subarkoses or sublitharenites (Fig. 4). Nearly half of the samples have ≥50% micritic matrix.

Grain size ranges from coarse silt to coarse sand with a mean of 2.38 phi (medium sand) and a standard deviation of 0.89 (poorly to moderately well sorted). Grains vary from very angular to rounded. Quartz grain properties in these sands may have been modified by the cushioning effect of lime mud and reaction with softer carbonate particles.

Depositional environment of the Thatcher Creek Member.—The Thatcher Creek Member records the existence of a complex shallow-marine setting which involved deposition in the subtidal, intertidal, and supratidal zones (Fig. 5). The thicker units were probably deposited in the subtidal zone, where current energy was strong enough to deposit cross-bedded, ripple-marked, and planar-bedded sands. The lack of a significant algal facies suggests that current action and abrasion may have been too strong for the establishment of boundstone varieties.

The less-quartzose and generally thinner-bedded units of the member were deposited in the shallower waters of the intertidal to supratidal zones. In some cases, thin, erosive-based channel deposits can be discerned; these often contain rip-up clasts of local origin, together with siliciclastic grains. Generally, the thinner-bedded units are dolomicrite and calcimicrite, and quartz sand and silt are either in discontinuous laminae or are absent. Mud cracks (as much as 10 cm in depth surrounding plates as much as 20 cm in diameter) and pseudomorphs after evaporite minerals in these carbonate mudstones indicate emergence and desiccation. Given the great dimensions of the Arbuckle carbonate platform, such evidence of emergence suggests that small, impermanent islands may have developed from time to time.

Slowed subsidence and/or global sea-level oscillation may have been the cause of a gradual regression in the aulacogen that began during the time of deposition of the uppermost McKenzie Hill sediments. Differential uplift of the exposed craton and/or increased subaerial exposure of quartzose sediments resulted in an influx of quartz detritus. The uppermost 30 m of the McKenzie Hill Formation is composed of laterally interweaving algal boundstones and micrites containing shallow-water features (mud cracks and evaporite pseudomorphs). Dispersed fine-grained quartz sand is found in the McKenzie Hill at least 30 m below the top of the formation.

The Thatcher Creek Member is the ultimate product of an influx of quartz sand which was associated with shallowing of marine waters during the time of deposition of the uppermost McKenzie Hill and lower Cool Creek Formations. As quartz sand was introduced into the aulacogen, tides, waves, and storms reworked the detritus. The sand was distributed in greatest abundance in the subtidal environment and in decreasing quantities in intertidal through supratidal areas. Quartz-rich beds were deposited as sheet sands resulting in units similar to Johnson's (1978) "sub-littoral sand sheets."

Herringbone Sandstone

Petrography.—The Herringbone sandstone is an informally named quartz sandstone and carbonate unit located ~3 m above the base of the Kindblade Formation (Fig. 2). It is herein given that name because of its distinctive "herringbone" cross bedding, small- and medium-scale trough and planar cross beds, which record bipolar current motion. The sandstone varies in thickness from 0.10 m to 2.33 m. The thickest outcrops are located on Bally Mountain; all other sections are <0.65 m thick (Fig. 6). As in the Thatcher Creek Member, quartz detritus is intermixed with intraclasts, ooids, peloids, micrite, and spar cement.

On Bally Mountain, the sandstone can be divided into three subunits based on sedimentary structures (Fig. 6) The lowest subunit (A), which rests disconformably on a dense, thin-bedded micrite, consists of quartz sandstone and micrite which are mostly parallel laminated.

Overlying the parallel-laminated subunit is the cross-bedded subunit B. The lowest 10 cm of this subunit (B1) is composed of undulose sandstone laminae with wavelengths of 30 cm and heights of 3–4 cm. Maximum angles of undulation dip are

<15°. A second overlying stack of undulations has been truncated by erosion. The morphology of these undulations closely resembles hummocky cross stratification (Harms and others, 1982; Walker, 1984).

Trough cross-sets (B2) with average thicknesses of 14 cm comprise the bulk of the B subunit in the thickest section on East Bally Mountain, but this unit thins progressively to the east. The B2 subunit is capped by a 0.30-m-thick band of bipolar cross

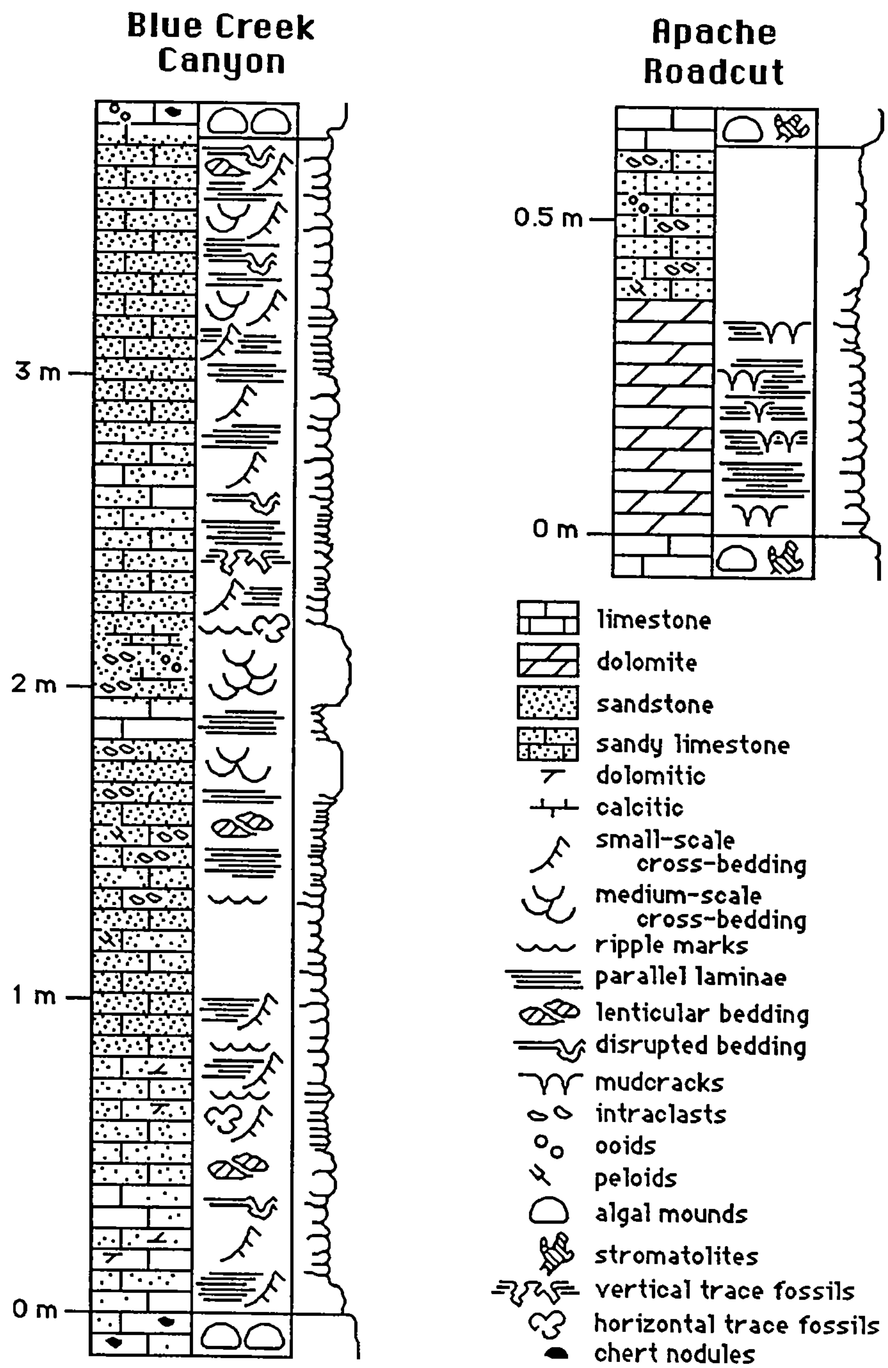


Figure 3. Representative measured sections of the Thatcher Creek Member. The Blue Creek Canyon section and other sections in the southern part of the area were probably deposited in the subtidal zone. Sections in the northern Slick Hills (e.g., the Apache road cut section) are generally thinner, less quartzose, and more dolomitic than the southern sections, suggesting highest intertidal to supratidal depositional zones.

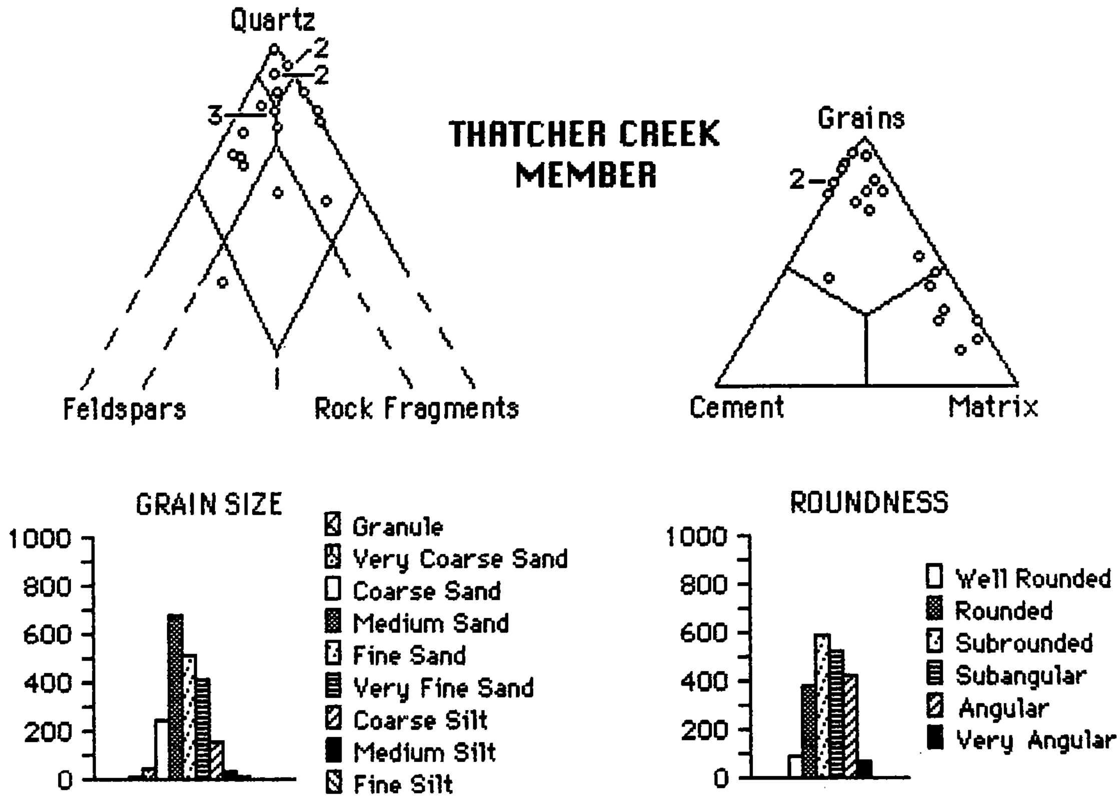


Figure 4. Statistical data on 21 samples of Thatcher Creek Member sandstones. In many cases the sandstones contain feldspars and rock fragments resulting in their classification as subarkoses, sublitharenites, lithic subarkoses, and lithic arkoses. Matrix is abundant in some of the samples. Grain-size mean is medium sand with a standard deviation of 0.89 (moderately sorted). Roundness ranges from very angular to well rounded, with a peak in the subrounded range.

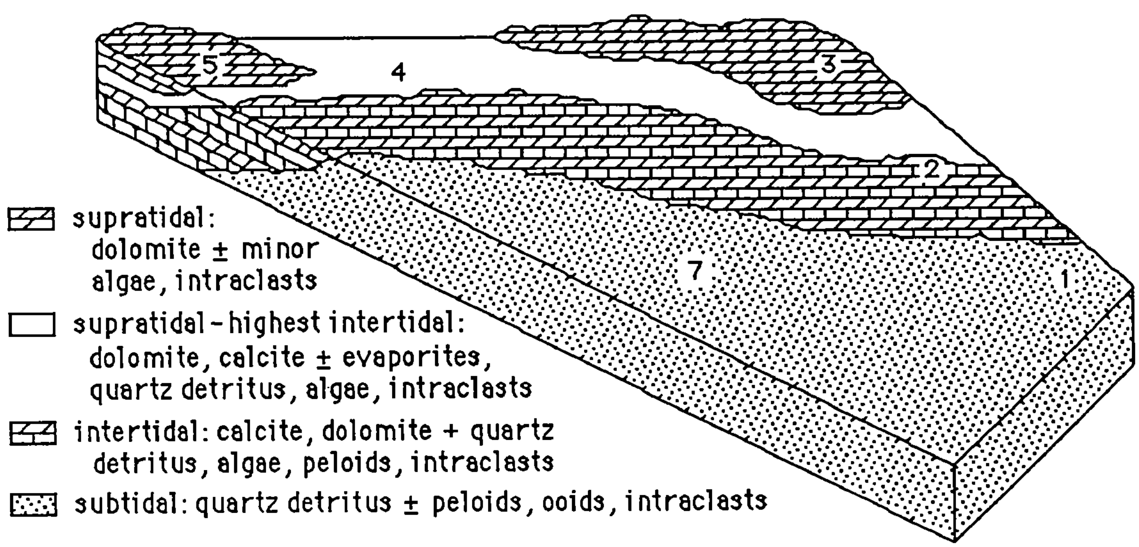


Figure 5. A hypothetical "moment in time" during deposition of the Thatcher Creek Member. Quartz sands were abundant in the subtidal zone, decreasing in abundance in the intertidal and supratidal zones. 1—Dolese Brothers Richards Spur quarry; 2—Chandler Creek; 3—Apache road cut; 4—Bally Mountain; 5—Mathews' Mountain; 7—Blue Creek Canyon.

beds whose sets intersect at a mean of 20° (B3). The bipolar subunit can be traced over most of the Slick Hills.

Subunit C forms the uppermost sandstone, which ranges in thickness from 0.30 to 0.51 m. In this unit medium-scale planar and trough cross-sets intersect at angles <21°. The entire Herringbone sandstone (A,B,C) is abruptly overlain by a 5-cm-thick fossiliferous biomicrite, which is itself overlain by a massive 45-cm-thick peloidal micrite.

Grain size of the siliciclastics in the Herringbone sandstone averages 2.20 phi, with nearly equal amounts of fine and medium sand (Fig. 7). The sands are moderately sorted (0.65). Grains are angular to well-rounded. In contrast to the Thatcher Creek Member, the Herringbone sandstone contains very minor amounts of feldspar and/or rock fragments; >85% of samples are quartz arenites. Less than 10% of the sandstones contain micritic matrix as a major component (i.e., the sandstones are better washed than those in the Thatcher Creek Member).

Depositional environment of the Herringbone sandstone.—The transition from the Cool Creek to Kindblade Formation reflects the gradual onset of generally deeper water conditions. Facies characteristic of intertidal to subtidal environments are found throughout most of the Kindblade, whereas supratidal to subtidal facies characterize the Cool Creek (Fig. 8). Consequently, in the Kind-

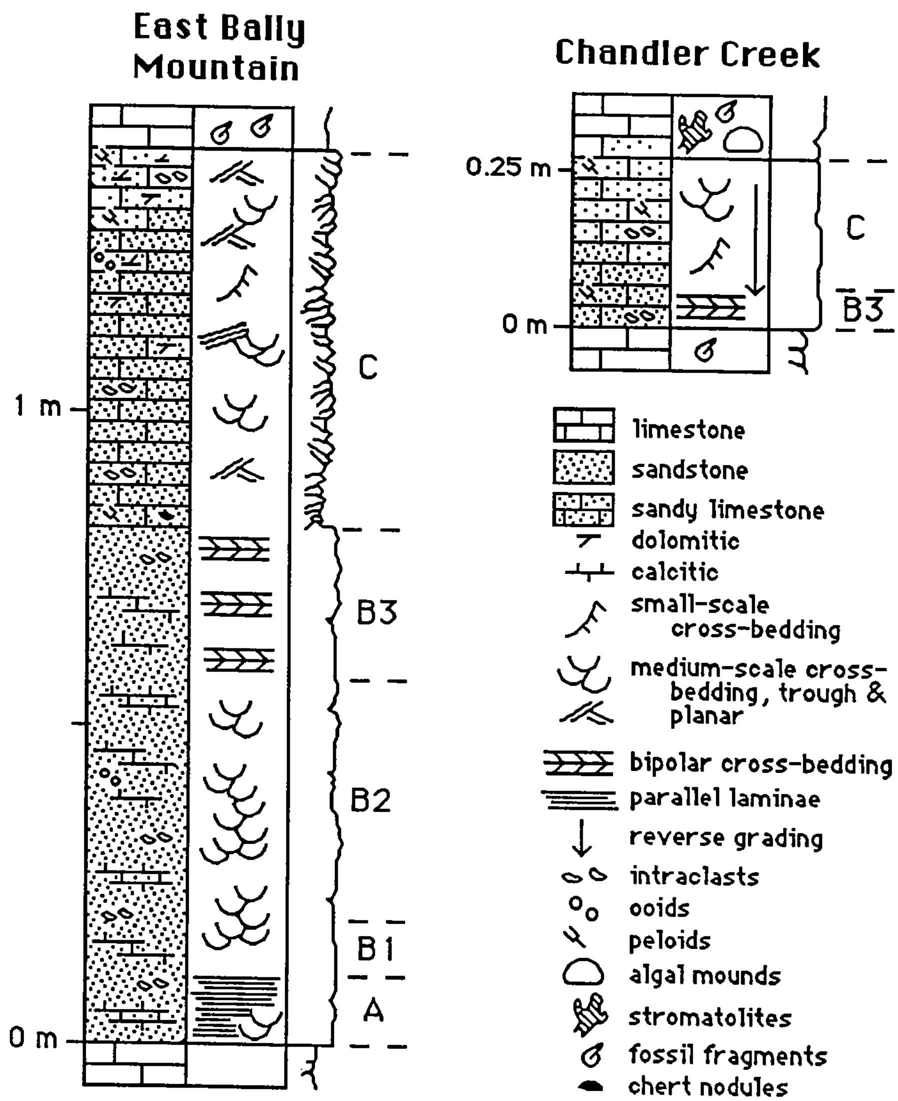


Figure 6. Representative measured sections of the Herringbone sandstone. The Bally Mountain section is a cross section of an offshore sand ridge or bar. The sand is divided into subunits based on sedimentary structures (see text for explanation).

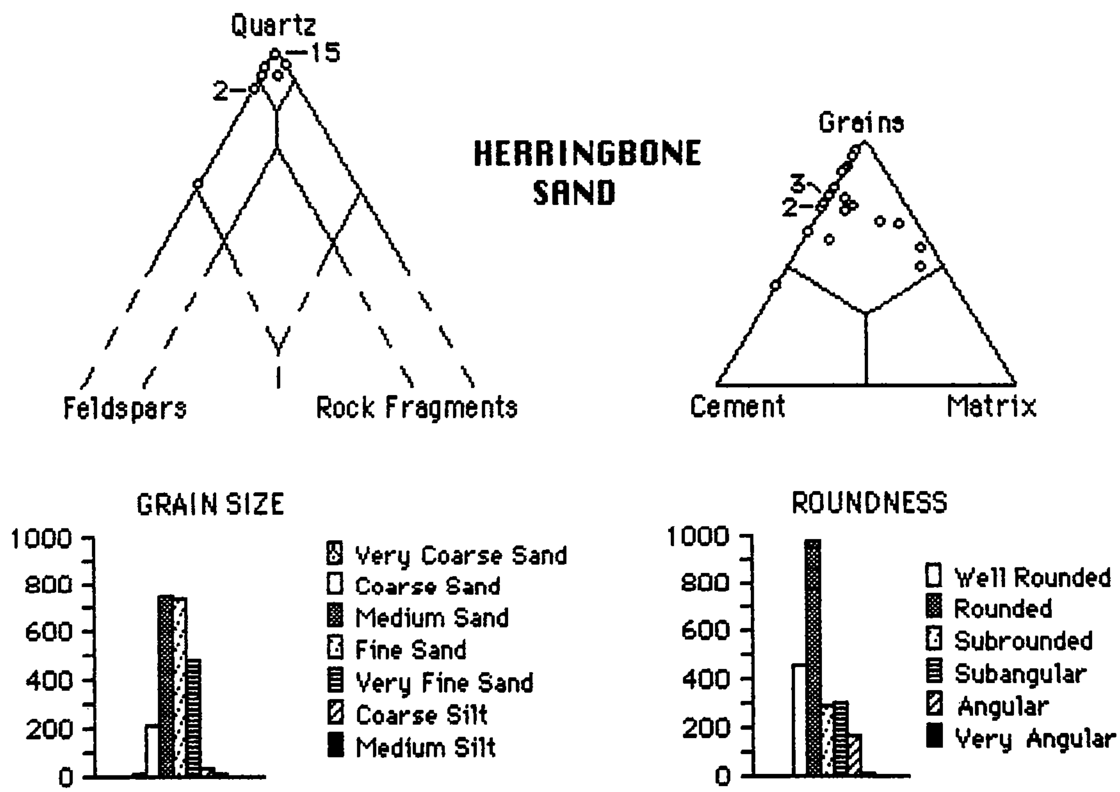


Figure 7. Statistical data on 21 samples of the Herringbone sandstone. Most of the samples are devoid of feldspars and rock fragments and can be classified as quartz arenites. Matrix is less abundant than in the Thatcher Creek Member samples. Grain size has a double peak at the medium and fine sand range and a standard deviation of 0.65 (moderately sorted). Most grains are well rounded to rounded.

blade, invertebrate fossils are more commonly seen and facies deposited in shallow restricted water are less frequently observed (Toomey and Nitecke, 1979; Donovan and Ragland, 1986). In general it would seem that the Kindblade was deposited during a period of general transgression, whereas the Cool Creek records a more regressive setting.

The Herringbone sandstone may be the remnant of a linear sand body, located at an unknown distance from the paleostrandline and perhaps similar to sand ridges and bars on modern continental shelves described by Reinson (1984) and Walker (1984). By analogy, the exposure on East Bally Mountain may be a cross section through the center of the bar.

It is suggested that the initial influx of sand moved into the area relatively quickly, resulting in the deposition of parallel-laminated to slightly cross-bedded sands above an erosional surface on the underlying micrites. The upper part of the parallel-laminated sand was subsequently reworked, possibly by storms, to form the undulatory bedding at the top of B1. Sand influx continued while the B2 unit was deposited. This sand was reworked and winnowed into clean sandstone cross beds.

Bipolar cross beds occur at the top of the B unit (B3). Although bipolar cross beds with sets at 180° to each other are not always diagnostic of tidal processes (Clifton and others, 1971; Miall, 1984), the large area over which they occur in this single subunit is strongly suggestive of tidal influence.

The planar and trough cross-bedded sandstones (unit C) overlying the bipolar sands are also slightly bipolar, but the cross beds are much less distinct than those in the underlying unit. Some of these sands were probably reworked from subunit B3. Eventually, all sand input ceased and relatively sand-free carbonate production resumed.

The Herringbone sandstone unit on West Bally Mountain differs from that on East Bally Mountain in that quartz sand is generally less abundant and more dispersed. Subaqueous shrinkage cracks indicate periodically quieter conditions with possible changes in salinity (Donovan and Foster, 1972). Most of the apparent correlatives to the B subunit on East Bally Mountain may have been deposited in an interarea between sand bars. During quiet periods, lime mud was deposited in an

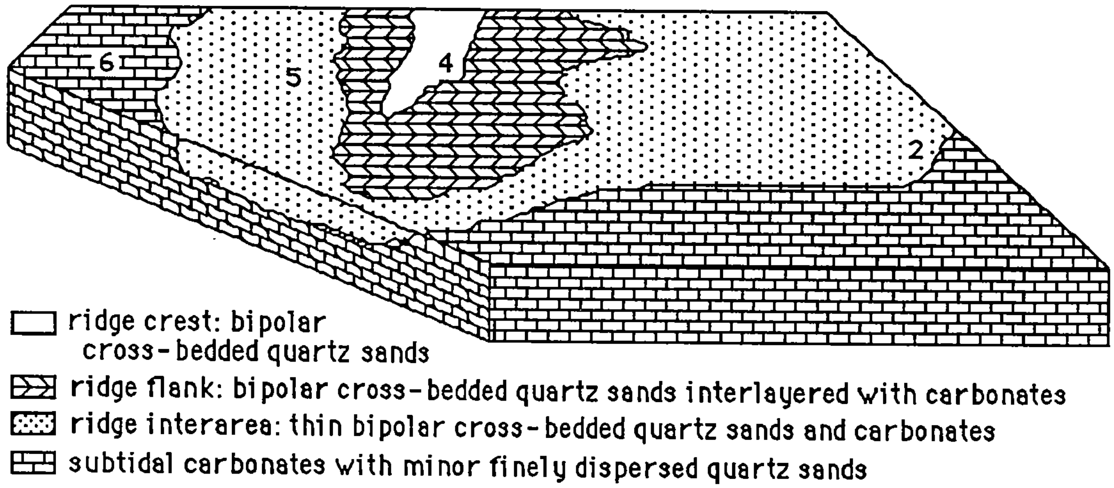


Figure 8. A hypothetical "moment in time" during the deposition of the Herringbone sandstone. The most abundant sands occur at the Bally Mountain outcrops (4) with sands thinning away from the sand ridge. 2—Chandler Creek; 4—Bally Mountain; 5—Mathews' Mountain; 6—Longhorn Mountain.

environment characterized by considerable biological activity, as evidenced by abundant peloids. Occasional storms and/or shifts in current patterns resulted in the periodic introduction of sand detritus into this environment.

The B unit on West Bally is capped by bipolar planar cross beds (B3), as on East Bally. As noted previously, this unit probably records a period of dominantly tidal influence during which sand was redistributed over a wide area by tidal activity.

The increased abundance of quartz sand in the C subunit at West Bally may represent deposition of sand on the flank of the ridge as sand at the East Bally Mountain site migrated westward. Quartz detritus influx either ceased before the ridge could migrate to the West Bally site, or migration was in a direction such that the main profile of the ridge is not presently exposed.

Elsewhere in the Slick Hills, the entire Herringbone sandstone member never exceeds 0.65 m in thickness. Each of these sections begins with the bipolar cross beds, suggesting that the initial sands of the sand ridge at Bally (units A, B1, and B2) did not remain permanently at the other Slick Hills sites. The bipolar sandstones represent a period of tidal-current reworking in which sand was removed from the bar and redistributed over a wide area. All outlying outcrops of the Herringbone sandstone are capped by the cross-bedded unit C. Quartz influx terminated abruptly at the top of the C unit, perhaps because in a generally transgressive setting the supply of siliciclastics was cut off.

General Features of the Quartz-Bearing Beds

The previous discussion has demonstrated that, although considerable facies variations are to be expected within quartz-bearing sandstones, the overall environment of their deposition was a shallow-marine setting. In this section we attempt to elucidate features and discern trends common to all such beds.

Petrographic considerations.—All the siliciclastic units examined are dominated by monocrystalline grains of quartz which show straight to slightly undulose extinction. Other grain types are uncommon; feldspars are increasingly rare in younger rocks. Grain size and sorting are facies-tied. Thus, for example, grains in cross-bedded sandstones are generally coarse and well sorted, whereas grains associated with intraformational conglomerates, while coarse, are less well sorted. Maximum grain size is approximately 2–3 mm.

Siliciclastic grain populations are not common in the Lower Paleozoic rocks of Oklahoma; the two stratigraphically closest occurrences are in the underlying Timbered Hills and overlying Simpson Groups. These three populations resemble each other in that they are dominated by monocrystalline quartz (showing straight or gently undulose extinction), although there is an overall tendency for mineralogical maturity to increase up section. Texturally the Simpson grains are finer, while in the Reagan Formation (Timbered Hills Group), small, well-rounded quartz pebbles as large as 10 mm in diameter are present. Pebbles and sand-sized grains are generally well-rounded, while sorting is tied to depositional facies.

It is conceivable that the three siliciclastic populations are of similar provenance. The rarity of polycrystalline quartz and detrital mica suggests that metamorphic terranes did not contribute substantially to the grain populations. Alternatively, but not exclusively, the detritus was the product of either recycling (conceptually more likely in the case of the younger populations) or extended and intensive weathering of a quiescent craton of low relief on which large tracts of granitic terrane were exposed. The latter suggestion is in accord with the character of the basement to the north and west of the Slick Hills, outside the boundaries of the southern Oklahoma aulacogen.

Facies relationships in the quartz-rich sandstones.—Although quartz-bearing sandstones, such as the Thatcher Creek Member and Herringbone sandstone, are distinctive in the field, small percentages of siliciclastic detritus can also be found widely dispersed throughout much of the upper part of the Arbuckle Group, even a few meters below the Thatcher Creek Member. However, there is a tendency for conspicuous amounts of quartz to be tied to a thinly bedded or laminated, frequently dolomitic facies, characterized by ripple marks, small-scale cross bedding, thin channels, rip-up clasts, and mud cracks. By analogy with descriptions of modern environments in the Persian Gulf and the Bahamas, this facies appears to be the product of high intertidal/supratidal sedimentation. It is the common capping component of thin (~3 m), highly variable shallowing upwards cycles, and forms a quantitatively significant proportion of the Cool Creek and, especially, the West Spring Creek Formations. Strata above and below this facies comprise a variety of deeper-water facies (oolitic and intraclast-rich grainstones, thrombolites, stromatolites, ribbon carbonates, mudstones, and wackestones). Quartz grains may be present in abundance in the grainstones and other clastic facies (for example, they often form the nuclei of ooids), particularly in rocks just above the laminated facies.

Discussion.—The presence of considerable amounts of siliciclastic detritus in the upper Arbuckle Group raises three related questions:

1) By what mechanism was detritus distributed across the immense early Ordovician platform?

2) What triggered the initial detrital pulse, as recorded by the Thatcher Creek Member?

3) Why is the quartz more common in the Cool Creek and West Spring Creek Formations than it is in the Kindblade?

The Lower Paleozoic stratigraphy of Oklahoma is dominated by shallow-water carbonate deposition on platforms which are very much larger than any modern examples. Introducing siliciclastic grains in any abundance to this platform probably required a general low-sea-level stand during which the proto-North American craton was exposed. In the absence of land plants it is reasonable to assume that a biological desert existed and that, as a consequence, sand populations were remarkably mobile. We suggest that the sand was blown and/or transported by sluggish rivers across the exposed Arbuckle platform and was then reworked in a variety of marine settings during a subsequent transgression.

The initial detrital pulse, as recorded by the Thatcher Creek Member, could be explained as a response either to distant uplift or to a general sea level fall. The mineralogical maturity of the member is an argument which can be used to support the second of these two mechanisms. If relative rises and falls in sea level are permitted a role in the development of the Arbuckle Group, then it is possible that the stratigraphic distribution of quartz sand mirrors the relative position of sea level (i.e., high during deposition of the Kindblade Formation when little quartz was available and low during deposition of the Cool Creek and West Spring Creek Formations). In this context, it is interesting to note that the end of Arbuckle Group deposition coincides with a major low stand (second order cycle) of Vail and others (1977), which would certainly be in accord with the large amounts of detrital quartz seen in the West Spring Creek Formation.

CHERT

Bedded Cherts

Petrography

While secondary silica, in the form of replacement nodules, void fills, and cement is abundant, primary chert facies are rare in the upper Arbuckle Group. Two such units, one each in the Cool Creek and Kindblade Formations, consist of 15- to 25-cm-thick beds of interlaminated carbonate micrite and microquartz (Fig. 9). The laminae are variable in thickness, ranging from 1.0 to 20.0 mm. Boundaries between the carbonate and silica laminae are gradational.

The most laterally persistent bedded-chert unit occurs ~15 m above the Herringbone sandstone in the Kindblade Formation and has been located at five sites in the Slick Hills. Assuming that the unit was continuous between these five sites at the time of deposition, the areal extent of the bedded chert was at least 130 km^2.

Near Chandler Creek (Fig. 2) the deposits above and below the Kindblade bedded chert are carbonates. On the other hand, on West Bally Mountain, the bed immediately underlying the chert is composed of cross-bedded quartz sandstone and calcimicrite. The mean grain size ranges from medium silt to coarse sand, and the grains are very angular to well rounded. The micrite contains small cauliflower chert nodules (<2 cm in diameter); these are possibly replaced evaporites (Chowns and Elkins, 1974) that formed in high intertidal to supratidal conditions. The top of the sandy unit is planar, perhaps because of subaerial exposure and erosion.

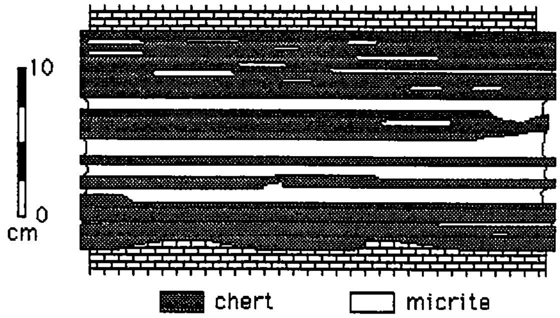


Figure 9. Sketch of bedded cherts.

Origin of the Bedded Cherts

Three lines of evidence associated with the Kindblade bedded chert suggest that the unit was deposited in shallow water: (1) the nature of the bounding sediments, (2) the nature of the included sediments, and (3) sedimentary structures in the latter.

Clearly the interlaminated chert and micrite was deposited in an unusual environment. Perhaps the closest modern analog is the Coorong Lagoon/lakes district of southern Australia, where similar laminae are the result of seasonal climatic variations (Alderman and Skinner, 1957; Skinner, 1963; Skinner and others, 1963; Peterson and von der Borch, 1965; Bathurst, 1975). If this analogue applies then it is necessary to postulate that part of the platform was temporarily cut off from open-ocean circulation as an isolated lagoon or marginal lake. Seasonal fluctuations within this water body would have resulted in variations in pH, as a result of which silica would have been alternatively dissolved and reprecipitated. Seasonal(?) flooding events, such as higher than normal tides and/or storms, would have led to washover of "normal marine" water into the lagoon. Such relative freshening may have accelerated algal growth and thus led to an increase in photosynthetic activity. In turn, this activity would have induced precipitation of calcium carbonate and a concomitant rise in pH (Alderman and Skinner, 1957). A high pH (9+), in what was presumably a hot climate, would have caused silica in the sediments to go into solution. The most likely source of this silica was the available adjacent quartz sand. Subsequently, as the lake/lagoon evaporated, algal activity ceased, pH decreased, and silica was deposited. At the same time salinity would have increased and, in some cases, may have caused shrinkage cracks to form in the still unlithified micrite, under conditions similar to those which caused the subaqueous shrinkage cracks discussed by Donovan and Foster (1972). As the next generation of silica was precipitated on top of the micrite, the cracks were filled or partly filled with silica (Fig. 10).

Some of the dolomicrite and calcimicrite laminae contain very small (0.10 mm) cauliflower chert nodules and calcite-filled pseudomorphs, possibly recording the replacement of gypsum and/or anhydrite. Presumably these evaporite minerals formed in the carbonate layers during the dry season, and were dissolved and replaced by silica during the ensuing wet season. The topmost bedded cherts at the Chandler Creek site contain very small, apparently subaerial mud cracks, suggesting that eventually the "Coorong-type" setting evolved into a more orthodox high intertidal setting.

Assuming that the Coorong analogy can be correctly applied to laminated cherts of this type, and that the postulated flooding events were seasonal, it follows that the chert/micrite couplets may be varves. If so, in the two cases under discussion, the silica-producing settings existed for a brief span of time, no more than about 10 years.

Secondary Silica

The most important diagenetic products in the Arbuckle Group are calcite, dolomite, and silica. Each occurs as cement and as replacement products. Calcite and dolomite compose the majority of these products; secondary silica is less abundant overall (Ragland, 1983,1984).

Chert is uncommon in the lower part of the Arbuckle Group (Barthelman, 1969; Brookby, 1969; Stitt, 1977,1983). Chert nodules begin to appear ~100 m above the base of the McKenzie

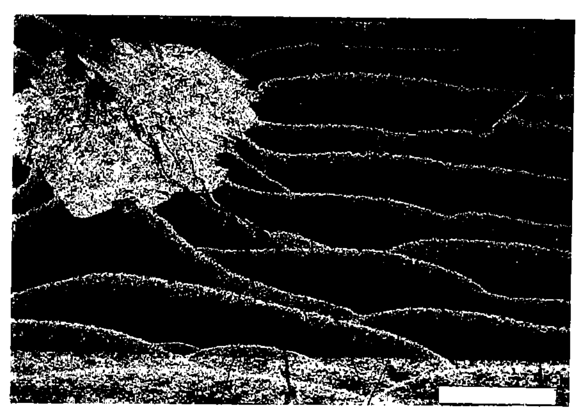


Figure 10. Silica nodule in a micritic lamina of the bedded-chert lithofacies. Cracks, possibly resulting from subaqueous shrinkage, acted as conduits for mobilized silica and were eventually cemented with silica. Silica nodule may be the replacement of an evaporite nodule (Chowns and Elkins, 1974, "cauliflower" cherts). Bar equals 1 mm.

Hill Formation, and above that secondary silica is an important component of the Group.

Morphology and Petrography of Secondary Silica

In the field, chert is conspicuous as nodules which adopt a variety of forms related to the character of the host sediment. Some of these forms (e.g., burrow-fillings and fossil mold-fills) appear to reflect obvious permeability pathways in the host sediment. Other forms (cauliflower chert) probably record the one-time existence of evaporative sulphate nodules (Cloyd and others, 1986). In some cases chert has formed at sites where a chemically propitious environment might be envisioned, e.g., in the center of stromatolites where algal decay presumably produced a locally more-acidic environment. However, in the majority of cases the chert is seen as nodules which, while clearly concentrated along particular layers or in particular zones, do not offer much firm evidence as to their origin. Individual nodules are usually flattened parallel to bedding and may be as much as 50 cm across. In places, nodules are so densely developed that they have coalesced into pseudolayers of chert. Chert-bearing horizons can usually be traced laterally for long distances (in one case for at least 4 km) without significant variation in either the macroscopic or microscopic form of the chert. This observation probably reflects the stability of the water table developed on the Arbuckle platform. Some chert horizons are associated with an early phase of fabric-selective dolomitization (Donovan and others, 1986).

As far as can be deduced from field evidence, many of the cherts formed relatively early in the history of the sediment; not only are textures fabric selective (and hence probably predate lithification), but also many intraformational conglomerates contain reworked chert clasts identical to in-situ chert nodules elsewhere in the sequence.

Almost every described microscopic morphology of secondary silica is present (Fig. 11). Microcrystalline chert is the most abundant form, occurring mostly as a replacement product and, more rarely, as pore-lining cement (Fig. 11A). Most fibrous chalcedony occurs as pore-lining material with minor amounts replacing calcite (Fig. 11A). Chalcedonic forms include chalcedonite (length fast), quartzine (length slow), spherules, zebraic fibers, and "feathery" forms that resemble lutecite. Equant megaquartz almost always occurs as pore-filling cement or as syntaxial overgrowths on detrital quartz grains (Fig. 11B). Occasionally, megaquartz crystals entrap remnants of the original host mineral, either carbonates or traces of evaporites (anhydrite and celestite, Fig. 11C).

Origin of Secondary Silica

As noted above, the earliest phases of silicification began before the sediments were lithified. Evidence supporting this conclusion includes compactional draping of carbonate laminae around silica nodules, silica-filled fenestrae, void-filling silica cement which predates calcite cementation, and the reworking of silica nodules as intraformational clasts (Fig. 11D). Initial porosities and permeabilities in the carbonate and siliciclastic sediments were probably very high (50–80% [Tucker, 1981]). Assuming silica mobility in interstitial waters, the silica-containing fluid could easily have moved to points of precipitation. Growth from such points was probably not continuous but occurred in stages as hydrochemical conditions allowed silica dissolution to alternate with precipitation. Multiple nucleation points were probably the norm. If growth continued for a relatively long time, small nodules intersected and merged forming large nodules and, eventually, pseudolayers.

The most likely source of silica in the upper Arbuckle Group is from the dissolution of detrital quartz grains. In this context it is interesting to note that there is a general correlation between the occurrence of detrital quartz and the abundance of chert. In thin section many detrital quartz grains show signs of significant corrosion; it is possible that the finer grain size modes of quartz have been entirely removed by dissolution. Siliceous sponges are an alternative source of silica; we have observed sponge spicules in both the Kindblade and the McKenzie Hill Formations. However, the quantity of such sponges is small and seems insufficient to account for the large volumes of secondary silica present in these rocks.

Chert Formation

Various mechanisms probably operated to produce the different chert textures observed. In this section we assess some of these mechanisms.

Epstein and Friedman (1982) demonstrated that carbon dioxide, oxygen, the bicarbonate ion, and pH vary substantially within modern reefs as a result of photosynthesis. Carbon dioxide and bicarbonate ion values remain low during maximum photosynthesis, while oxygen and pH values escalate. They hypothesized that carbonates are deposited and silica dissolved during the daylight hours of photosynthesis with maximum activity prior to sunset.

If this mechanism was operative in the Ordovician, early silicification of the Arbuckle Group sediments could have begun during deposition, especially in those sediments that were rich in algae. Photosynthesis during daylight hours would have facilitated dissolution of silica (quartz detri-

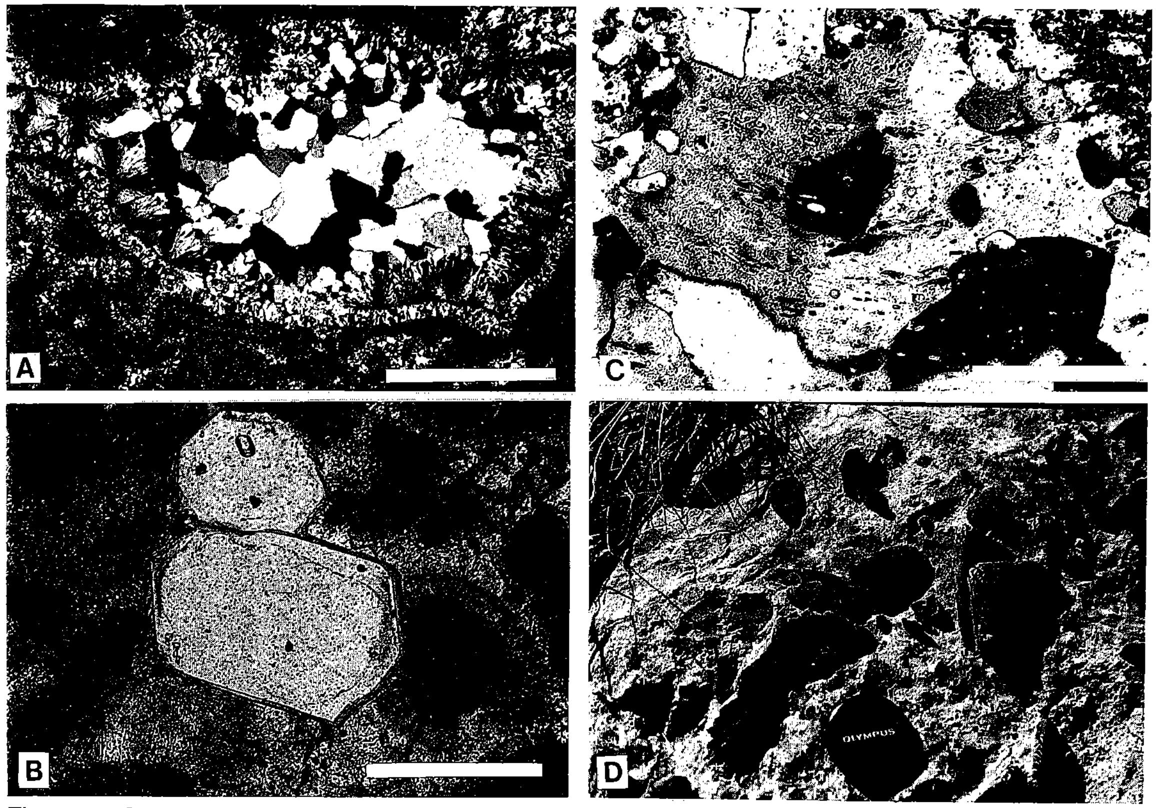


Figure 11. Secondary silica in the Arbuckle Group. *A*—Vug in silicified stromatolite is lined with a thin coat of fine-grained chalcedony. Radial fibrous chalcedony partially lines the vug. Final infill is drusy megaquartz. Scale bar equals 1 mm. *B*—Syntaxial overgrowth on a detrital quartz grain. Overgrowth partially fills pore spaces and partially replaces oolitic cortices. Scale bar equals 250 μ. *C*—Megaquartz containing remnants of anhydrite laths. The fabric of the original mineral is retained in the curved lath traces. Scale bar equals 250 μ. *D*—Silica intraclasts mixed with carbonate intraclasts and matrix, attesting to early silicification.

tus) as mounds, reefs, and encrustations added to their carbonate mass. With the onset of evening, reversals of the hydrochemical conditions would have resulted in cessation of the dissolution of silica, but not necessarily the dissolution of carbonates. Epstein and Friedman (1982) recorded maximum pH values of 9.95 during peak photosynthesis, but a minimum of only 7.6 during least-active photosynthesis; the latter value is not necessarily low enough to dissolve calcium carbonate. The highest pH levels were recorded in algal buildups (green and brown algae). If the blue-green algae of the Early Ordovician produced as high (or higher) pH levels, silica dissolution during the day would have been extensive over the aulacogen. However, at each individual site dissolution probably occurred only for a brief time during the period of maximum photosynthesis; hence, interstitial movement would have progressed incrementally under this diurnal control.

A commonly observed occurrence of chert involves hand samples which appear to be completely silicified. However, when such nodules are examined in thin section, the silica is seen to be intimately mixed with carbonate matrix. In such cases, assuming that the mechanism outlined above was operative, silica probably remained dispersed throughout the sediment, suggesting that: (1) movement was too rapid to allow concentration at a nucleation point; (2) a sufficiently attractive(?) nucleation point was unavailable; (3) movement was slow and permanent cementation by silica occurred in a dispersed manner, blocking further fluid movement; and (4) hydrochemical conditions changed, trapping dispersed silica.

However, in the general case, we suggest that extended periods of unchanging hydrochemical conditions resulted in the growth of nodules around specific nucleation points as mobilized silica aggregated (although silica in solution may not have precipitated entirely each evening as the pH decreased but may have remained in solution, or near-solution, as a gel). The nature of the nucleation sites is uncertain, but may have involved precipitation on detrital quartz dust, particularly at sites where decomposing organic matter produced acidic "haloes." The attraction of the silica molecules to nucleation points would have resulted in the continued growth of the nodules. The interior parts of the mass of the silica would have stabilized because of three-dimensional attraction forces between the silica molecules. Silica at the edges of nodules continued to dissolve and reprecipitate until such sites were in turn shielded from the effects of the surrounding microenvironment and incorporated within the nodule. In general, nodules continued to grow until a change in conditions took place; in this context, rinds found on many nodules and pseudolayers may record such a change. Thus, while the interiors of the nodules were fixed and probably solidified, the edges (i.e., outermost few millimeters) represent zones that were partially silicified, incorporating a mix of silica (now chert) and carbonate material.

Very early silica-nodule growth often resulted in the displacement of surrounding matrix (Ragland, 1983). For example, primary sedimentary laminae were distorted in all dimensions if growth was rapid and roughly equidimensional. Alternatively, if growth was slow, the nodules may have been modified by gravity as the nodule accumulated, resulting in preferential sideward, and, sometimes, downward growth. Concurrent overlying sedimentation may also have contributed to the flattening of nodules, producing pseudolayers and minimal matrix distortion.

In addition to silica precipitation, replacement of carbonate probably occurred, if the pH levels were low enough and dissolved CO_2 concentrations high enough to cause simultaneous calcite dissolution and silica precipitation. Replacement of carbonate micrites in such circumstances is difficult to confirm visually, as 1:1 replacement of microcrystalline carbonate by microcrystalline silica reproduces the original texture at the microscopic level. However, such a mechanism is more clearly proven in circumstances where fossil and ooid replacement has taken place.

The early nature of silica-rich interstitial fluids is also shown by the cementation of fenestrate in some stromatolites. In such cases, minor amounts of chalcedony and very small (~0.1 mm) megaquartz crystals with wavy boundaries preserve the pre-compaction fenestral fabric (Fig. 11A).

As a result of penecontemporaneous silica cementation and replacement, the true thicknesses of the sediments may have been more closely preserved. For example, some partially silicified stromatolites show compactional draping of unsilicified layers against silicified layers. Although initial differences in relief and growth rates may be responsible for such draping, the coincidental occurrence at the edge of silicified zones suggests differential compaction in unsilicified sediment.

We speculate that high concentrations of silica in interstitial fluids moving just below the sediment surface led to comprehensive replacive silicification of some sediments, as well as the infill of much porosity by chalcedony and megaquartz. One of the most interesting illustrations of this process involves a rare occurrence of free-floating silica spherules (Fig. 12). In this case the spherules occur in small vugs in a silicified stromatolite. The architecture of the silica spherules resembles that of calcitic ooids found within algal vugs on the Gavish Sabkha along the Gulf of Aqaba (Gerdes and Krumbein, 1987). Particles of micrite or previously silicified micrite may have acted as nuclei around which chalcedonite fibers grew radially, as the particles were gently tumbled in silica-rich pore fluids by tides and currents. The free rotation of the particles probably did not last too long as their free edges would have been attracted to chalcedonic travertine-like growth from the side walls of the vugs. As the critical size for suspension was surpassed, the spherules dropped into the lowest parts of the vugs. Silica growth in the form of quartzine fibers continued around the free edges of the vug. Eventually cementation was completed by a final infill of megaquartz crystals. Complete silification appears to have been a single event as no apparent breaks occur between the chalcedony, the later elongate megaquartz crystals, and the final equant megaquartz. The spherules do not appear to be a replacement product as the progression from flamboyant, elongate crystals to more equant crystals follows the edges of the vugs and the edges of the spherules. Assuming a sufficient source of silica, adequate fluid movement, and high photosynthetic rates (Epstein and Friedman, 1982), silica could have dissolved in one section of the algal buildup, migrated to another section (probably very close), and thus partially silicified the buildup. The interstitial fluid, supersaturated with respect to silica, would have provided a microenvironment where silica spherules could accrete.

In some cases, particularly in the Cool Creek and West Spring Creek Formations, early silicification has preserved traces of, and pseudomorphs after, evaporites. Inferentially, early mobilization of silica in freshly buried sediments replaced evaporite nodules, in some cases preserving the textures or trapping trace amounts of the minerals (Fig. 11C). Pseudomorphed textures include fine and coarse laths after anhydrite, blades and discoidal crystals after gypsum, alabastrine gypsum rosettes, and hopper-shaped crystals resembling halite. When silica was not available, evaporites presumably were flushed from the system with slightly freshened ("normal marine") ground water.

One potent control on chert formation may have been the bacterial oxidation of organic matter (mostly algal) in near-surface conditions with resulting increases in pCO_2. This would have re-

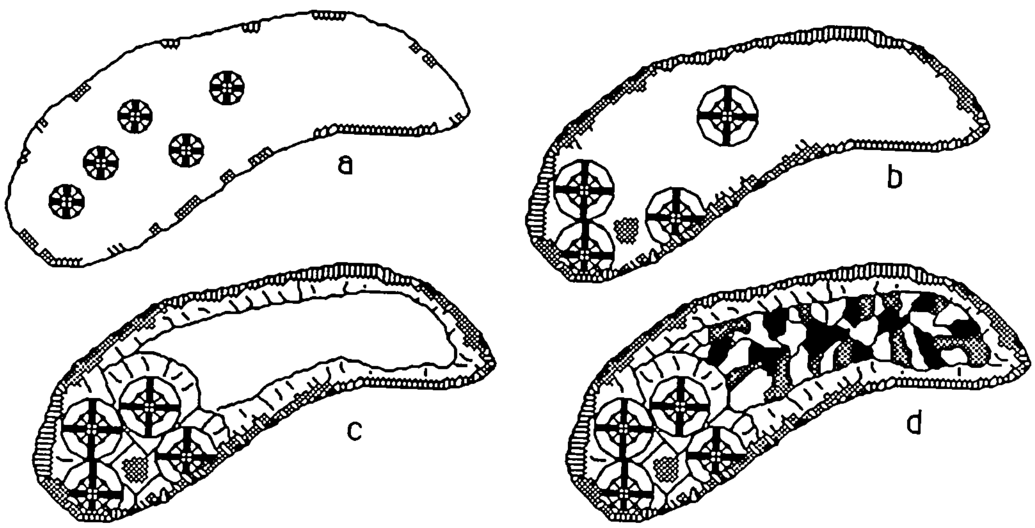


Figure 12. Stages in the infill of a void by a variety of secondary silica textures: a—free-moving spherules of chalcedony; b—gravitational settling of the spherules to bottom of void and growth of pore-lining chalcedony; c—growth of pore-lining fibrous chalcedony; d—final infill of void by equant megaquartz.

sulted in the dissolution of carbonates and early replacement by silica if a source was available (Holdoway and Clayton, 1982). Assuming that such a mechanism was operative, high concentrations of mobilized silica would have replaced carbonate morphologies on a nearly 1:1 basis (Holdoway and Clayton, 1982). Differences in permeability and porosities of the material being replaced could have resulted in preferential silicification of some components of the sediment or some parts of grains. Variations in the controlling mechanisms (e.g., availability of silica, pH, pCO_2, etc.) could have led to multiple stages of silicification and cementation.

Near-surface silicification and other diagenetic processes may have been aided by the physiography and climate of the aulacogen. The shallow water which covered vast areas of submerged continent presumably would have responded in its entirety to fluctuations in temperature. Thus, interstitial waters in near-surface sediments would have remained very warm on a constant basis, aiding in dissolution of silica and precipitation of calcite. Changes in other parameters would have allowed alternating precipitation and dissolution of minerals. If Knauth and Epstein's (1976) estimate of a temperature of 34°C for sea water over the North American continent in the Early Ordovician is correct, interstitial waters may have been much warmer, especially during the paleosummer.

A modified version of Knauth and Epstein's (1976) mixing zone model may have been the mechanism by which many of the pseudolayers of chert were formed. During low stands, emergent areas of the platform, and even restricted lagoons, would have been influenced (i.e., freshened) by meteoric water (rain). On the great Arbuckle platform we envisage that lenses of meteoric water would have developed at both platform margin and island sites. Meteoric water filtering down and mixing with modified interstitial sea water may well have promoted both silicification and dolomitization in intertidal and supratidal facies. This mechanism of silicification would have been particularly potent during periods of progradation and regression, and may well account for those occurrences of chert nodules which can be traced over great distances. The general model for mixing of this type is well established; we suggest that it may be possible to apply it to the Arbuckle section in a quite detailed fashion. Minor rain storms may not have contributed enough fresh water to initiate silicification, but large storms (e.g., hurricanes) would have produced sufficient amounts of rain to percolate into the sediments. Smaller storms between the large ones would have reinforced the fresh-water content. The thin, flattened pseudolayers of silica nodules may thus represent packages of "rain events" in much the same way as hardgrounds in the Bahamas are thought to represent packages of events (Hine and others, 1981). The generally silica-free units between pseudolayers may represent "dry seasons" complicated, of course, by a variety of other factors not necessarily related to seasonality (e.g., availability of quartz sand as a source). Once formed, the pseudolayers would have acted as permeability barriers, partially preventing vertical fluid movement and thus restricting the influx of meteoric water, except through breaks in the pseudolayers.

Early carbonate cementation in the unsilicified layers would also have slowed percolation at the onset of the next rainy season.

Textural evidence suggests that syntaxial quartz overgrowths began forming early in the cementation history of the Arbuckle Group, i.e., in near-surface conditions (Fig. 11B). If the overgrowths were a late diagenetic feature, carbonate cements, by occluding permeability, would have inhibited such overgrowths (Maliva, 1989). In some cases, multiple generations of alternating silica and calcite precipitation reflect rapid changes in pore-water chemistry that occurred during early diagenesis (Ragland, 1984). Silica growth occasionally continued long enough to replace and incorporate parts of carbonate grains, such as ooids, before carbonate cementation finally sealed the rock (Fig. 11B).

CONCLUSIONS

This study of silica in the upper Arbuckle Group has recognized that both detrital quartz and varities of secondary silica (chert) are important features of the sequence. The detrital quartz sand populations are mineralogically mature and consist of well-rounded grains. Sorting reflects the reworking into a variety of shallow-marine settings on the Arbuckle carbonate platform. In general terms, more siliciclastic detritus is found in the Cool Creek and West Spring Creek Formations than in the Kindblade; this distribution may reflect the fact that the Kindblade was a period of general transgression, whereas the other two formations were deposited during relative low stands when greater areas of the craton were exposed and more sand was available.

Two of the most distinctive bodies of quartz sandstone in the upper Arbuckle Group represent two different morphologies and processes, but each marks a significant change in their respective environments. The Thatcher Creek sandstone Member is the ultimate reflection of an influx of sand associated with regional shallowing of marine waters during the time of deposition of the uppermost McKenzie Hill Formation and lower Cool Creek Formation. As quartz sand was introduced to the area of the aulacogen, possibly by eolian or fluviatile processes, tides, waves, and storms reworked the sands. The sands were distributed in greatest abundance in the subtidal environment and in decreasing quantities in intertidal through supratidal zones. The quartz-rich sections (i.e., subtidal zones) were deposited as wave-controlled sheet sands. This environment was not entirely siliciclastic, but a mixed carbonate/clastic mixed system deposited on a shallow platform.

During a transgression beginning in latest Cool Creek times, the deposition of massive micrites and boundstones was briefly, but convincingly interrupted by an influx of quartz sand. This sand was reworked and redeposited as sand bars or ridges, one of which remains preserved as the Herringbone sandstone member. Our interpretation of this member suggests that quartz sand was spread throughout the subtidal zone by ebb and flood tides molding some of the sands into bipolar cross beds.

Bedded cherts are an infrequently occurring but environmentally significant facies in the upper Arbuckle Group. In these beds, microcrystalline chert alternates in thin laminae with microcrystalline carbonates. We suggest that the cherts probably were deposited in conditions similar to those of the modern Coorong district of southern Australia. The chert/carbonate couplets may represent fluctuations in algal bloom and decay associated with seasonal changes in climate.

Almost every texture of secondary silica, including equant microcrystalline chert, fibrous chalcedony, and megacrystalline quartz, is present in the upper Arbuckle Group. Nodules of various sizes are found from the middle McKenzie Hill Formation upward. Very early silica dissolution, precipitation, and replacement may have been linked to abundant algal growth as photosynthesis caused fluctuations in interstitial fluid chemistry. Preservation of traces of evaporites and pseudomorphs after evaporites by silica have facilitated interpretation of paleoenvironments. In addition, early silicification probably also occurred when marine and meteoric waters mixed in a variety of emergent or near-emergent settings.

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THE SADDLE MOUNTAIN DOLOMITE AT MAUKEEN HILL— A DOLOMITE INTERFACE IN THE UPPER ARBUCKLE GROUP, WESTERN SLICK HILLS, OKLAHOMA

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ABSTRACT.—The Saddle Mountain Dolomite is a newly recognized interval of comprehensive dolomitization in the western part of the Slick Hills in southwestern Oklahoma. The interval that has been dolomitized is within the upper Arbuckle Group and includes the topmost part of the Cool Creek Formation, most of the Kindblade Formation, and part of the West Spring Creek Formation. Contacts between dolomite and original limestone are well exposed on Maukeen Hill, where the dolomitization front follows an irregular course across the hillside through a stratigraphic thickness of 1,800 ft. Contacts between limestone and dolomite are knife-sharp. Fabrics in the dolomite are complex and display considerable variation in crystal size and shape. Cathode-luminescence studies indicate that at least two periods of dolomitization have occurred. For the most part porosity values are low, but in some areas as much as 15% intracrystalline porosity has been preserved.

We suggest that initial dolomitization took place during a substantial marine regression at the end of Arbuckle Group deposition. At this time, meteoric fluids dolomitized the Arbuckle Group on the Texas and Oklahoma craton, while in the southern Oklahoma aulacogen the group was not altered. Within this geographic framework the Saddle Mountain deposit displays the best exposed contact between the cratonic and aulacogen mineralogies. Following burial and exposure to temperatures of ~60°C, the dolomite was modified in the subsurface, prior to uplift in the Pennsylvanian and exposure at the surface in the Permian.

The Saddle Mountain Dolomite provides an analog to the "Brown Zone" hydrocarbon reservoir. The exposures on Maukeen Hill indicate that the Brown Zone is not an easy target, particularly in areas of great structural complexity.

INTRODUCTION

With the exception of the upper part of the West Spring Creek Formation, the entire Arbuckle Group is well exposed in the Slick Hills of southwestern Oklahoma (Fig. 1). One of the more impressive features of the outcrop is the occurrence of thick sequences of dolomite at two stratigraphic levels: (1) near the top of the Cambrian Fort Sill Formation; and (2) in the Ordovician uppermost Cool Creek, Kindblade, and lower West Spring Creek Formations. The latter occurrence is at the approximate stratigraphic level of the "Brown Zone," a prolific, though enigmatic, reservoir in the Anadarko and Ardmore basins.

A feature of especial interest is that neither dolomite horizon persists throughout the Slick Hills; in fact, over most of their outcrop the stratigraphic intervals concerned are represented by unmodified platform limestones. The dolomite near the top of the Fort Sill Formation—the Bally Dolomite (new name)—is found only at Bally and Zodletone Mountains in the northwestern part of the Slick Hills (Donovan and others, 1986a) and at McKenzie Hill on Fort Sill (to the south of the Wichita Mountains). The contact between the dolomite and unaltered Fort Sill Formation found elsewhere in the Slick Hills is not exposed. The stratigraphic horizon is similar to that occupied by the Royer Dolomite in the Arbuckle Mountains, but differs in that the topmost part of the Fort Sill is only partly dolomitized and none of the Signal Mountain Formation has been affected.

Large-scale dolomites in the upper Arbuckle Group are found only in the northwestern part of the Slick Hills, specifically on Longhorn and Unap Mountains and in folded terrain just to the north of the Meers Valley—the Saddle Mountain Dolomite (new name). The great significance of the last-named locality is that the contact between dolomite and limestone is well exposed on Maukeen Hill (Fig. 2). Here the dolomitization front can be studied through ~1,800 ft of section as it bends around the Saddle Mountain syncline (Donovan and others, 1989). In this paper we describe the field relations of this dolomite. Geochemical and isotopic parameters of the dolomite are addressed in a complementary paper by Gao and Land in this volume.

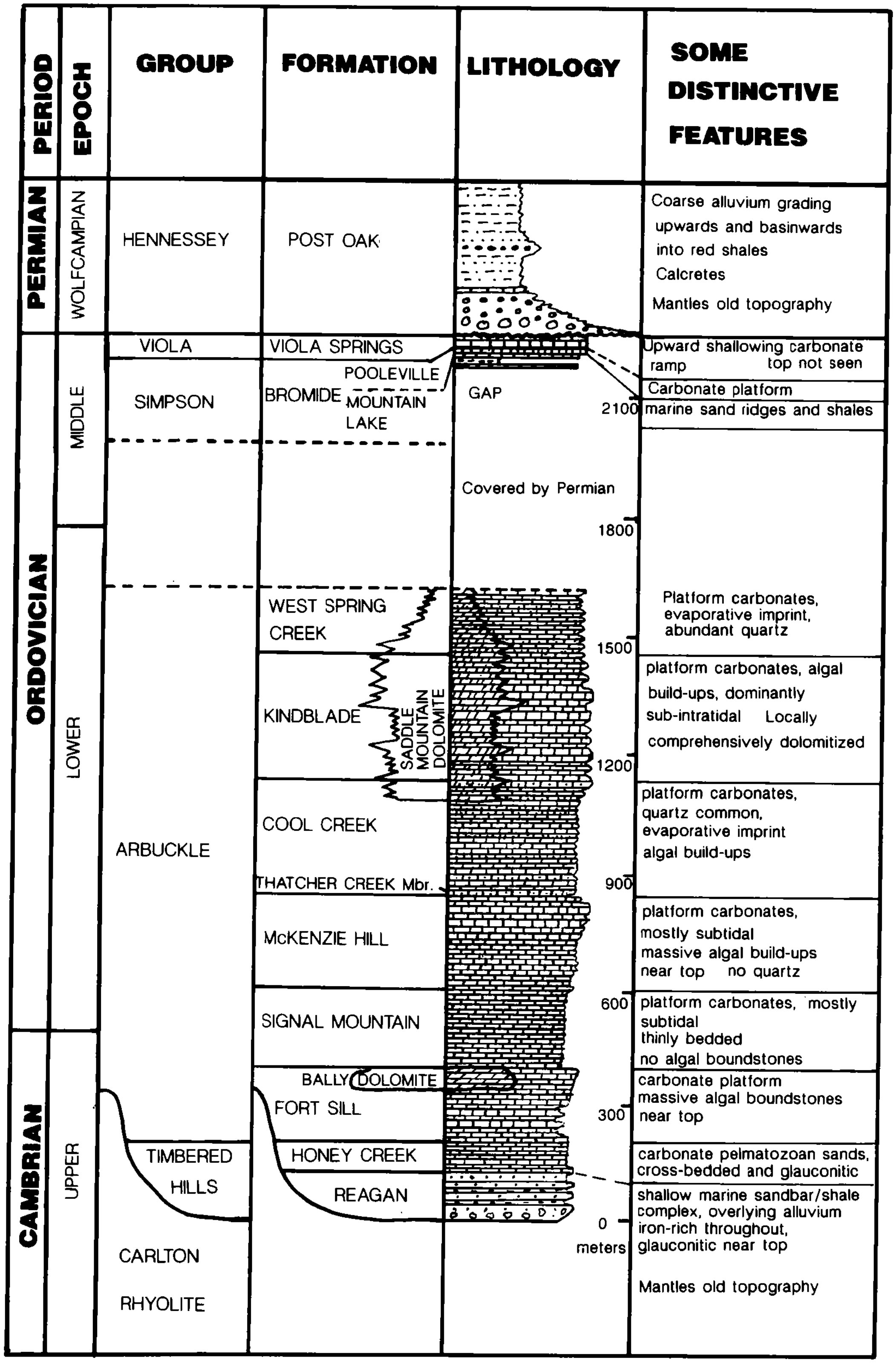

Figure 1. Stratigraphic log of the Paleozoic section that is exposed in the Slick Hills, showing the position of the Saddle Mountain and Bally Dolomites.

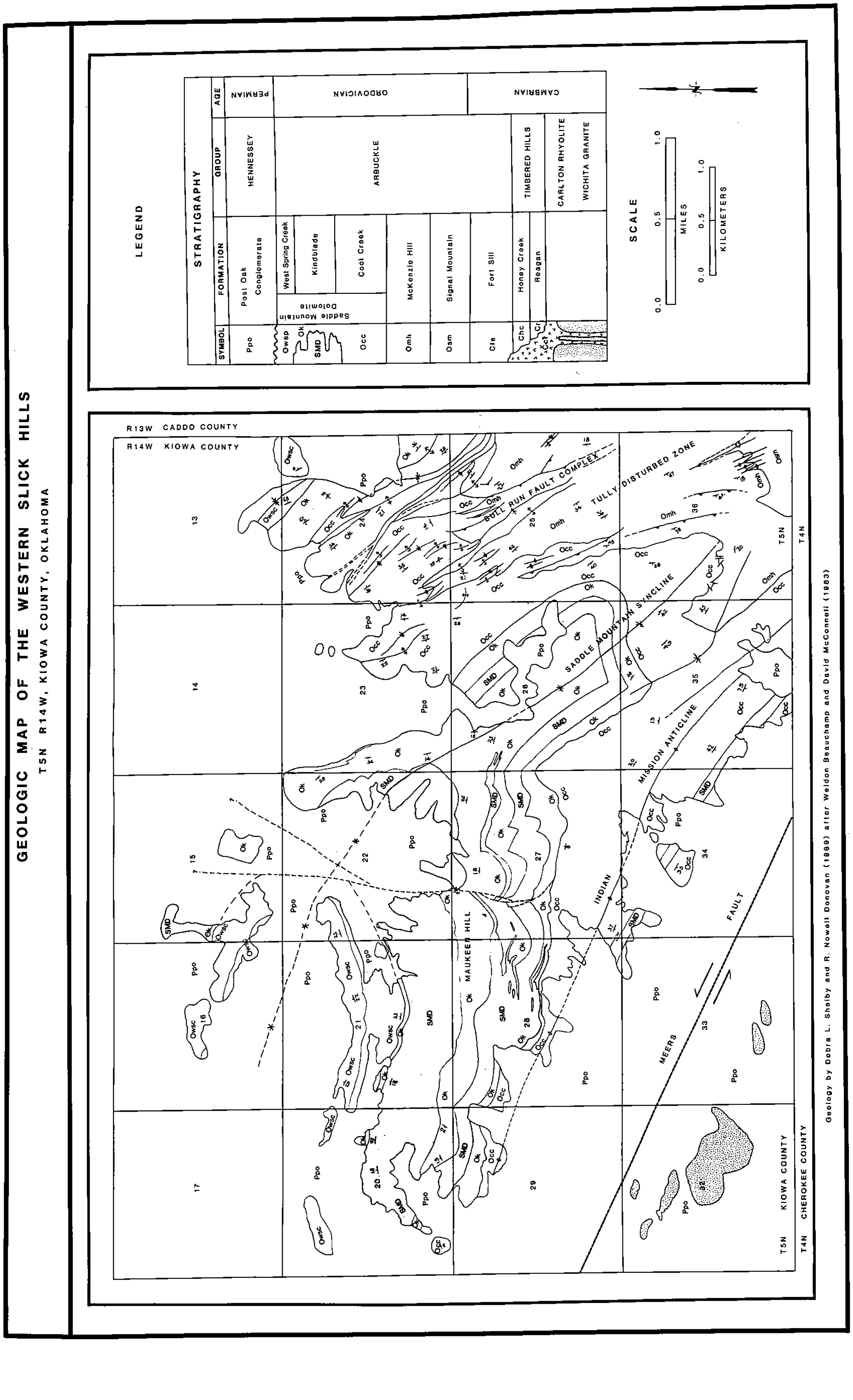

Figure 2. Geologic map of the western Slick Hills showing the outcrop of the Saddle Mountain Dolomite.

THE SADDLE MOUNTAIN DOLOMITE

In the area of Maukeen Hill the upper part of the Arbuckle Group is folded by the Indian Mission anticline and Saddle Mountain syncline (Fig. 2). To the north, south, and west, these folds are partly buried by Permian rocks (the Post Oak Conglomerate). To the east the Arbuckle outcrop is contiguous, comprising tectonically complicated relationships involving the lower part of the group.

The maximum interval of dolomite observed occupies the interval from the top of the Cool Creek through the lower part of the West Spring Creek Formation. Not all of this thickness has been dolomitized; the central part of the Kindblade and several zones of the West Spring Creek have not been altered.

In its pristine state the upper part of the Arbuckle Group consists of generally thinly bedded platform carbonates arranged in meter-scale upward-shallowing cycles. These upward-shallowing cycles mostly are composed of limestone but do contain a number of thin (as much as 5 ft) beds of fine-grained dolomite. Some of this dolomite has evaporitic connections (Cloyd and others, 1986), and some does not (Donovan and others, 1986b): it all appears to be very early, if not penecontemporaneous. Replacement chert is common in the form of nodules of various types. Both dolomite and chert are more commonly found in the Cool Creek and West Spring Creek Formations than in the Kindblade. Throughout all three formations the original calcitic textures are highly variable, comprising a potpourri of intraformational conglomerates, algal boundstones, mudstones, and a spectrum of clastic carbonates ranging from grainstones (some of which are oolitic), through packstones to wackestones. In other words the primary lithologies are diverse, both with respect to grain size and composition.

Contacts between unaltered Arbuckle strata and the Saddle Mountain Dolomite are very sharp; in most cases calcite is completely replaced within a few millimeters. Silica (chert and quartz grains) is unaffected by the replacement. While many contacts are parallel to bedding for long distances, others are highly discordant, particularly on Maukeen Hill (Figs. 3,4).

Grain size in the Saddle Mountain Dolomite varies from micrite to medium-coarse sparite (1 mm). This variation is largely inherited from the precursor calcite crystal size and may be finely tuned. For example, dolomitized algal boundstones frequently exhibit a bimodal crystal size; finely crystalline within relict algal laminae, and medium to coarsely crystalline between algal heads. In essence the dolomite convincingly mimics the precursor limestone. This can be seen at outcrop scale, where it is easy to recognize the original carbonate cycles, as well as under the microscope, where some dolomite fabrics are characterized by abundant ghosts.

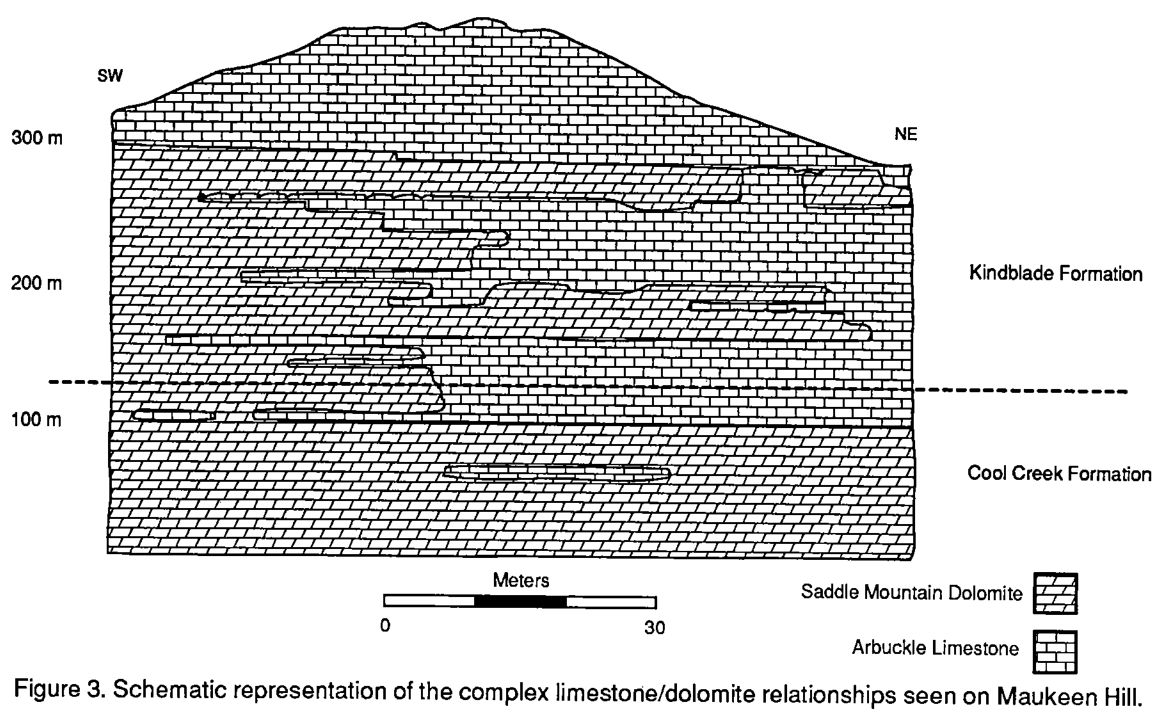

Figure 3. Schematic representation of the complex limestone/dolomite relationships seen on Maukeen Hill.

Figure 4. View of lower part of Maukeen Hill showing dolomite/limestone contact in the lower part of the Kindblade Formation.

Most crystal fabrics in the Saddle Mountain Dolomite are replacive; a small amount of cementing dolospar partly fills vugs and veins. A wide variety of grain sizes and textures is found. Thus anhedral, subhedral, and euhedral mosaics are all present (and may grade into each other), many crystals show undulose extinction (the degree of "sweep" varying from slight to extravagantly baroque), and textures may be inclusion-free or inclusion-rich (in some cases within the same crystal). In the terminology of Gregg and Sibley (1983) the most commonly seen crystal fabrics are idiotopic-S and xenotopic-A; these two fabrics grade intimately into each other, and the latter is slightly coarser than the former (Fig. 5). In addition, in some areas idiotopic-E fabric is found; this fabric grades into idiotopic-S and is generally associated with intracrystalline porosities of up to 15% (Fig. 6). Some xenotopic-C fabric (i.e., baroque) is found in association with veins and vugs (Fig. 7).

Clearly a wide and somewhat confusing spectrum of crystal types is present in the dolomite; cathode-luminescence studies confirm this complexity. In general, idiotopic fabrics tend to fluoresce more brightly than do xenotopic, but the reverse relationship may also be the case. In some areas it is clear that xenotopic textures postdate idiotopic and may even have a syntaxial relationship with the latter. The intimacy of variation revealed by the fluorescence within a single thin section suggests that whole-rock isotope or geochemical analyses are unlikely to be helpful in determining the genesis of the dolomite. Staining suggests that some of the variations in fluorescence are due to variations in iron content (the more quenched dolomite being slightly richer in iron).

The Slick Hills are recently exhumed Permian topography, formerly covered by the Post Oak Conglomerate, a locally derived Lower Permian alluvial facies. Boulders and pebbles of Saddle Mountain Dolomite are widespread in the Post Oak and provide a clear *terminus ad quem* for the timing of dolomitization. Traces of Permian exposure are manifest as limited dedolomitization (to nonferroan calcite and iron oxide) and small vadose fissure infills of speleothemic calcite.

DOLOMITIZATION IN THE ARBUCKLE GROUP

Early Dolomitization

As noted above, the earliest dolomites found in the Arbuckle Group are thin dolomicrites that occur in some of the meter-scale ("mini") upward-shallowing cycles which typify so much of the group. Dolomites of this type are usually associated with the shallowest facies of such cycles; they comprise a spectrum of lithologies, grading from dolomitic limestone through calcitic dolomite, to true dolomite. In most cases dolomitizing fluids followed available permeability paths (e.g., burrows in the case of "leopard skin texture," Donovan and others, 1986b). Dolomite crystals are euhedral or subhedral (i.e., idiotopic), micrite or microspar. Porosity in these dolomites approaches zero. An early age for dolomitization of this type can be inferred from three lines of evidence:

1) Where present, dolomitization is consistently tied to the shallowest water facies of the "mini cycles," suggesting a tie-in to contemporaneous water tables.

2) Dolomite is fabric selective and apparently predates lithification.

3) Some intraformational conglomerates contain clasts of dolomicrite associated with undolomitized clasts and matrix (while it is possible that such clasts may have been suitable sites for fabric-selective dolomitization, it is more likely that they were eroded from already existing dolomites).

Early dolomites of this type are found throughout the Arbuckle Group, but are most common in the Cool Creek and West Spring Creek Formations. Two basic types of dolomitic facies can be distinguished: with or without evidence of an evaporative imprint (Ragland and Donovan, 1986; Donovan and others, 1986b). The former facies is characterized by: (1) pseudomorphs after gypsum, anhydrite and, rarely, halite (Ragland and Donovan, this volume); (2) relict traces of anhydrite and celestite in cherts; (3) cauliflower cherts with lutecite, length-slow chalcedony, and flamboyant megaquartz, that have been interpreted as a replacement of anhydrite (Cloyd and others, 1986); and (4) collapse breccias, thought to record the near-surface dissolution of evaporites (Donovan and others, 1988).

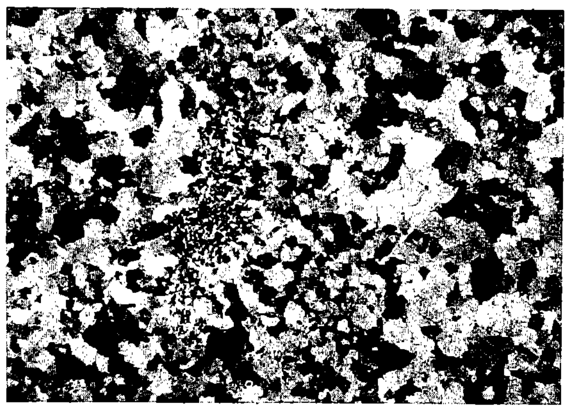

Figure 5. Saddle Mountain Dolomite: complex sparry dolomite that shows very little porosity and consists of an intergrowth of both idiotopic and xenotopic crystals. There is a small chalcedony nodule in the center of the picture. Polarized light, field of view 8 mm wide.

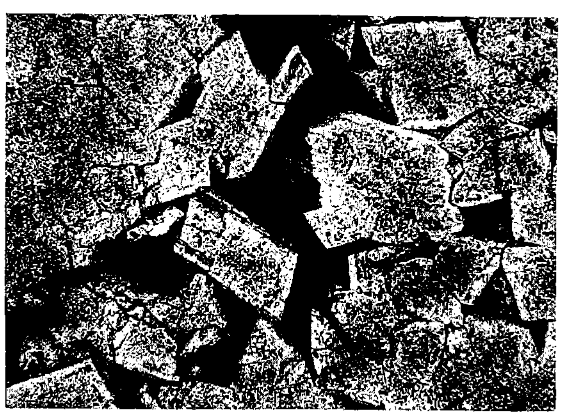

Figure 6. Saddle Mountain Dolomite: euhedral dolosparite with well-developed intracrystalline porosity. Ordinary light, field of view 4 mm wide.

Figure 7. Saddle Mountain Dolomite: vein-cutting dolomicrite (left of view). Initial crystals lining vein wall are baroque, later crystals (right of view) are idiotopic and show zones of inclusions. This sequence is interpreted as a record of deposition from brines that were variable in composition and slowly cooling with time. Polarized light, field of view 4 mm wide.

Where such evidence is present, it is likely that a sabkha-type model for dolomitization is applicable. This suggestion is buttressed by the occurrence of nodular anhydrite in both the Cool Creek and West Spring Creek Formations (K. S. Johnson, personal communication). Where evaporative evidence is lacking, it is possible that dolomitization resulted from ground-water mixing during low stands.

Formation of the Saddle Mountain Dolomite

Dolomitization on the scale exhibited by the Saddle Mountain Dolomite must have involved the movement of fluids on a scale which is an order of magnitude greater than that of the penecontemporaneous dolomites described above. Furthermore, the complexity of the petrographic textures in the dolomite suggests that it has been subject to more than one phase of dolomitization. The problem of dolomite formation resolves into several related questions: (1) the position of the Saddle Mountain area within the southern Oklahoma aulacogen, (2) the timing of initial dolomitization, (3) the mechanism of that dolomitization, and (4) the diagenetic history of the deposit.

Position of the Saddle Mountain Area within the Southern Oklahoma Aulacogen

The Saddle Mountain area lies on the southwestern flank of the southern Oklahoma aulacogen, on the northern edge of the Wichita uplift. Although the character of the Arbuckle Group in the Hollis/Hardeman basin to the south of the Wichita uplift is poorly known, it is clear that, in general, it becomes increasingly dolomitic as it merges into the dolomites of the Ellenburger Group in Texas. It has been noted that, whereas most of the Arbuckle Group within the confines of the southern Oklahoma aulacogen consists of limestone, it is generally dolomitized on the surrounding craton (e.g., Johnson and others, 1989). Schematic representations of the passage from limestone to dolomite suggest that, within a zone on the periphery of the aulacogen, the two lithologies must interfinger. In this context it is apparent that the Saddle Mountain Dolomite represents by far the best exposed example of such an interfingering. If this is the case then it follows that the mechanism for dolomite formation must be compatible with mechanisms which can account for the widespread cratonic dolomitization of the Arbuckle Group. It also follows that, as the basic pattern of dolomitization is related to the shape of the depocenter (i.e., the contours of the aulacogen), then the mechanism producing that dolomitization was probably early and predates dismemberment of the aulacogen by Pennsylvanian tectonism.

Timing of Initial Dolomitization

The timing of initial dolomite formation is clearly constrained by the small-scale penecontemporaneous dolomites and the Permian conglomerates discussed previously. Furthermore, although the limestone/dolomite contacts on Maukeen Hill are discordant to bedding, they are not discordant to structure (in particular folding). Hence, the major dolomitization predates Pennsylvanian tectonism associated with dismemberment of the aulacogen.

What little dolomite is spatially related to structure comprises quantitatively insignificant areas of coarse baroque textures which appear to be controlled by faults and fractures. Post- or syntectonic dolomites of this latter type are not restricted to the Saddle Mountain Dolomite but can be found through much of the Slick Hills (e.g., Cloyd and others, 1986).

This field evidence, together with the basic pattern of Arbuckle dolomitization noted above, appears to eliminate the potentially attractive solution that initial dolomitization was accomplished by brines derived from siliciclastic sediments in the Pennsylvanian/Permian Anadarko basin. While there is ample evidence that a variety of fluids (including those responsible for fracture-related dolomitization) has found and continues to find conduits to the surface through the Slick Hills (Donovan, 1986; Younger and others, 1986), the folding of the dolomite clearly indicates pre-Pennsylvanian replacement. However, it is possible that fluids derived from the developing Anadarko basin could have modified existing dolomite prior to major folding and uplift.

In summary, field evidence suggests that initial dolomitization was not penecontemporaneous but probably took place early enough in the history of the deposit to permit extensive diagenetic modification during pre-Pennsylvanian burial. Dolomitization was completed by the end of the Pennsylvanian or Early Permian, by which time the Slick Hills were thrust northward, structurally emplaced above the Anadarko basin, and exposed at the surface. At this time a phase of hydrocarbon migration from the basin ensued. The evidence for timing and geometry of this migration is provided by hydrocarbon contamination of speleothems in karst which can be securely dated on the basis of Permian vertebrate faunas (Donovan, 1987). This karst is found only a few miles from Maukeen Hill, hence it is not surprising that some of the intracrystalline pores in the Saddle Mountain Dolomite contain traces of biodegraded hydrocarbon.

Mechanism for Initial Dolomitization

In order to produce dolomites of the size of the Saddle Mountain, a mechanism which can flush the upper Arbuckle with large volumes of fluids is

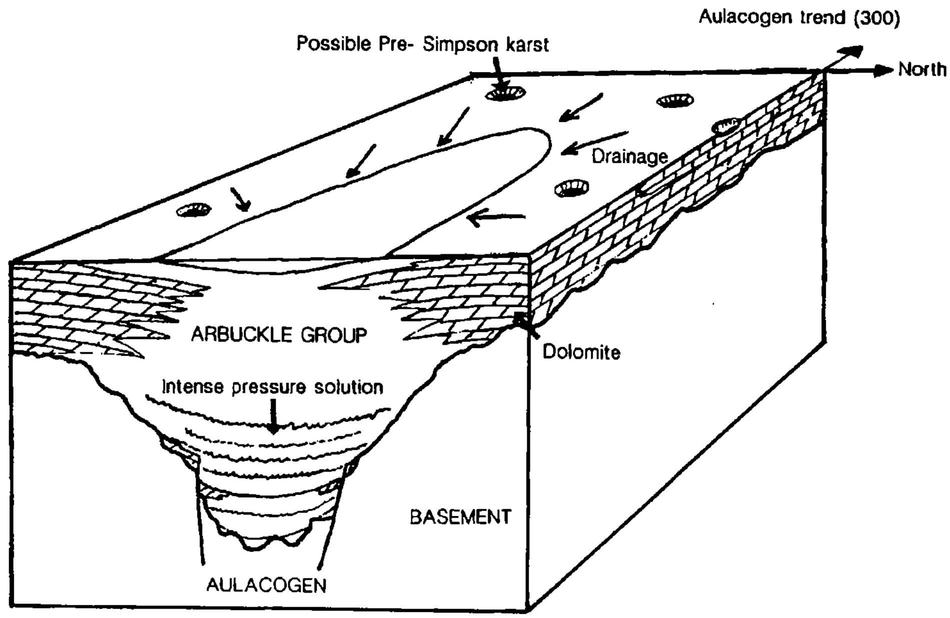

Figure 8. Model illustrating possible geographic and hydrologic relationships that existed during the post-Arbuckle/pre-Simpson regression. At this time the aulacogen is envisioned as a gulf controlled by continuing subsidence along the axis of the aulacogen and augmented by pressure solution in the underlying carbonates. Arrows indicate surface drainage and possible ground-water flow paths.

required. In the present context, the most attractive such mechanism is provided by the large-scale movement of meteoric waters related either to sea-level falls or to relative uplift. Within the geographic framework of the aulacogen and adjacent craton, the mechanism requires that thinner cratonic sequences were substantially dolomitized while much thicker aulacogen sequences were unaffected, except for topmost sequences near the margins of the embayment (Fig. 8). This mechanism also requires that undersaturated ground water in the Arbuckle Group within the aulacogen was not displaced. The implication here is that the negative relief of the aulacogen was enhanced relative to the adjacent craton; in other words, ground water within the aulacogen was immobile because of its low position relative to the craton. This relationship requires some analysis, because there is little or no evidence of differentiation in water depth (bathymetry) within the Arbuckle Group that can be related to the geography of the aulacogen. Thus, in the case of the topmost West Spring Creek Formation, shallow-water carbonates were deposited across the entire craton and aulacogen (the conclusion here is that short-term sedimentation rates on the Arbuckle platform were greater than subsidence rates in both the aulacogen and craton).

It appears that bathymetric differentiation of the aulacogen and craton began after the Arbuckle carbonate platform had ceased to function, as can be inferred from facies patterns in the overlying Simpson Group (R. E. Denison, personal communication). Opinions vary as to whether or not the aulacogen was subject to tectonic modification at this time. Thus Longman (1981) considered that facies patterns within the aulacogen were to some extent controlled by tectonism, while McPherson and others (1988) highlighted the role of sea-level fluctuations during Simpson Group deposition. As there is no evidence that would support the contemporary uplift of the craton relative to sea level (as opposed to uplift relative to the aulacogen), we suggest that the substantial global fall in sea level at the end of the Lower Ordovician (Vail and others, 1977) resulted in the flushing of the craton by meteoric water and was the primary cause of dolomitization of Saddle Mountain type. In southern Oklahoma this fall is heralded by facies patterns within the West Spring Creek Formation which show an increasing cratonic influence. Thus, within this formation small-scale ("mini") cycles contain increasing amounts of cratonic quartz, significant evaporative imprints, and even red shales. Both Longman (1981) and McPherson and others (1988) have suggested that the subsequent breakdown of the carbonate platform and inundation of both craton and aulacogen by siliciclastics was a function of dramatically lowered sea levels.

The lingering question is: How did the bathymetric and hydrologic differentiation necessary to maintain unmodified ground water within the aulacogen develop? It is possible that continued differential subsidence during a fairly lengthy period of nondeposition associated with the sea-level fall may have been sufficient. In addition, pressure solution (i.e., stylolite formation) in the aulacogen may have played a role in basin differentiation. Within this context, it should be noted that Donovan and others (1989) estimated that as much as 20% of the Arbuckle section in the Slick Hills may have been removed by pressure solution parallel to bedding. Clearly much of this removal postdated Simpson deposition; nevertheless, a good deal of dissolution may have taken place in a section already as much as 7,000 ft thick by the end of the Early Ordovician.

It is possible that the initial dolomitization may have formed (or been enhanced) during later low sea-level stands, although the post-Arbuckle/pre-Simpson drop appears to have been the most profound. Alternatively, pre-Pennsylvanian tectonic uplift may have further or initially exposed the Arbuckle to dolomitizing fluids. If the latter mechanism is applicable, the most likely time for it to have operated would have been during the Middle Devonian, prior to deposition of the Woodford Formation, when the Arbuckle was locally exposed (e.g., in the Oklahoma City area). However, in the Slick Hills area there is no evidence of exposure (i.e., limited inversion) at this or any other time prior to the Pennsylvanian. Furthermore, it is pertinent to note that in the Sugar Hills, a few miles to the northwest of Saddle Mountain, the upper part of the Simpson Group and the lower 400 ft of the Viola Formation show no signs of extensive dolomitization (Donovan and others, this volume).

Diagenetic History of the Saddle Mountain Dolomite

As indicated above, our preferred scenario is that the Saddle Mountain Dolomite was dolomitized at a relatively early date, early in the Middle Ordovician. Thus, the initial dolomite probably comprised low-temperature idiotopic fabrics that crystallized in a schizohaline environment. However, as noted, the present fabrics seen in the dolomite are complex, involve xenotopic fabrics, and suggest that more than one period of dolomitization has affected the deposit. To account for this we suggest that partial recrystallization and/or syntaxial overgrowth (that filled much intracrystalline porosity) took place from the Late Ordovician to Early Mississippian, when the dolomite was buried to depths of 3,500 ft and was subjected to temperatures of ~60°C (Fig. 9). It seems likely that the existing volume of dolomite increased substantially at this time. The source of magnesium ions during this time may have been from the pressure solution of existing dolomites within the Arbuckle Group. Subsequently, in early Pennsylvanian time, as dismemberment of the aulacogen commenced, it is possible that magnesium-rich brines may have moved from the developing Anadarko basin and further modified the original dolomite.

THE SADDLE MOUNTAIN DOLOMITE, AN ANALOG FOR THE "BROWN ZONE"?

One of the most elusive hydrocarbon targets in the Anadarko and Ardmore basins is the "Brown Zone," a dolomitic reservoir located at or near the top of the Arbuckle Group. The Saddle Mountain Dolomite may well represent an exposure of this reservoir. Clearly the contacts displayed on Maukeen Hill confirm that the Brown Zone is likely to be elusive. If our analysis is correct, the zone is likely to be best developed at or near the margins of the Lower Paleozoic aulacogen. Two caveats are necessary. In the first place, dolomites within the Pennsylvanian/Permian basins are likely to have undergone further deep burial diagenesis, whereas volumetrically significant diagenesis of the Saddle Mountain Dolomite appears to have ceased by the Permian. Secondly, the original southern Oklahoma aulacogen has been subject to great tectonic dislocation during the Pennsylvanian. McCoss and Donovan (1986) and Granath (1989) have suggested that deformation was essentially the result of an application of stress that was transpressive in a left-lateral sense to the N. 60° W. (300°) trend of the aulacogen. The net result of this stress is that faults parallel to the aulacogen trend will have a component of left-lateral displacement.

On empirical evidence, McConnell (1989) has suggested that this component is approximately equal to, or slightly greater than, the compressive (i.e., vertical) component perpendicular to the aulacogen trend. This being the case it follows that, as most Pennsylvanian faults are subparallel to the aulacogen trend, the dolomite/calcite contact will be successively displaced to the northwest by an amount equal to, or slightly greater than, the vertical displacement of a given fault when traced to the north.

ACKNOWLEDGMENTS

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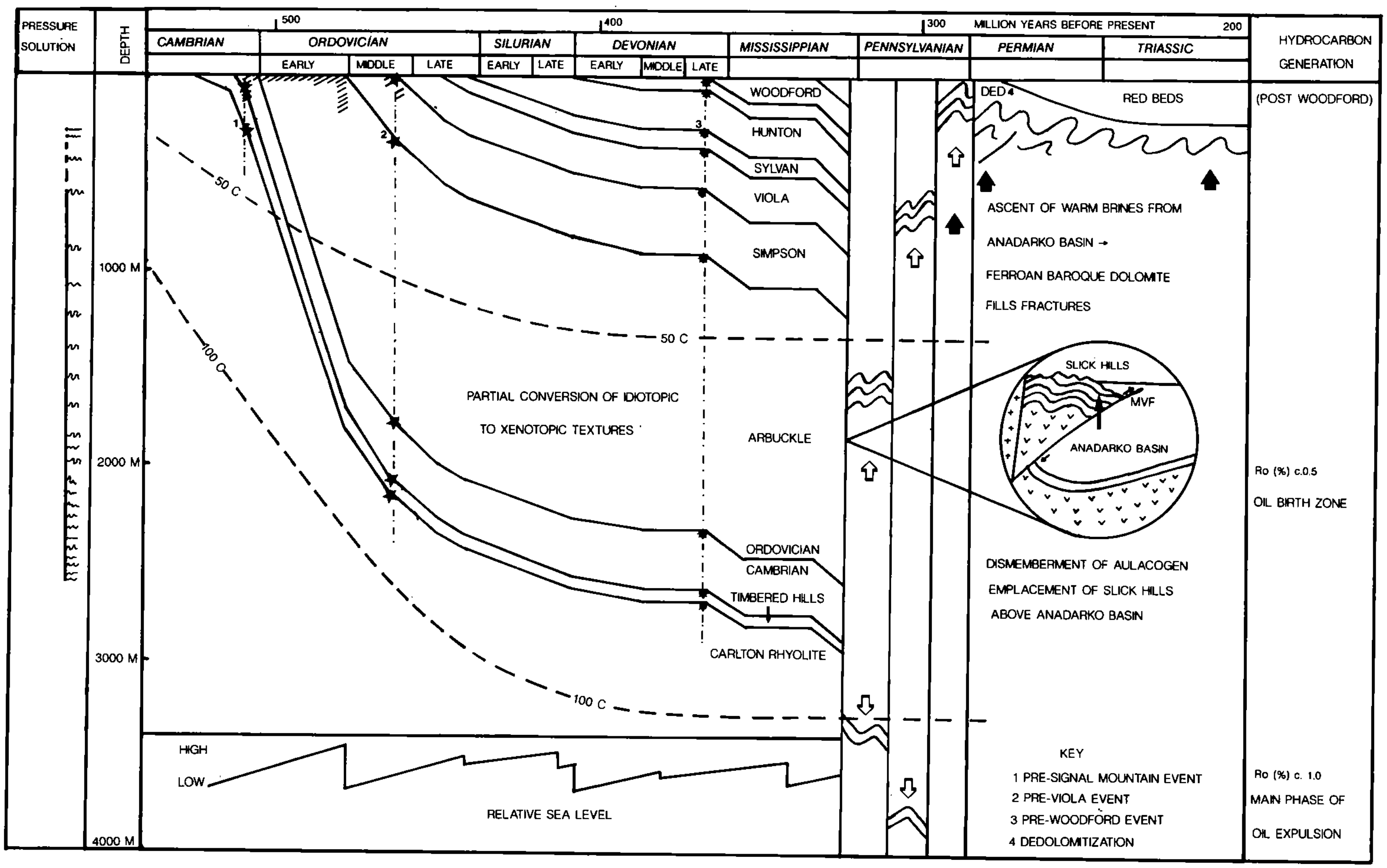

Figure 9. Time/subsidence relationships illustrating the burial history of the Saddle Mountain Dolomite. Subsidence curves (modified from Donovan, 1986) illustrate initial rapid subsidence during Cambrian–Ordovician time, a slow rate of basin development during Silurian–Mississippian time, and dismemberment of the aulacogen into two distinct parts during the Pennsylvanian and Permian (the Wichita uplift and Anadarko basin). Note that the Saddle Mountain Dolomite was uplifted at this time and in consequence has since remained at a high, and therefore cool, crustal level. The sea-level curve (from Vail and others, 1977) illustrates the significant drop in sea level that took place at the end of Arbuckle Group deposition. Vitrinite reflectance and temperature (from Cardott, 1989) show that the Woodford Formation, likely principal source rock in the area, did not mature until the dismemberment.

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DEPOSITIONAL AND DIAGENETIC HISTORY OF THE LATE ORDOVICIAN MONTOYA GROUP, SACRAMENTO MOUNTAINS, SOUTH-CENTRAL NEW MEXICO

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ABSTRACT.—The Montoya Group represents the southwestern exposure of Late Ordovician sheet-like carbonates on the North American craton. The central location of the Sacramento Mountains in the Montoya Group distribution and excellent exposures of the strata provided an opportunity to analyze and interpret the depositional fabric, diagenetic sequences, and conodont biostratigraphy of the pervasively dolomitized and silicified carbonates.

The Montoya Group consists of three formations: Second Value, Aleman, and Cutter Dolomite. The Second Value is subdivided into the Cable Canyon and Upham Members.

Montoya depositional history began after an extended erosional period. Middle Edenian transgressing seas deposited coarse-grained, quartzose sand across the south-sloping carbonate ramp. This thin sand bed, the Cable Canyon Member of the Second Value, was deposited in intertidal environments. The overlying Upham Member of the Second Value rests conformably on the Cable Canyon and was deposited in deepening, well-agitated subtidal water. Water depths continued to deepen during the Maysvillian as grain-supported carbonate sediments of the Upham Member were replaced by carbonate-mud-dominated sediments of the Aleman Formation. Most of the Aleman was deposited below storm wave base. Shoaling conditions ended Aleman deposition with the early Richmondian exposure of the carbonate ramp; however, deposition continued south of the study area. Following a short period of exposure, peritidal depositional conditions developed, and the Cutter Dolomite was deposited within the Richmond. A second exposure of the ramp ended Montoya deposition.

Diagenetically, the Montoya underwent cyclic alteration after deposition. Dolomitization and sulfate emplacement affected the Second Value–Aleman succession during early burial and were penecontemporaneous with deposition of the Cutter. Exposure of these sediments at the end of the Second Value–Aleman and Cutter depositions, respectively, resulted in silicification, dissolution, and meteoric cementation of the carbonates. A second burial further dolomitized the units and was possibly accompanied by emplacement of sulfate minerals. During this episode, recrystallization of the dolomite and cementation eliminated the early secondary porosity which had developed during subaerial exposure. Recrystallization and silicification were less pervasive in the Cutter Dolomite. Cenozoic deformation raised the Montoya from deep burial into another phase of freshwater diagenesis in both phreatic and vadose environments. Most of the porosity found in the Montoya at the surface is a product of these diagenetic processes.

INTRODUCTION

Purpose

The Montoya Group of the southwestern United States represents Upper Ordovician shallow-water marine sediments which extended across the North American craton and into the marginal miogeoclines. Intensely dolomitized and sparsely fossiliferous carbonates generally characterize the Montoya Group in outcrop and in the subsurface. Although generalized regional stratigraphic, petrographic, and paleontologic relationships of the Montoya have been presented by various workers, data are insufficient to provide a detailed understanding of depositional environments and diagenetic processes.

Approximately 135 m of Montoya carbonates and thin clastics crop out near the base of the Sacramento Escarpment in south-central New Mexico. This is the easternmost surface occurrence of the Montoya in New Mexico. Other Montoya outcrops extend westward across New Mexico and southward into west Texas. The Montoya dips into the subsurface in southeastern New Mexico and west Texas. Good exposures in the Sacramento Mountains offer an opportunity to examine the Montoya in detail (Fig. 1).

This study has three primary purposes: (1) clarification of Montoya stratigraphic nomenclature; (2) description and interpretation of Montoya depositional fabrics; and (3) description and interpretation of Montoya diagenetic fabrics.

Methods

Five stratigraphic sections were selected for measurement and sampling, based on geographic location, stratigraphic continuity, and accessibility. The sections represent a north–south transect along the trend of the Sacramento Escarpment (Fig. 1). Each section was measured, described, and sampled for petographic and paleontologic analysis. Selected samples of diagnostic features (i.e., sedimentary structures and fossiliferous beds) were also collected. The Water Station section (WS) was chosen as the reference section of the Montoya Group in the Sacramento Mountains because of its vertical continuity, exposure, and accessibility.

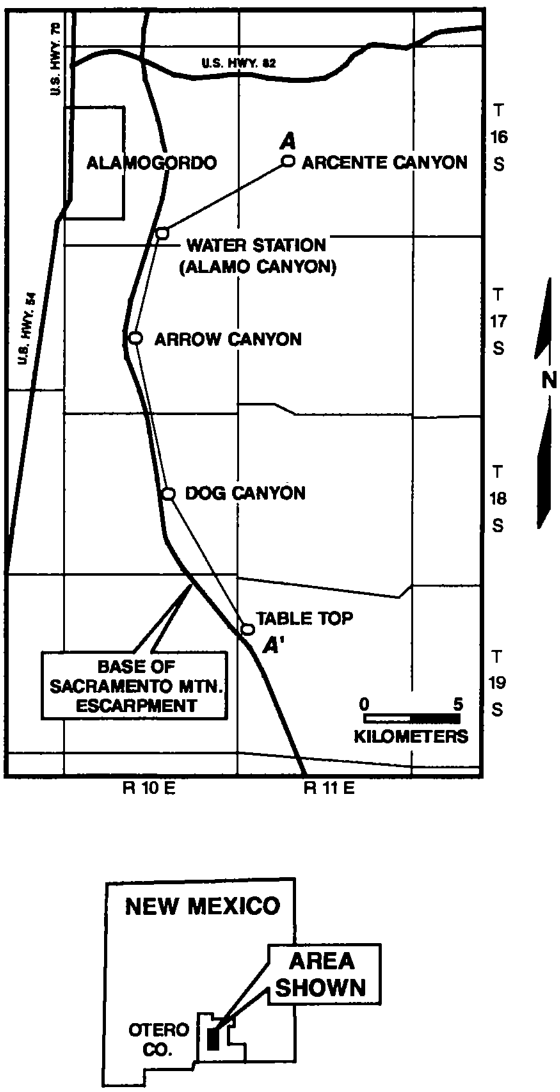


Figure 1. Index map for study area, showing measured-section localities, stratigraphic cross section A–A′ orientation, and base of west-facing Sacramento Mountain Escarpment.

Nomenclature

A host of workers have studied the Late Ordovician rocks in southern New Mexico and west Texas since Richardson (1908) did the initial work in the Franklin and Hueco Mountains, Texas. The subsequent contributors to Late Ordovician stratigraphic work on the southwest margin of the Midcontinent are Darton (1917), Dunham (1935), Entwistle (1944), Kelley and Silver (1952), Pray (1953,1958), Flower (1953,1955,1958,1959,1961, 1965,1969), Howe (1959,1964,1965a,b,1966,1967), Pratt and Jones (1961), Pratt (1967), Hayes (1975), Sweet (1979b), Measures (1985), and Evans (1985). Most of these contributors recognized the significant components of the Upper Ordovician in either a regional or local framework. The nomenclature used by these authors has not been consistent, but has defined the primary divisions within the Upper Ordovician strata of the region.

Locally, Pray (1953,1954,1961) recognized divisions within the Upper Ordovician in the Sacramento Mountains. Pray (1953) divided the Upper Ordovician rocks in the Sacramento Mountains into the Montoya Formation and Valmont Formation. Each formation had an upper and lower member. Pray's (1953) nomenclature was a departure from conventional Upper Ordovician regional nomenclature, which had named the strata the Montoya Limestone (Richardson, 1908), Montoya Dolomite (Entwistle, 1944), or the Montoya Group (Kelley and Silver, 1952). The Montoya Group is used for this discussion because the Upper Ordovician strata in the Sacramento Mountains can be divided into three mappable formations: Second Value Formation, Aleman Formation, and Cutter Dolomite (Fig. 2). The Second Value Formation is subdivided into the Cable Canyon Member and the Upham Member. This nomenclature coincides with that of similar units described by Pratt (1967) and Hayes (1975).

Regional Structural History

Orogenic episodes affected lower Paleozoic strata throughout the Paleozoic, in the Cretaceous, and in the Cenozoic. Periodic crustal warping exposed miogeoclinal deposits on a S-dipping

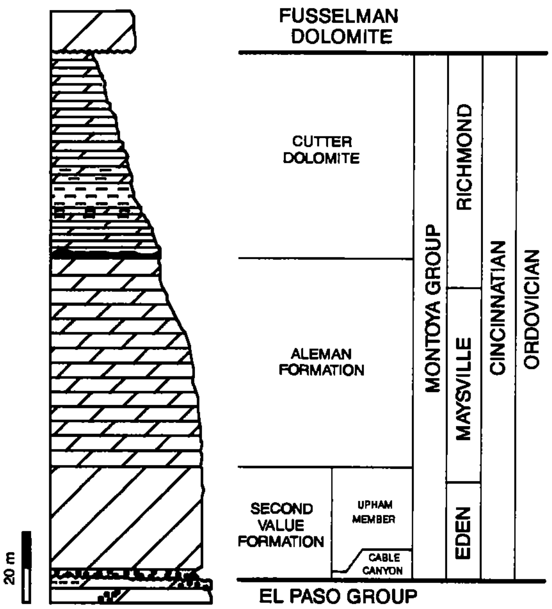


Figure 2. Diagrammatic vertical section of the Montoya Group, Sacramento Mountains, New Mexico.

ramp to erosion several times during the Paleozoic. A major structural event in the late Paleozoic truncated much of the sedimentary section. Erosion at that time had a marked effect on the regional distribution of the Montoya and other lower Paleozoic strata (Kottlowski, 1963).

Laramide compression folded and faulted the strata in the area, but the last significant deformation of the Montoya Group in the Sacramento Mountains occurred in the Tertiary. Neogene block faulting created the horsts in which the Paleozoic rocks are now exposed across westernmost Texas, southern New Mexico, and eastern Arizona.

Regional Montoya Stratigraphy

Where the Montoya has not been truncated by Paleozoic or Cretaceous erosion, the group's thickness ranges from 100 to 300 m. Erosion has removed the Montoya north of lat. 33° and west of Greenlee County, Arizona. The Montoya is also absent on the crest of a Pennsylvanian arch which extends from southwest Culbersen County, Texas, north into Otero County, New Mexico (Hayes, 1975). The Montoya thickens to the east (subsurface only) and west of the arch.

Throughout most of its distribution, the Montoya rests unconformably on the Lower Ordovician (Canadian) El Paso Formation. However, Middle Ordovician deposits between the Montoya Group and the El Paso Formation (Harding remnants) have been described at localities in the region (Flower, 1961).

The Cable Canyon Member or the Upham Member of the Second Value Formation can rest on the unconformity below the base of the Montoya. Only one or both of the members can make up the thickness of the Second Value Formation, which regionally ranges from 16 to 35 m. The Cable Canyon Member, a sandstone, can be absent, or as thick as 12 m. The massive carbonate of the Upham Member is always present, and ranges in thickness from 5 to 35 m. The two lithologies of the Second Value Formation, sandstone and dolomite, are consistent throughout the region, except that limestone is present in the southernmost area of the Second Value distribution.

The Aleman is characterized by its high chert content, mud-rich lithology, and marine fauna. According to Hayes (1975), the Aleman gradually thickens to >50 m from its northern and western truncation edges. The unit ranges from 14 to 84 m regionally.

The Cutter Dolomite is distinctly different from the underlying Aleman and Second Value Formations. The unit is a uniformly white, thin- to thick-bedded dolomite. The regional thickness ranges from 30 to >60 m, partly a result of the regional unconformity at the top of the Cutter Dolomite (top of the Montoya Group).

The Fusselman Dolomite (Silurian, possibly Upper Ordovician to Devonian; J. E. Barrick, personal communication, 1983) overlies the Montoya Group. In the few localities where the Fusselman has been removed, Devonian Oñate strata rest unconformably on the Montoya Group.

DEPOSITIONAL FABRIC OF THE MONTOYA GROUP IN THE SACRAMENTO MOUNTAINS

Upper El Paso Group Lithofacies

The uppermost beds of the El Paso Group, below the unconformity which separates the El Paso Group and the Montoya Group, are bioclastic pelletal wackestones. Crinoids and rugose corals dominate the fauna. Crinoids are disarticulated and abraded. Sandy intervals in the upper El Paso Group are typically fine-grained quartz sands which are coarser near the top of the unit. The quartz grains are similar in size and character to those found in the overlying Cable Canyon Member of the Second Value Formation. Cross-bedding occurs in the finer-grained sandstones of the El Paso. Small nodules and lenses of chert have replaced the carbonate, but chert is a minor constituent of the strata.

The disconformity between the El Paso and the Montoya is most often marked by an abrupt lithologic change from sandy dolostone to sandstone. Only a few centimeters of relief are recognizable at the contact. Thinning and thickening of the overlying Cable Canyon Member reflects low, broad

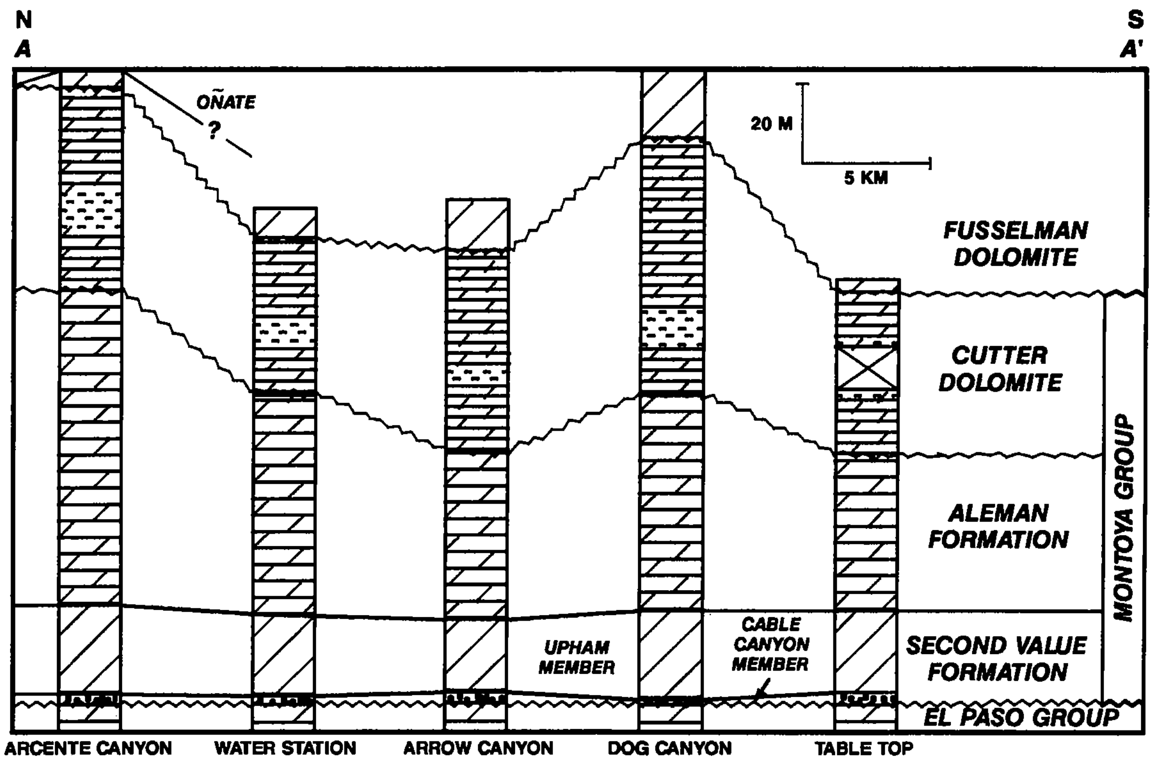


Figure 3. North–south stratigraphic cross section of the Montoya Group, Sacramento Mountains, New Mexico. Note the relatively consistent thickness of the Second Value Formation and the highly variable thicknesses of the Aleman Formation and Cutter Dolomite. Datum: base of the Montoya Group.

relief on the unconformity (Fig. 3). Stratigraphic leaks of quartz sand in fissures on the erosional surface are common. The contact of the El Paso Group and the Montoya Group is readily identifiable throughout the Sacramento Mountains.

Montoya Group Lithofacies

Second Value Formation

The basal unit of the Montoya Group, the Second Value Formation, represents initiation of deposition following a prolonged period of erosion. The Second Value Formation is easily divided into its two members in the Sacramento Mountains: the sands of the Cable Canyon Member and the massive dolomites of the Upham Member. The thickness averages 24 m in the Sacramento Mountains (Fig. 3).

Cable Canyon Member.—The Cable Canyon Member of the Second Value ranges from <0.5 m to >2 m in thickness (Fig. 3). The undulose erosional surface below probably affected the thickness of the sand. The Cable Canyon Member weathers medium- to dark-gray at the base of the massive cliff formed by the Second Value Formation. The darker color indicates the large percentage of dolomite cement and matrix between the grains of the quartz arenite. Feldspar, carbonate, and chert clasts and fossil grains are minor constituents of the relatively clean sandstone (Fig. 4A). Microcline is the only feldspar that has escaped replacement. Scarce fossil debris of crinoids and small rugose corals (*Paleophyllum*) occur in parts of the Cable Canyon Member where the mud matrix is more prominent. The sandstone is poorly sorted. Generally, grain sizes range from 0.1 mm to >2.0 mm. Angularity has been added to the grains with the development of minor quartz overgrowths.

Few sedimentary structures occur in the Cable Canyon Member. Although no cross-laminations were recognized in this study, cross-laminations at other localities were noted by other workers (Kelley and Silver, 1952; Pratt, 1967; Flower, 1961). Small vertical traces, which appear to be burrows filled with quartz grains, represent the only preserved biogenic structures within the Cable Canyon Member.

In the thicker beds of the unit, a noticeable increase in dolomitic matrix and decrease in grains are evident toward the top of the bed. However, the fabric remains grain-supported. The upper contact of the Cable Canyon is generally sharp and locally appears to have symmetrical wave ripples.

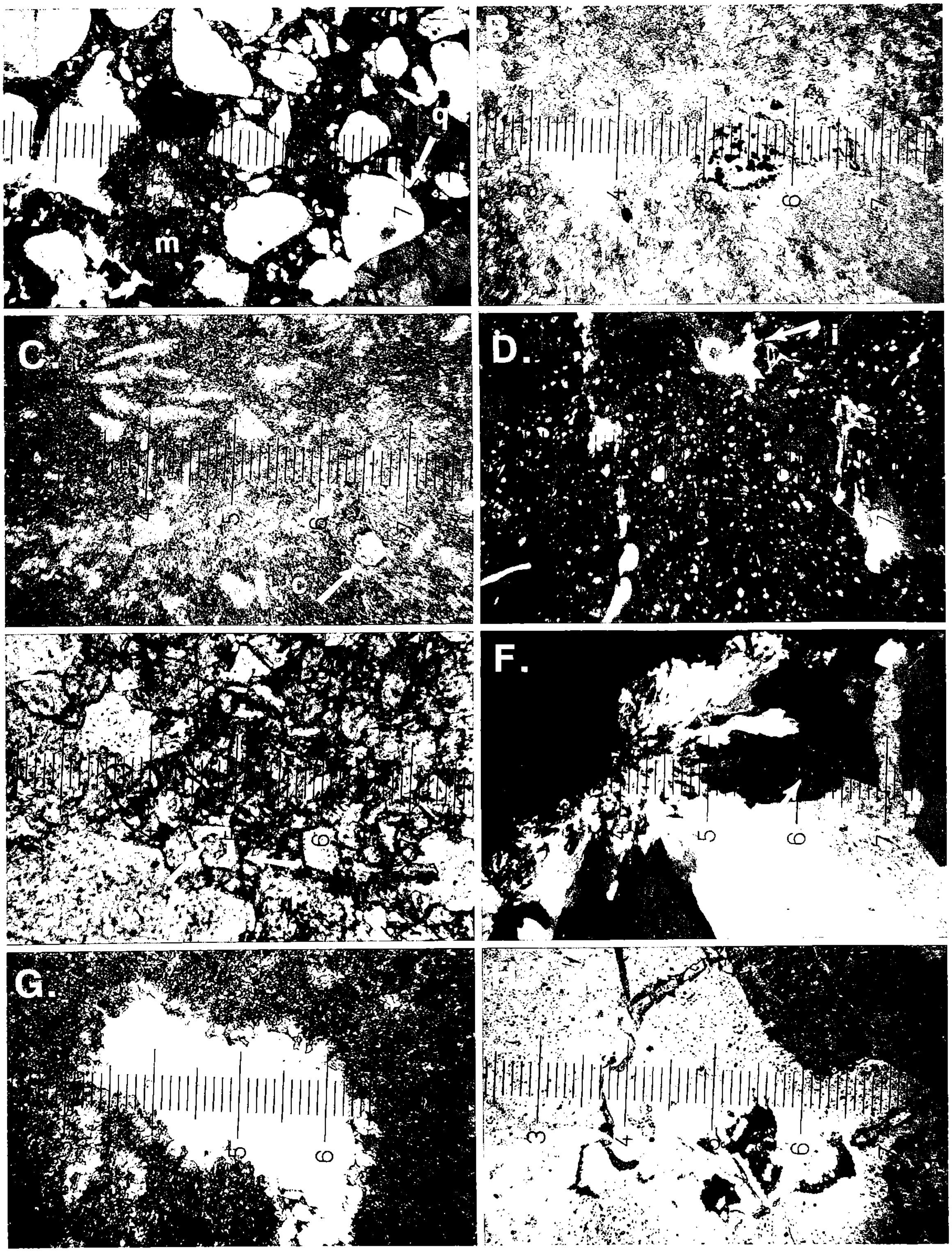
Upham Member.—The Upham Member averages 22 m in thickness throughout the study area (Fig. 3). The Upham forms a steep cliff at the base of the Montoya Group in the Sacramento Mountains. The dark-brown dolostone is massively to crudely bedded at the top. Quartz grains are absent above the basal 12 m of the Upham. The dolostones are bioclastic wackestones–packstones with some grainstones (Fig. 4B). Packstones increase toward the top of the member. Current laminations are recognized locally. A zone of intraclasts also occurs locally. Chert is not abundant in the Upham Member; however, nodules, spheroids, and disseminated masses replaced the carbonate in the upper few meters of all measured sections. The remainder of the silica in the Upham Member replaced shells (Fig. 4B). Dolomite also replaced many shells. Broken bioclastic grains (mostly crinoids and rugose corals) dominate in the Upham Member. The remaining biota consists of tabulate corals, stromatolites, brachiopods, gastropods, trilobites, nautiloids, and conodonts. Silica and dolomite replacement cause poor fossil preservation. Echinoderms are completely disarticulated; plates and ossicles constitute the greatest percentage of the fossil material in the Upham Member. The echinoderm grains and rugose corals exhibit evidence of transport. The solitary corals occur usually as individuals 3–4 cm in size lying on their sides, or as large fragments. Intervals of abundant rugose-coral debris alternate with noncoralline intervals within the Upham Member. The rugose corals belong to the genus *Paleophyllum,* a form characteristic of the Second Value (Flower, 1961). Rare chain corals, *Catenipora,* also occur within the lower 12 m of the Upham Member in the Sacramento Mountains. Where *Paleophyllum* and *Catenipora* are found together, they are in growth position.

A single stromatolitic horizon was identified within the Upham Member. Approximately 11 m above the base of the Second Value Formation, "stromatolitic" laminae have been replaced by chalcedony. The wavy morphology of these laminae is indicative of subtidal mats (G. Strathearn, personal communication, 1984). The laminated clasts probably became detached by wave action during a storm. At the Water Station section, small oncoids (1 cm in diameter) occur at this stratigraphic horizon instead of algal laminae. In Arcente Canyon, a *Catenipora* horizon occupies the same stratigraphic level as the "algal" horizon.

Brachiopods, gastropods, and sponge spicules gradually increase in abundance within the upper 5 m of the Upham Member. The brachiopods are small rhynchonellid and atrypacean forms which are preserved both articulated and disarticulated (Howe, 1965,a,b,1967). Moderately high-spired and loosely coiled gastropods and sponge spicules were retrieved from insoluble residues. The addition of these faunal components toward the top of the Second Value indicates that environmental conditions within the area favored faunal diversification, or may only reflect improved preservation. However, the faunal diversification continues upward into the Aleman.

The contact of the Upham Member and the Aleman Formation is drawn at the base of the bed containing the lowermost continuous band of chert nodules associated with a decrease in dolomite crystal size. The transition from the characteristically massive Second Value strata to medium-bedded, chert-rich Aleman strata takes place over several meters as noticeable bedding develops. Dark, coarse-crystalline (0.7–1.0 mm) and light, fine- to medium-crystalline (0.2–0.5 mm) dolostones are interbedded 20–30 m above the base of the Montoya Group. These are coarse-crystalline and medium-crystalline fabrics under the modified classification of Pray (1953). The faunal diversification continues to increase upward into the Aleman Formation.

Figure 4 (opposite page). Photomicrographs of Montoya Group depositional and diagenetic fabrics. Each small division on the scale is 10 μ. *A*—Cable Canyon Member fabric. Note sub-rounded texture, minor quartz overgrowths (q) and dolomitic matrix (m). *B*—Upham Member fabric. Intensely dolomitized fabric with recrystalized matrix and preserved crinoid fragments. *C*—Aleman Formation fabric. Recrystalized dolomite with dolomite-centered biomolds and one shell replaced by chert (c). *D*—Cutter Dolomite fabric. Note the dolomicrite matrix and selective recrystallization and dissolution of a remnant intraclast (i). The open porosity is moldic and fenestral voids. *E*—"Pillbox" structures (p) from the Second Value–Aleman succession. Note the euhedral and recrystalized dolomite in addition to the calcite cement (cc) in the center of the dolomite rhombs. *F*—Anhydrite nodule replaced by chalcedony and megaquartz from the Aleman. Tiny blebs (b) are remnants of andydrite. *G*—Moldic pore with scalenohedral cements from the upper Aleman Formation. The cements (c) were emplaced as calcite vadose cements and later dolomitized. *H*—Chalcedony replacement of shells and microquartz replacement of matrix from the Aleman Formation. Some shells (s) were stabilized to calcite and remained unreplaced.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Aleman Formation

The Aleman Formation is a fine- to medium-crystalline, medium-bedded dolostone that contains abundant, dark ribbon and nodular chert in bands oriented parallel to bedding. The Aleman forms a series of gentle slopes and short cliffs above the massive Upham cliff. Short cliffs in the Aleman exposure represent intervals of increased chert percentages. In the Sacramento Mountains the thickness of the Aleman ranges from 42 to 84 m (Fig. 3).

Chert masses associated with shell debris grade upward into chert ribbons at the top. The upper 5–10 m of the Aleman contains 30–50% replacement chert. In all the measured sections except the Arcente Canyon section, a 30- to 50-cm bed of white, black, and rust-colored chert marks the top of the Aleman Formation. At Arcente Canyon, the top chert bed is absent, but ribbon cherts remain abundant.

The light-gray Aleman Formation is a bioclastic wackestone (Fig. 4C). Packstone lenses containing small invertebrates are present. Silica commonly replaced the skeletal remains. Thin beds of fossiliferous carbonate silts, as well as wave-laminated beds, occur in the upper Aleman.

The macrofauna present in the Second Value Formation persists and diversifies through the transitional contact into the Aleman Formation. *Paleophyllum* disappears within the lower 10 m of the Aleman. Other corals are not present. Echinoderm remains occur through the lower 30 m of the Aleman. As the mud matrix becomes dominant, this fauna is replaced by a diverse brachiopod fauna.

Most of the brachiopods are well preserved by silica replacement (Fig. 4H). Brachiopods begin to dominate the fauna ~15 m above the base of the Aleman Formation. Rhynchonellid and atrypacean brachiopods carry over from the Second Value faunas. Dalmanellids, orthids, and strophomenids add to the diversity of the brachiopod fauna in the Sacramento Mountains. The size and abundance of individuals increase toward the top of the unit. Intervals rich in brachiopods alternate with brachiopod-poor intervals to within 5 m of the top of the Aleman. Variable preservation may have produced the alternating intervals. Brachiopods are absent from the cherty beds at the top of the Aleman. An intensely burrowed bed occurs in the Dog Canyon section at 55 m and in the Water Station section at 60 m above the base of the Montoya Group. The burrows resemble *Chondrites*, a below-wave-base trace fossil (Chamberlain, 1978). Other faunal constituents in the Aleman Formation include tiny gastropods, sponge spicules, nautiloids, and conodonts. Sponges probably increased in abundance as the quieter-water fauna developed in the lower and middle Aleman.

The entire Aleman fauna disappears near the top of the formation. Chert continues to be abundant, and truncated current laminations replace the diverse fauna within the upper beds. Intense silicification removed any evidence of fossils in the uppermost 3 m of the Aleman. The multicolored chert bed at the top of the Aleman contains no apparent fossils or sedimentary structures.

The contact between the Aleman Formation and the overlying Cutter Dolomite is sharp. Gray, cherty Aleman dolostones lie below the white, nonsiliceous dolostones of the very fine-crystalline to dolomicritic Cutter Dolomite.

Cutter Dolomite

Within the Sacramento Mountains, the Cutter ranges from 40 to 60 m thick (Fig. 3). The variable thickness reflects the effects of post-Montoya erosion. The light-gray to white, thin to medium dolostone beds form a continuous slope of step-like benches above the slope-and-cliff series of the Aleman and below the massive cliff of the overlying Fusselman Dolomite. The Cutter Dolomite generally weathers ruggedly. Abundant vertical fractures and a few large chert nodules are characteristic. The nodules are lenticular masses oriented parallel to bedding. Abundance of the large nodules increases toward the top of the Cutter, as do thin, continuous chert beds. The Cutter is generally a bioclastic, intraclastic wackestone with the matrix exclusively dolomicrite (Fig. 4D).

Approximately 14 m above the base of the Cutter, a 5- to 10-m-thick argillaceous dolomite and yellow, dolomitic shale is present. The clastic interval is areally continuous in the Sacramento Mountains. The shale is thinly laminated. Quartz silt occurs within the shale. Above the shale, argillaceous material decreases, and the dolostones are again dominant.

In the absence of a diverse fauna, abundant sedimentary structures are useful in the Cutter for environmental analysis. Cyclic bedding is apparent in the Cutter Dolomite. Cycles comprising a thick dolostone bed, laminated medium-bedded dolostone bed, and thin dolostone bed develop or partially develop over 3- to 6-m intervals. Wave laminations are present in some beds. Beds containing laminations in the upper 3 to 4 cm are common and represent thin stromatolite mats. Laminations with sediment-filled mud cracks were recognized in the Cutter. Other upper laminated surfaces contain abundant phosphatic debris which could represent brief hiatuses. Birdseye structures are found in the Cutter, but are not associated with the stromatolite mats, and possibly are a response to desiccation and shrinkage (Shinn, 1968). Beds of intraclasts occur above the areally continuous shale interval. The intraclasts indicate postdepositional erosion and transportation. Intraclast zones are laterally restricted within beds at the Water Station and Table Top sections.

The absence of allochems is the product of a sparse fauna in the Cutter. Bivalves, tabulate corals, conodonts, stromatolites, and a local burrowed horizon comprise the total evidence of the fauna and flora. Silicified bivalves, whole and broken valves, and bivalve molds are easily recognized in the Cutter section from both thin section and hand samples. Several beds above the shale interval contain tabulate corals. From one to four different coral beds occur in each section in the Sacramento Mountains. The tabulate corals (*Paleofavosites*) are preserved in life position as low domal masses, as large as 30 cm in diameter and 10 cm in height, resting in dolomicrite. Evidence suggests the possibility of extensive stromatolite growth in the Cutter. "Puckered stromatolite mats" have been dolomitized or replaced by chert. The morphology of these stromatolite mats indicates high intertidal–supratidal origin. A localized burrowed bed occurs near the top of the Table Top section. These *Chondrites* burrows are similar to those in the Aleman. *Chondrites* indicates shallow, quiet-water deposition near a tidal flat or on the open shelf (Chamberlain, 1978). Some small, unidentifiable burrows occur in the top 3 cm of upper Cutter beds at the Arrow Canyon section.

Lithologically, the upper 10 m of the Cutter Dolomite shows no significant change, although the macrofauna is virtually absent. Phosphatic lags, laminations, and chert are common in the uppermost beds of the Cutter. The white dolomicrite beds are in sharp contact with the overlying brownish-pink, coarsely crystalline beds of the Fusselman Dolomite (Fig. 3). An undulose erosional surface with relief of >2 m marks the disconformity at the top of the Cutter.

Fusselman Dolomite

The Fusselman in the Sacramento Mountains is >50 m thick in the south, but absent due to erosion in the north. Two units of the Fusselman crop out in the Sacramento Mountains; a coarsely crystalline, brownish-pink lower member and a darker, more finely crystalline and cherty upper member. The lower member rests on the Cutter in all the measured sections except in Dry Canyon, where the Fusselman is absent due to a post-Fusselman disconformity.

DIAGENETIC FABRICS OF THE MONTOYA GROUP

Diagenetic fabrics in the Montoya Group are a result of depositional fabrics of the original sediments and postdepositional processes. The diagenetic data complement information provided by the depositional fabric in analysis of the postdepositional history of the Montoya Group. These carbonates contain fabrics in which several diagenetic processes can be recognized. Of these processes, dolomitization and silicification are the most significant.

Fabrics of the Second Value–Aleman Succession

Diagenetic fabrics in the Second Value and Aleman Formations are similar, but depositional textural variations of the sediments contributed to the significantly different outcrop appearances. The sequence and factors that gave rise to the observed diagenetic fabrics which caused the variations of these fabrics are discussed below for the Second Value–Aleman succession.

First-Generation Dolomitization

Burial of the wackestones, packstones, grainstones, and quartz arenite followed deposition. During burial, the lime mud was the first to be dolomitized by magnesium-rich brines. The metastable aragonite was probably stabilized to micritic dolomite and euhedral dolomite rhombohedra (Fig. 4D). Stabilization of aragonite to dolomite or calcite is a rapid process in recent carbonates of the Trucial Coast and Barbados, respectively (Kinsman, 1969; Matthews, 1968). Moving from the base of the Second Value up through the Aleman, the abundance of paramorphically replaced shells decreases significantly. Cemented biomolds and silicified shells indicate that the dolomite did not completely replace the sediment, leaving shells susceptible to later dissolution and replacement (Fig. 4C).

First-Generation Sulfate Emplacement

The association of sulfate replacement and dolomitization, common in many dolostones, is evident in the Second Value–Aleman succession. Anhydrite replacement as nodules and porphyroblasts followed dolomitization. Many shells probably were replaced by anhydrite, but evidence of that has been removed by later diagenesis.

In the Second Value, the evidence for the presence of sulfates is concentrated in the Upham Member as "stair-step" anhydrite porphyroblast molds and "pillbox" structures. The original anhydrite spar which replaced the dolomite matrix was dissolved, leaving "stair-step" molds. "Pillboxes" (dolomite rhombohedra with intracrystalline pores) represent dissolution of anhydrite which had replaced the centers of first-generation dolomite crystals (Jacka, 1984; Fig. 4E). In addition to anhydrite porphyroblasts and "pillbox" structures described in the Upham, altered anhydrite nodules occur in the Aleman. The anhydrite nodules have been replaced by silica and are now cauliflower cherts with tiny blebs of anhydrite remaining in the replacement chalcedony and megaquartz (Fig. 4F) The anhydrite must have followed dolomitization, since it replaces the early dolomite fabric.

First-Generation Freshwater Influx

Evidence of a meteoric phreatic and overlying meteoric vadose zone exists in the carbonate rocks as an overprint on the dolomite/sulfate-influenced Second Value–Aleman succession. Meteoric water dissolved shells, anhydrite, and matrix, leaving biomolds, anhydrite molds, rare intracrystalline pores, and dissolution channels (Fig. 4C,F). Blocky and scalenohedral calcite cements were precipitated in pores according to the diagenetic environment (Fig. 4G). As sea level rose and the water table moved updip, pores filled with cement.

In addition to the calcite cementation, the meteoric water probably gathered silica from the unstable sponge spicules, which subsequently nucleated in the fossils as shell replacement by chalcedony (Fig. 4C). Once shells were replaced, microquartz replaced the matrix to form the reddish-orange to black nodules, ribbons, and spheroids, which are found primarily in the Aleman. Silica replacement in the uppermost Aleman obliterated depositional fabric, and the top Aleman bed was completely silicified. Chalcedony and megaquartz replaced sulfates (Fig. 4F) and cemented moldic and intrabiotic pores. Also, poorly developed quartz overgrowths formed on the quartz grains of the Cable Canyon Member as the silica-rich fluids passed through the deposit (Fig. 4A). Unreplaced dolomite in the silica and silica replacement of the sulfates indicates that the freshwater influx was the third diagenetic event. Dissolution, calcite precipitation, and silicification are dependent on the rate of fluid movement (Matthews, 1968) and silica saturation of the fluid. Silicification probably preceded dissolution and calcite precipitation, because many of the shells were partially silicified and their outer edges were dissolved. The bedding-parallel orientation of the chert ribbons and nodules suggests that the meteoric water probably moved laterally, downdip.

Second-Generation Dolomitization

A second dolomitization event is evident in the Second Value and Aleman carbonates. The dolomite matrix was recrystallized, dolomite replaced calcite cements paramorphically, dolomite replaced feldspar grains, and dolomite spar-cemented open pores. The dolomitization continued long enough for dolomite to replace calcite cements (Fig. 4G) and to begin to replace silica. Recrystallization, which obliterated most of the preserved depositional and diagenetic fabrics, was a rapid process, because nucleation sites were present (similar to recrystallization in calcite; Matthews, 1968). The mosaic texture of the dolostones in the Second Value–Aleman succession was created during recrystallization (Fig. 4B,C). The differing crystal sizes of the Second Value and Aleman dolomite matrices probably reflect differences in the original sedimentary fabrics. Medium- to coarse-crystalline dolostones formed where grains were abundant. Where the fabric was mud-supported in the Aleman, finely crystalline dolostones formed. The second dolomitization also produced pore-filling cements.

Second-Generation Sulfate Emplacement

Minor evidence of second-generation sulfates occurs as fracture- and pore-widening molds and as matrix replacement. Second-generation sulfates are probably related to the second-generation dolomitization during burial. Early diagenesis ended with the emplacement of these sulfates in the Second Value and Aleman.

Dolomitization

The Second Value and Aleman were most likely not dolomitized prior to shallow burial. For several geologic reasons, most models which describe burial dolomitization (Adams and Rhodes, 1960; Deffeyes and others, 1965; Badiozamanie, 1973; and others) cannot be applied to the Montoya. One model which may have implications for the Montoya was postulated by Jacka and Franco (1974). They proposed that fluids rich in magnesium, chloride, and sulfate ions, originating beneath the coastal plain, moved downward and laterally through subtidal sediments to dolomitize them. The water, possibly fresh water, initially would gather ions as it dissolved updip sulfates during downdip migration. This model applies to the Montoya, specifically the Second Value–Aleman succession, in that dolomitization would have begun soon after its deposition, during early burial. The dolomitizing fluid must have originated updip from the subtidal deposits in the study area. Supratidal, sulfate-rich deposits were described by Geeslin and Chafetz (1982) in the Aleman; therefore, a source for dissolved magnesium and sulfate ions was available. The final question arises: Could the dolomitizing brines migrate 320 km downdip without appreciable change to dolomitize the Second Value and Aleman? The dolomitization of shelf-edge Guadalupian carbonates in the Permian basin, as described by Jacka and Franco (1974), required the fluids to move at least tens of kilometers. Guadalupian dolostones of the Permian basin grade into deeper-water limestones downdip. The same diagenetic transition occurs in the deeper-water facies of the Second Value and Aleman southwest of the Sacramento Mountains. This point favors long-distance lateral migration of dolomitizing fluids through the Montoya.

Fabrics of the Cutter Dolomite

Much of the original depositional fabric of the Cutter Dolomite is preserved, although the rock has been extensively dolomitized. The preservation of the depositional fabric may represent early, rapid dolomitization of the mud-rich sediments. The same sequence of diagenetic events which altered the Second Value–Aleman section probably affected the Cutter, but these events appear to have started before burial, or penecontemporaneously with deposition. The diagenetic sequence of the Cutter and an explanation of its expression is described as follows.

First-Generation Dolomitization

Dolomitization was the first diagenetic process to affect the Cutter sediments. Dolomite probably replaced aragonite mud in the sabkha and peritidal carbonates. The generally micritic fabric of the Cutter has some very fine-crystalline mosaic textures and rare, coarse-crystalline idiotopic textures. Also, first-generation dolomite cements line or fill birdseye vugs and mud cracks created by near-surface and surface desiccation.

First-Generation Sulfate Emplacement

Sulfate emplacement followed and may have accompanied dolomitization of the muddy sediments, although evidence for abundant sulfates is not present in the Cutter. Some anhydrite nodules (now cauliflower cherts; Chowns and Elkin, 1974) and porphyroblasts replaced the carbonates. Anhydrite may have also replaced the shells. Collapse breccias near the top of the Water Station and Table Top sections suggest more-extensive replacement by sulfates and subsequent dissolution of sulfates.

First-Generation Freshwater Influx

As in the Second Value–Aleman succession, a meteoric-water overprint on early dolomitization and sulfate-emplacement features is present in the Cutter Dolomite. However, chert and calcite cement is not as abundant in the Cutter as in the Second Value and Aleman. The decrease in chert probably indicates the lack of an internal silica source (sponge spicules). Chalcedony replaced some bivalves. Minor amounts of dolomicrite were replaced by microquartz and chalcedony. Microquartz replacement of the possible "algal" mats is represented by thin (10 cm), continuous beds at the top of Cutter. Large concretions in the Cutter represent replacement of *Paleofavosites* by chalcedony, and continued replacement of the surrounding mud matrix by microquartz. Black to rusty-brown, lenticular nodules occur in the upper Cutter Dolomite where the tabulate corals are prominent, but not all of the chert masses contain a coral nucleus. Orientation of the lenticular chert nodules parallel to bedding probably indicates that meteoric water migrated parallel to bedding. Chalcedony also replaced a few of the anhydrite nodules. Both silica forms, chalcedony and megaquartz, occur in minor amounts as cement in moldic, intercrystalline, and fracture porosity. The most-extensive silicification occurred in the upper Cutter. Only minor silica replacement is evident in the lower Cutter, but it does extend to near its base. Evidence of other meteoric-water processes, such as dissolution and calcite cementation is not abundant.

Second-Generation Dolomitization

A second phase of dolomitization, also seen in the Second Value–Aleman succession, is represented by recrystallization of a few of the Cutter dolostone beds and dolomite cementation in open pores (Fig. 4D). Where recrystallization did occur, very fine-crystalline mosaic textures formed. Burrows in bioturbated horizons have been selectively recrystallized. Second-generation dolomite cements filled open birdseye vugs, biomolds, anhydrite molds, and fractures. Compromise boundaries are common in voids filled with ferroan and nonferroan dolomite spar. Dolomite cement in fractures through chert also began to replace the chert. Dolomite spar is present in solution channels, nontectonic fractures, and intergranular pores of the collapse breccias. Sulfate cements probably precipitated in the few unfilled pores during and following dolomitization, but evidence for such an event is minimal. The Cutter Dolomite contains no evidence of additional diagenesis.

The same cyclic diagenetic sequence which affected the Second Value–Aleman section altered the Cutter Dolomite, but the events probably took place later, had shorter duration, and began before burial of the sediment. In both cases, the rocks were left nonporous and impermeable as postdepositional diagenesis ended with pore-occluding cementation and recrystallization.

Cenozoic Diagenesis

The effects of Basin and Range deformation (Cenozoic) and post-structural diagenesis are similar throughout the entire lower Paleozoic section in the Sacramento Mountains. Cenozoic faulting raised the Montoya Group from the subsurface—where significant diagenetic changes probably had not occurred since the Devonian—to the surface. Brittle carbonates fractured during the uplift. The greatest amount of fracturing occurred near the fault plane on the western escarpment of the Sacramento Mountains. The thin-bedded Cutter fractured most intensely. The mechanically produced porosity in all three Montoya formations decreases rapidly eastward away from the escarpment. Blocky, equant calcite cement is common in the tectonically produced fracture porosity.

APPLICATION OF CONODONT DATA

Relationship of Conodont Faunas to Water Depth

The distribution of condont genera found in the Montoya Group appears to be related to depth, as is the case for Upper Ordovician carbonates of the North American craton. The genera and their generalized depth relationships are listed in Table 1; these paleoecological interpretations are based on observations of faunal distributions at many North American Midcontinent Province sites (Sweet, 1979a,b).

The relationship between these genera and lithofacies aided in interpretation of depositional environments of the Montoya Group of the Sacramento Mountains. A comparison of overall faunal depth relationships of the Montoya conodont faunas in the Sacramento Mountains to those of the Franklin Mountains and Florida Mountains (Sweet, 1979b) also helped in the reconstruction of the paleogeographic position of the study area during the Late Ordovician.

Elements of *Panderodus* dominate the fauna of the Second Value Formation in the Sacramento Mountains (Fig. 5). The abundant *Panderodus* is associated with shallow-water stromatolite mats, bioclastic wackestones, and bioclastic packstones. Elements of *Plectodina* appear in the middle and at the top of the formation, associated with packstone fabrics. A few *Phragmodus* elements also occur within these two intervals. A low-percentage belodinid peak separates the two trends. *Oulodus* elements also occur in the break. The overall trend is toward deepening water through the Second Value, as shown by the occurrence of deeper-water conodont elements and deeper-water lithofacies at the top of the Second Value.

Shelf faunas are more abundant in the mud-rich, quiet-water Aleman deposits. Elements of *Panderodus* remain abundant, but elements of *Plectodina* and *Phragmodus* dominate periodically in the lower Aleman. Belodinid and *Aphelognathus* + *Oulodus* elements increase in abundance, then disappear or occur only in low percentages between the peaks of deeper-water fauna. The high percentage of *Plectodina* and *Phragmodus* species indicates deeper-water conditions for the Aleman. The macrofauna diversifies concurrently with the increase in abundance of *Plectodina* and *Phragmodus*.

TABLE 1. — CONODONT GENERA AND ENVIRONMENTS

Genus	Environment	
Rhipidognathus	Intertidal	Shallowest
Aphelognathus + *Oulodus*	Shoal water	
Belodinids	Normal marine—Shallow water	
Panderodus	Normal Marine—Shallow Water	
Plectodina	Subtidal—Shallow water	
	Intermediate platformal	
Phragmodus	Shelf platformal	Deepest

In the upper 20 m of the Aleman, the fauna changes abruptly to a fauna rich in *Aphelognathus* + *Oulodus*. At the same time, *Phragmodus* elements disappear and *Plectodina* elements grade out. One last sharp *Panderodus* surge occurs near the top of the Aleman. The shallower-water fauna is in current-influenced beds.

The abundances of conodont genera in the Franklin Mountain and Florida Mountain sections as recorded by Sweet (1979b) are important to discern the paleogeographic position of the Sacramento Mountain Montoya sediments. The greater abundance of *Plectodina* elements in the Second Value of the Franklin–Florida section indicates that deeper-water conditions developed south of the study area. The lower Aleman faunas of Sweet (1979b) contain a persistent *Plectodina–Phragmodus* peak. *Plectodina* elements remain abundant to the top of the Aleman but *Aphelognathus* + *Oulodus* elements are the most abundant. The water depths shallowed as the upper Aleman was deposited in the Franklin and Florida Mountain areas. No change in faunal composition occurs at

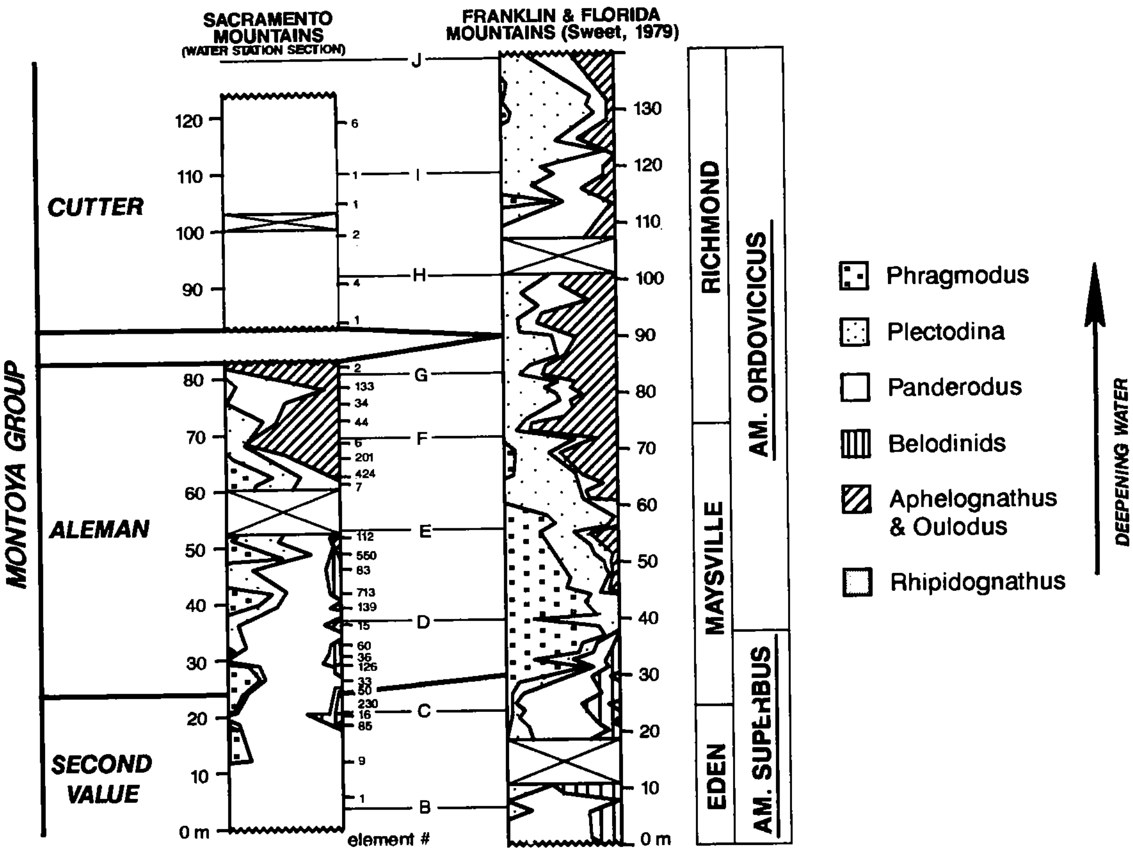

Figure 5. Relative-abundance logs from the Sacramento Mountains and a composite from the Franklin and Florida Mountains, Texas and New Mexico. Note the lower abundance of deeper-water faunas in the Sacramento Mountains section as compared to the Franklin and Florida section, 130 km to the south.

the Aleman/Cutter boundary in the Franklin–Florida section. However, a brief but sharp increase of shallow-water genera is evident. The lower Cutter depth-related generic abundances are similar to those of the upper Aleman, but in the upper Cutter, abundant deeper-water *Plectodina* elements with short-lived incursions of *Phragmodus* elements occur in the Franklin–Florida section. Although the underlying parts of the curve indicate fluctuation trends similar to those of the Sacramento section, the shift toward deeper-water faunas in the upper Cutter of the Franklin–Florida section is in contrast to the shallowing faunas in the Sacramento section.

The sharp change in lithology from the Aleman to the Cutter is accompanied by a virtual loss of conodont elements. Only 15 elements were recovered from the residues of the Cutter Dolomite, including *Rhipidognathus, Panderodus, Aphelognathus*, and *Oulodus*—all shallow water elements. Given the low abundance of elements, no other useful data are recognizable from the Cutter Dolomite conodont faunas.

Relative-Abundance Log and Graphic Correlation

The conodont fauna recovered from the Montoya in the Sacramento Mountains does not provide a direct dating of depositional events for the Late Ordovician in the study area; however, some limited correlations of the Sacramento Second Value and Aleman with the Montoya section in the Franklin–Florida section (Sweet, 1979b) and the Midcontinent Standard Reference Section (MSRS) are possible, using relative-abundance logs (Fig. 5) and graphic correlation (Fig. 6). Correlation of the Cutter Dolomite with the Franklin–Florida section and the MSRS uses only the depositional and diagenetic information gathered from the field.

Relative-abundance logs from the Sacramento Mountains Montoya and the Franklin–Florida Montoya are correlated on the basis of "inflection points" which indicate "shoaling epochs or deep-water maxima of chronostratigraphic significance" (Sweet, 1979b). The "inflection points" are

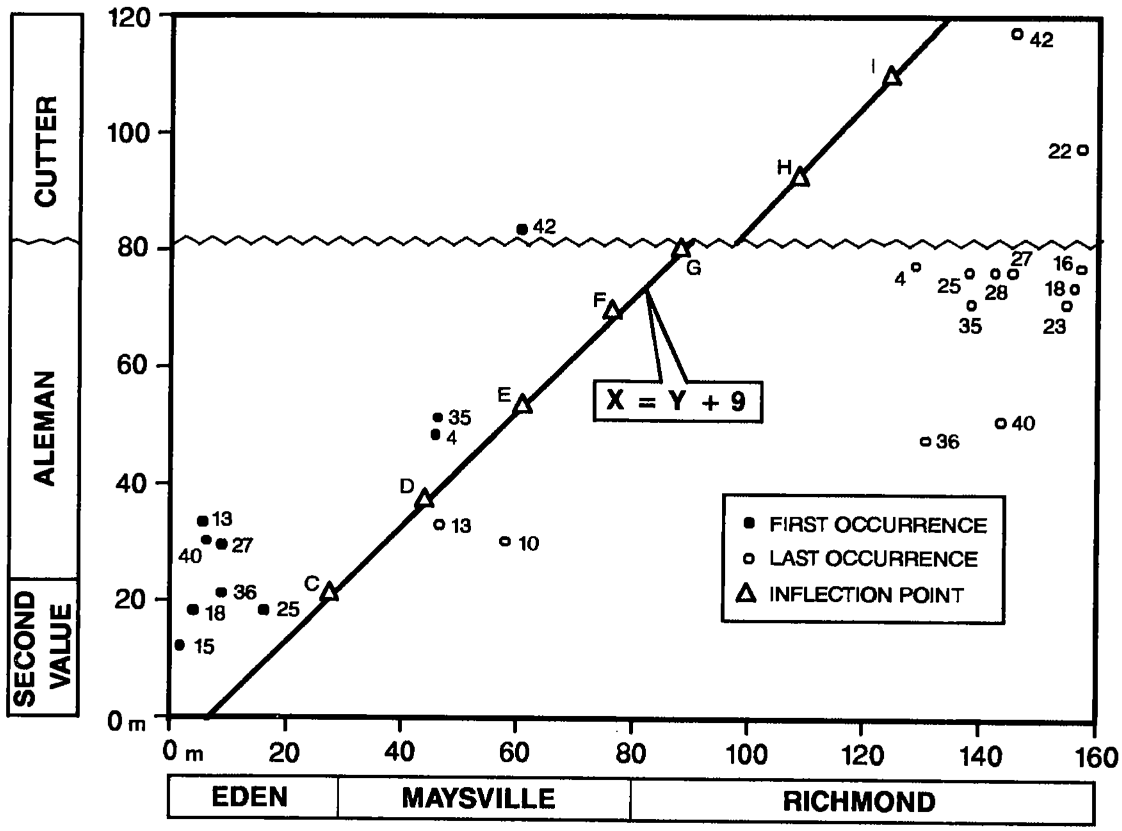

Figure 6. Conodont graphic correlation of the Midcontinent Standard Reference Section (MSRS; X: Sweet, 1979b) versus the Sacramento Mountain section (Y). Species identification can be found in Brimberry (1984) or Sweet (1979b). The line of correlation is defined by first and last occurrence points in the Second Value–Aleman succession, but the break (82 m) and line of correlation in the Cutter is implied by depositional and diagenetic information.

interpreted in the Second Value and Aleman Formations of the Sacramento Mountains and are correlated with the documented "inflection points" of the Franklin–Florida section, indicating the dating/timing of Montoya depositional events in the Sacramento Mountains.

The relative-abundance logs indicate that Late Ordovician deposition of the Second Value and Aleman Formations began later and ended earlier in the Sacramento Mountains than in the Franklin and Florida Mountains. Deposition of the Second Value began in the Edenian (Fig. 5). Carbonates appear to have been deposited continuously through the Maysvillian and into the early Richmondian where the top of the Aleman is correlated. The Cutter Dolomite probably was deposited within the Richmondian based on lithostratigraphic correlation with the Cutter in the Franklin–Florida section, which was deposited entirely within the Richmondian (Sweet, 1979b).

Graphic correlation of the Sacramento Mountains section with the MSRS also helps to date Montoya depositional events (Fig. 6). Again, data sufficient to provide meaningful correlations exist only in the Second Value and Aleman Formations. Detailed listings of the conodont data used for this graphic correlation can be found in Brimberry (1984) and Sweet (1979b).

Graphic correlation of the Sacramento Mountains section with the MSRS, with Late Ordovician conodont species first and last occurrences and "inflection points," indicates the same dating of the Second Value and Aleman depositional events as the relative-abundance logs. In addition to the dating of events, the graphic correlation also indicates that the "rate of rock accumulation" is the same for the Sacramento Mountains and the MSRS. The break (unconformity) separating the Aleman and the Cutter Dolomite is inferred from the depositional and diagenetic fabrics discussed earlier. No first or last occurrences of conodont species are meaningful in the Cutter Dolomite. The slope of the line of correlation is extended above the unconformity, as implied by lithostratigraphic correlations of the Sacramento Mountains section to the Franklin–Florida section discussed previously.

MONTOYA GROUP HISTORY

The three formations of the Montoya Group in the Sacramento Mountains were deposited in two distinct depositional systems. Although diagenesis has obliterated most of the primary sedimentary fabrics, the remaining lithologic and paleontologic data allow interpretation of the history of the Second Value and Aleman Formations and Cutter Dolomite. Not only are the depositional environments of the two systems important, their paleogeography and stratigraphic durations are significant to previous interpretations of the Montoya Group.

Second Value–Aleman Deposition

The erosional surface at the top of the El Paso Group apparently dipped to the south. Low relief existed on the erosional surface. The basal sandstone unit of the Montoya Group, the Cable Canyon Member of the Second Value, reflects the remnants of the topography at the beginning of the Late Ordovician (Edenian). The coarse sandstone thins over paleohighs and thickens in paleolows (Fig. 7A).

The Cable Canyon Member in the Sacramento Mountains was deposited as a sheet sand in shallow subtidal waters during a rapid transgression. Much of the quartz was probably derived from broad highlands to the northwest and north (Hayes, 1975) where the Bliss Sandstone (Cambrian) was exposed. The major clastic influx must have ended, or the highlands were submerged abruptly, as indicated by the abrupt transition from sandstone to clastic-free carbonates of the Upham Member of the Second Value Formation.

The transgression continued as the early Edenian seas washed and abraded shell fragments in the shallow subtidal waters characteristic of the Upham Member depositional environment (Fig. 7B). Intermediate- to high-energy water currents above storm wave base characterized Upham depositional processes. Grainstones or packstones grade landward and seaward to mud-rich deposits (wackestones). The lower Upham Member is mud-dominated and contains tabulate and rugose corals, echinoderm fragments, and stromatolites which probably lived in subtidal water above storm wave base. Some shoaling must have occurred as indicated by the higher-energy fabrics of grainstones and cross-laminations near the top of the Upham.

Below the contact between the Second Value and the Aleman, a facies change began as the transgression continued into Maysvillian time. The transitional interval around the Second Value/Aleman contact is marked by short pulses of the transgression, followed by short shoaling periods. Mud-dominated beds are interbedded with grain-supported beds across the contact. Through the transitional interval, coral and echinoderm fragments decrease in abundance as brachiopods increase.

Above the top of the transitional interval between the Second Value and the Aleman, sediments of the Aleman indicate deposition in subtidal water below storm wave base (Fig. 7C). Carbonate mud accumulated and quiet-water faunas diversified in water depths probably >10 m. The quiet, open marine water environments of the Aleman permitted faunal diversification of brachiopods, with the addition of bryozoans, nautiloids, sponges, and burrowing animals.

The transgressive trend ended in the upper Aleman at the beginning of the Richmondian. Deeper-water *Plectodina* and *Phragmodus* species are replaced by "shoal-water" species of *Aphelognathus* + *Oulodus* at the top of the Aleman. Cross-laminated carbonate sediments were deposited. The abundant brachiopod fauna diminished. The only survivors in the macrofauna at the top of the Aleman were a few large, thick-shelled brachiopods. It is evident that progradation or shoaling during the upper-Aleman regression moved deposition through and above storm wave base.

Post-Aleman Exposure

All of the evidence centering on the contact between the Aleman Formation and the Cutter Dolomite indicates that a previously undescribed unconformity is present at that level in the Sacramento Mountains (Fig. 7D). Continuous deposition from the Aleman into the Cutter has been described by Pray (1953), Hayes (1975), and Sweet (1979b). This appears to be the case in the Franklin and Florida Mountains sections; however, a regional break described by Flower (1958,1959,1961, 1965,1969) occurs at this contact in the Sacramento Mountains.

The abrupt lithic break from cherty to noncherty beds is only part of the evidence for the discontinuity. Drastic sedimentologic changes and faunal breaks support the interpretation of subaerial exposure between Aleman and Cutter deposition. The noncherty, nonfossiliferous beds above the contact were deposited in peritidal environments, as opposed to entirely subtidal environments below.

Meteoric calcite cement morphologies are present in the Second Value and the Aleman. Although dolomite paramorphically replaced calcite during burial, the blocky meteoric phreatic cement and scalenohedral meteoric vadose morphologies are stratified in the Second Value and

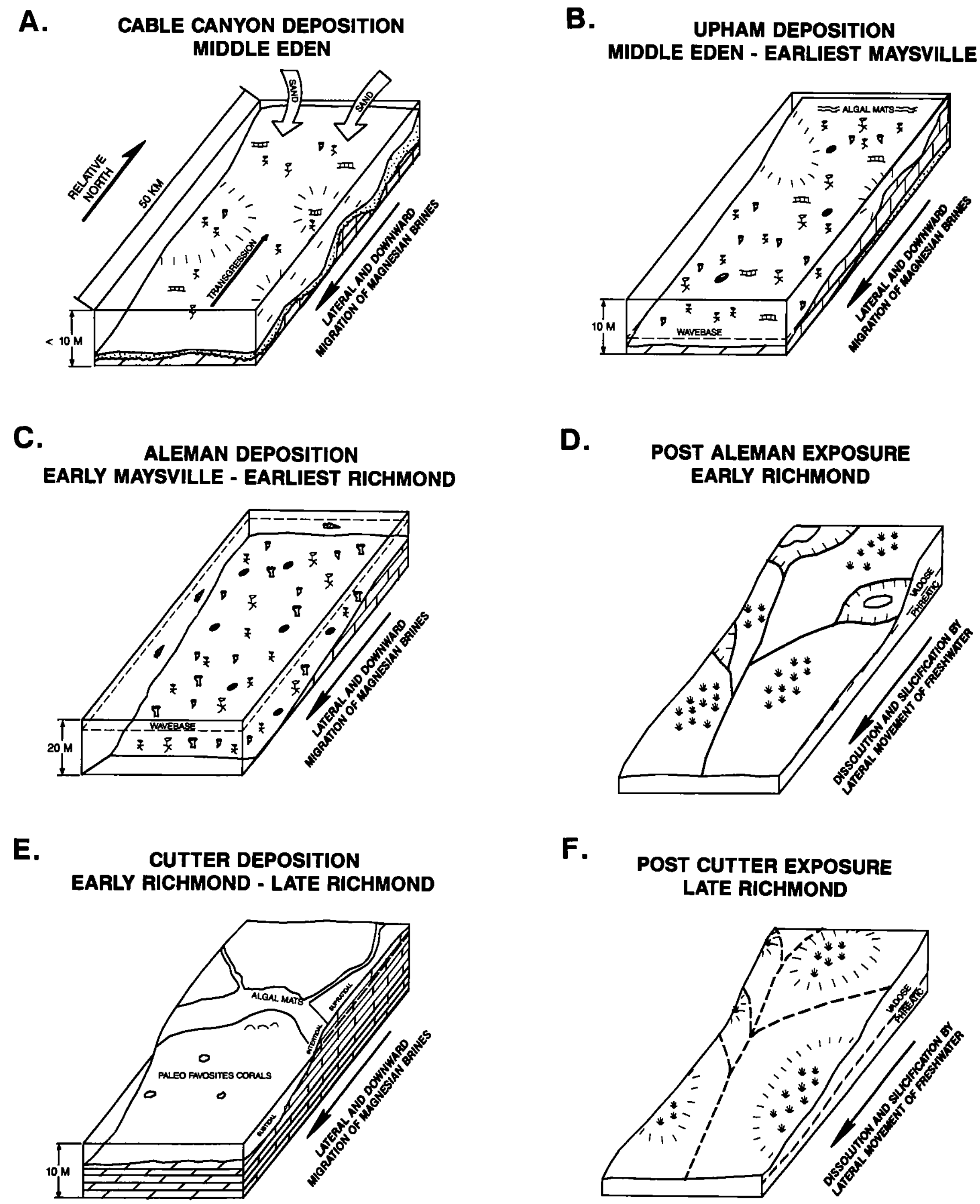

Figure 7. Block diagrams showing Montoya Group depositional history by event and timing, oldest (A) to youngest (F), during the Late Ordovician.

the Aleman. These cements are in moldic voids created by freshwater dissolution during subaerial exposure. The meteoric cements in the upper Aleman are not present directly above the Aleman/Cutter contact. Other evidence for subaerial exposure in the early Richmondian includes the variable thickness of the Aleman in the Sacramento Mountains.

Cutter Deposition

Following the exposure and erosion at the top of the Aleman in the early Richmondian, epeiric seawaters returned to the study area (Fig. 7E). Peritidal facies (Cutter) deposition began and persisted during the Richmondian. Wave and tidal currents were probably weak, as evidenced by the abundant lime-mud deposits.

Each phase of the peritidal facies tract is represented in the Cutter Dolomite depositional cycles. Supratidal evidence in the Cutter includes mud cracks, thin algal laminations, birdseye vugs, and intraclasts within thin beds. Tidal channels, puckered stromatolites, truncated ripples, and laminated beds are indicative of intertidal sedimentation. Thicker beds containing in situ *Paleofavosites* and *Chondrites* trace fossils represent some of the evidence for subtidal Cutter deposition. The shale interval is thought to have been deposited in the subtidal realm of the pertidal tract during a time when clastic input drowned out carbonate production.

Deposition of the upper Cutter Dolomite was probably interrupted by short periods of exposure which included erosion and diagenetic changes. Phosphatic lags present in the upper Cutter, indicate short episodes of erosion or nondeposition. Dissolution breccias and other freshwater phenomena probably identify periods of exposure and subaerial diagenesis.

Post-Montoya Erosion and Fusselman Deposition

The contact of the Cutter Dolomite and the Fusselman Dolomite is unquestionably unconformable (Fig. 7F). Contrast between the light-colored Cutter Dolomite and the brown Fusselman accentuates the contact. The Cutter Dolomite varies in thickness due to relief on the underlying erosional surface. Erosion at the end of Montoya deposition was probably caused by epeirogenic crustal shifting which exposed Cutter deposits to erosion and freshwater diagenesis before the end of the Ordovician.

Fusselman deposition followed erosion of the Montoya Group. The gentle slope of the epeiric sea floor had been modified by the post-Cutter cratonic tilting and erosion. The depositional surface was more undulose than any Montoya surface, as an epeiric sea once again transgressed over the area and Silurian deposition occurred.

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THE ARBUCKLE/OUACHITA FACIES BOUNDARY IN OKLAHOMA

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ABSTRACT.—The shelf rocks of the Arbuckle and Timbered Hills Groups of the southern Arkoma basin are mainly dolomites and sandstones, about 2,200–4,500 ft thick, dipping basinward to the south or southeast. The equivalent black shales of the Mazarn and Collier Formations have not been found in four marginal shelf wells.

The structural edge of the shelf is buried about 2–3 mi deep, south or southeast of the surface trace of the Ti Valley fault, with a fault displacement of 3,000 ft or more. This submarine block has been named Bengalia, or Ancestral Ouachita Mountains, and was probably active since Precambrian time; the fault face of the block is named the Bengalia fault. This is the probable source for Precambrian granite boulders in the Collier Limestone and Blakely Sandstone. The Johns Valley boulders were probably also derived from Bengalia, when slight uplift occurred beneath Bengalia. The large landslide masses within the Atoka Formation also were probably derived from Bengalia.

The Precambrian basement rocks below Bengalia were probably uplifted by pressure from below and by lateral pressure from the southeast, especially during Morrowan time, but beginning in Mississippian time with growth faulting along the Bengalia fault. Atoka growth faulting tilted the beds slightly southeastward or southward. With major post-Atoka compression, the Ouachita basin facies was thrust over the shelf facies along the Ti Valley fault, amounting to 10–15 mi of movement. The Ti Valley detachment appears to be 5 mi or more south or southeast of Bengalia, ~5 mi deep. Anticlines and horst blocks occur north or northwest of the detachment, below the Ti Valley fault, with several thousand feet of closure or displacement, involving basement-attached rocks.

Bedding-plane faulting has occurred in the Womble, Stanley, and other shales above the Ti Valley fault, causing thin-skin thrusting to occur simultaneously at many levels, concealing the deeper high-angle faults. Bengalia and the eastern Arbuckle Mountains were in the way of the Ti Valley fault, causing the bedding planes of the shelf facies to be displaced ahead of the Ti Valley fault, creating the Pine Mountain, Katy Club, and Choctaw faults. The pressure against Bengalia possibly caused some uplift and cracking of Bengalia, squeezing Atoka and older rocks against the Bengalia face, and possibly sourcing Bengalia with hydrocarbons.

The Ti Valley detachment must be close to the interface boundary between the shelf and basin facies, judging from gravity and magnetic interpretations, and changes in the Bigfork Chert. The Bigfork Chert in Black Knob Ridge contains much gray limestone, and was first described as the Viola Limestone, being a hybrid Viola facies. At the detachment interface, the Bigfork probably gradually contains more black shale and chert eastward 4–5 mi, judging from a well nearby. The underlying Simpson–Arbuckle–Timbered Hills rocks at the detachment probably have interbedded black shales, but this area has not been drilled. Only two wells have been drilled into Bengalia north of the Arbuckle Mountains; these were not to basement, and the Arbuckle Group was not tested, although much gas occurred in 800 ft of section of the Arbuckle in one well. Downside Bengalia has not been drilled.

Hydrocarbons could have moved into these rocks at many times from Cambrian to post-Atoka time, with excellent proven source beds in the Ouachita basin facies, and sandstones and fractured dolomites in the shelf facies. The basin was a live basin because of illite-bearing shales in the sequence, allowing fluids and gas to move with dilation. Tectonic dolomite could be formed at depth with dilation. Theoretical reserves of the Bengalia–Ti Valley detachment area should be ~2 quadrillion ft^3 of gas. Shallow zones <15,000 ft deep could have oil.

Ever since the 1890s, geologists recognized that somewhere beneath the Ouachita Mountains of Arkansas and Oklahoma there must be a facies change from a northern shelf of shelly carbonates to a southern basin of black, graptolitic shales (Fig. 1). Various facies diagrams have been proposed, the facies boundary being placed anywhere from 20 to 150 mi south of the Ouachita frontal fault.

The Ouachita Mountains have been studied for 130 years or more, and several thousand articles have been written about them. Surface mapping began in the late 1800s, and oil and gas tests began about 1907. Geophysical studies began about 40 years ago.

In the early 1950s, oil companies began shooting seismic lines in the mountains, but no deep wells were drilled. Shallow coring was done around the eastern Arbuckle Mountains to find the position of the Ouachita frontal fault. In 1958 the Max Pray-Wyrick well was drilled to 12,088 ft in sec. 26, T. 1 N., R. 14 E., in Atoka County, in search of the shelf facies beneath the basin facies; they drilled through Ouachita rocks mostly, but probably entered the Atoka Formation at ~10,000 ft.

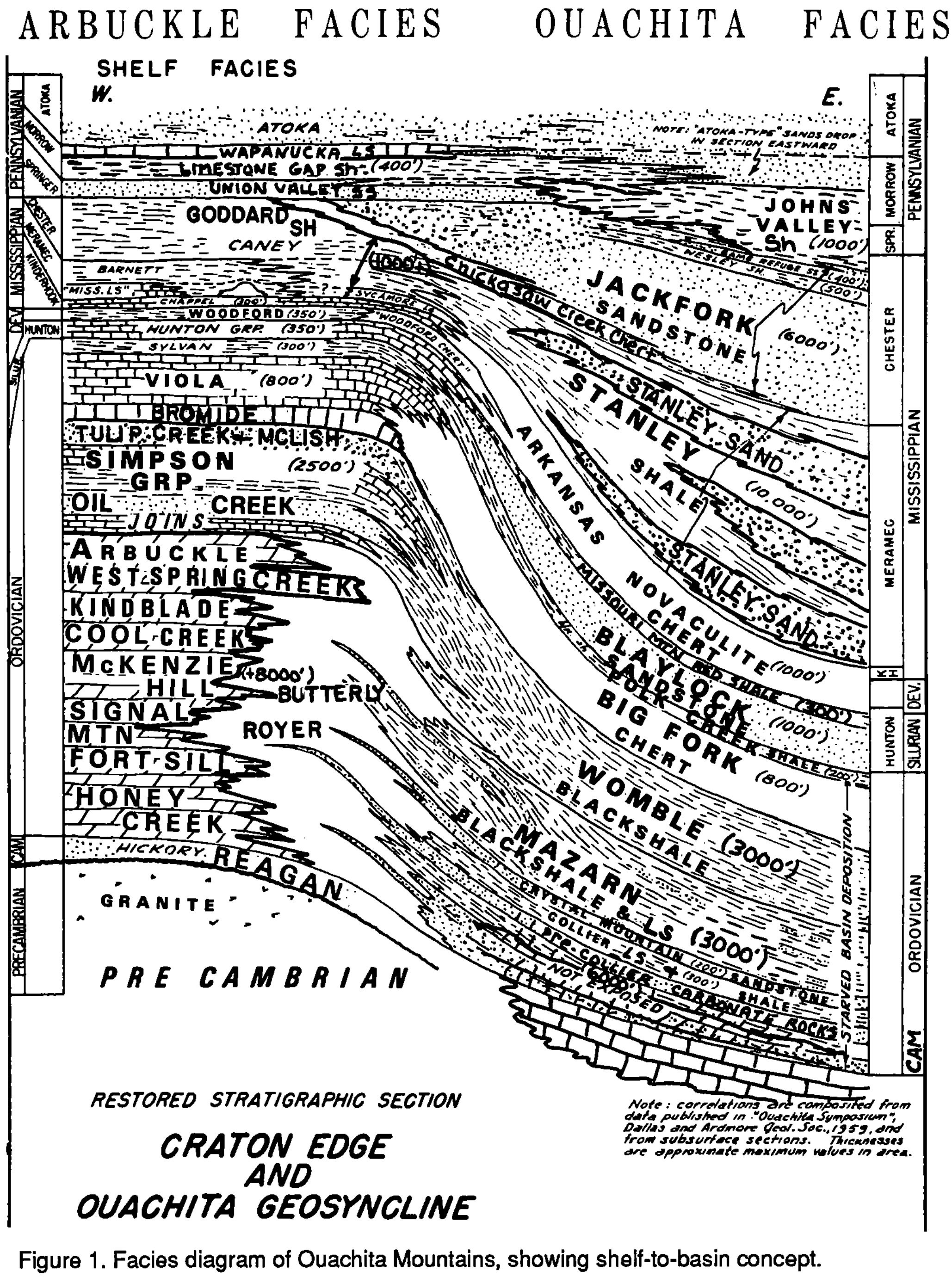

Figure 1. Facies diagram of Ouachita Mountains, showing shelf-to-basin concept.

In the past 25 years, about 10 wells have been drilled to basement around the edge of the frontal Ouachita fault of the Arkoma basin (Figs. 2,3). The Cambrian–Ordovician Arbuckle–Timbered Hills Groups are about 2,200–4,500 ft thick in the basement tests, the rocks being mostly dolomites and sandstones (Fig. 4). About eight wells have been drilled to the Arbuckle Group in the frontal Ouachita belt, but none have been drilled to basement. Only four wells were drilled through the Womble Shale, penetrating the Ti Valley fault from 2,800 to 12,000 ft. The prominent sandstones on the shelf are Reagan, Gunter (Butterly), and Roubidoux (= lower Cool Creek or Thatcher Creek Sandstone = Crystal Mountain Sandstone), and can be correlated over wide areas. Generally the rocks are tight, except where fractured, and where tectonic dolomite is present. Where black shales are present next to the fractured zones, large reserves of hydrocarbons may be present, such as those in the Wilburton field in T. 5 N., R. 18 E., north of the Choctaw fault.

The structure of the shelf is that of normal faults dipping basinward (Fig. 5). One of the last faults basinward from the surface trace of the Ti Valley fault is the Bengalia fault, with 3,000 ft or more of displacement, in places being part of a horst-block system termed Bengalia by Kramer (1933). Bengalia was named after the town of Bengal, Latimer County, near which an exotic block of Woodford–"Caney" rocks occurs in the Atoka Formation. Bengalia is the probable source of the Precambrian granite boulders in the Collier Limestone and Blakely Sandstone in Arkansas, the Arbuckle to "Caney" boulders in the Johns Valley Shale, and the Pinetop to Goddard blocks in the Atoka Formation, as described by Stone and Haley (1984), Shideler (1970), Kramer (1933), and Ulrich (1927).

In Late Mississippian and Morrowan time, Bengalia was uplifted high enough to prevent deposition of the Wapanucka Limestone, as seen on seismic lines south and southeast of the Ti Valley fault, and proven by drilling of the Texaco-Lucy well in sec. 8, T. 2 S., R. 12 E., Atoka County, Oklahoma. The eastern Arbuckle Mountains, east of Boggy Depot, Atoka County, show the same relationships, with Wapanucka pinching out against the Arbuckles, and the Atoka onlapping over Goddard and older rocks, as seen in the Mobil-Stewart Ranch well in sec. 36, T. 4 S., R. 11 E., Atoka County, Oklahoma. With deposition of the Atoka Formation, Bengalia was buried, and growth faulting continued. The eastern Arbuckle Mountains were uplifted earlier than the Ouachita Mountains, and the Ouachitas were later thrust over various formations of the Arbuckles, commonly with the Womble Shale resting upon the Atoka and Goddard Formations. Bengalia acted as a buttress, over which the Ouachita rocks were thrust along bedding-plane faults. The Devonian and younger rocks of the shelf facies reacted to the Ti Valley thrust by also being displaced along bedding-plane faults ahead of the Ti Valley fault, giving rise to faults in the shelf facies, such as the Pine Mountain, Katy Club, and Choctaw faults. Bengalia was also compressed by these movements, and the granite basement reacted, giving rise to some deep-seated reverse faults, such as the Yourman fault in the Wilburton field.

Seismic lines show that the only primary fault in the frontal Ouachitas is the Ti Valley fault, which detached ~14 mi southeastward from the surface trace, primarily in the Mazarn–Womble Shales, ~5 mi deep, the angle of the sole-fault averaging ~20° to the surface. The overlying Bigfork Chert crops out at Stringtown, in Black Knob Ridge, and contains much gray limestone, which was described as Viola Limestone by Decker (1933). This hybrid Viola–Bigfork must have been present near the detachment, showing that the facies change must have been close to this area, ~14 mi southeast of the Ti Valley trace. There are horst blocks and anticlines ~4 mi deep in front of the detached area, apparently rooted below the Ti Valley fault. Other, higher faults occur within the Stanley–Jackfork sequence, such as the McGee Valley and Potapo Creek faults, which may not be connected to the Ti Valley fault.

The Potato Hills are not rooted directly below, but occur above a major bedding-plane fault ~12,000 ft deep, dipping S, originating in the Mazarn–Womble sequence ~3 mi south of the Albion anticline, and ending in the Stanley shale; this fault was named the Jackson Creek fault by Pitt and others (1982). About 10,000 ft lower is a detachment fault, probably the Ti Valley fault, possibly bringing Mazarn and younger rocks over Arbuckle and higher units. The American Quasar-Cabe well in sec. 11, T. 2 N., R. 20 E., was drilled to 15,512 ft in 1978; it was considered possible that the Arbuckle facies would be found below the upper major thrust, but it was not. A core near total depth contained lower Ordovician palynomorphs, probably from the Mazarn. The facies change from Arbuckle shelf to Ouachita basin probably occurs under the northern Potato Hills, ~25,000 ft deep.

Gravity data of the Ouachita Mountains show a series of rooted residual highs along Bengalia and northward, with a deep basin beneath the Lynn Mountain syncline, and a gravitational break between Bengalia and the Ti Valley detachment (Fig. 6). A magnetic break also corresponds with the gravitational break, falling off basinward. The Potato Hills area is a gravitational plateau—not rooted, but part of a block between two large gravity lows. The 110-milligal low of the Lynn Mountain syncline is one of the deepest in North America, and probably represents ~35,000 ft of Ouachita basin facies. Thick Precambrian rhyo-

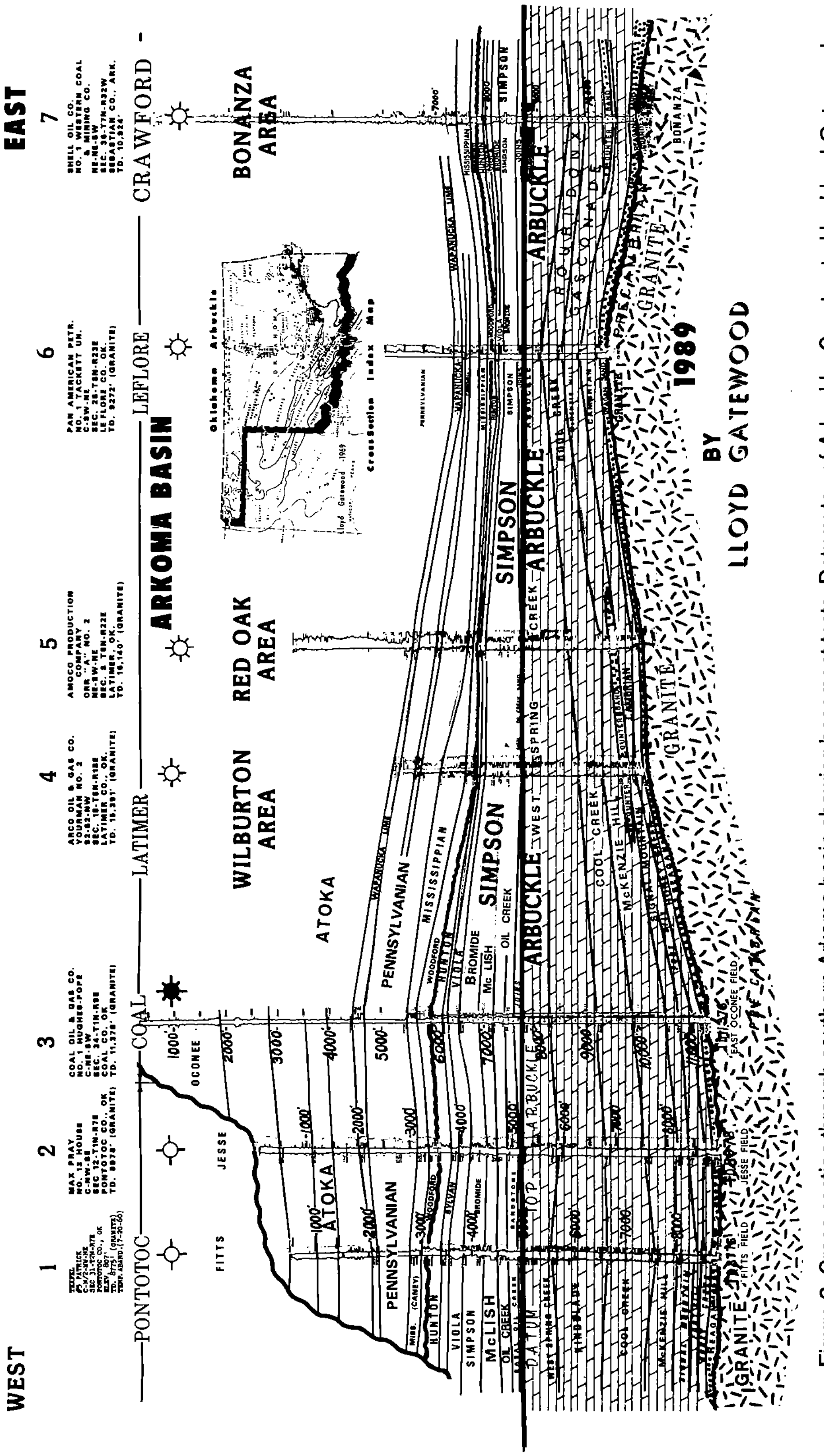

Figure 2. Cross section through southern Arkoma basin showing basement tests. Datum: top of Arbuckle. Constructed by Lloyd Gatewood.

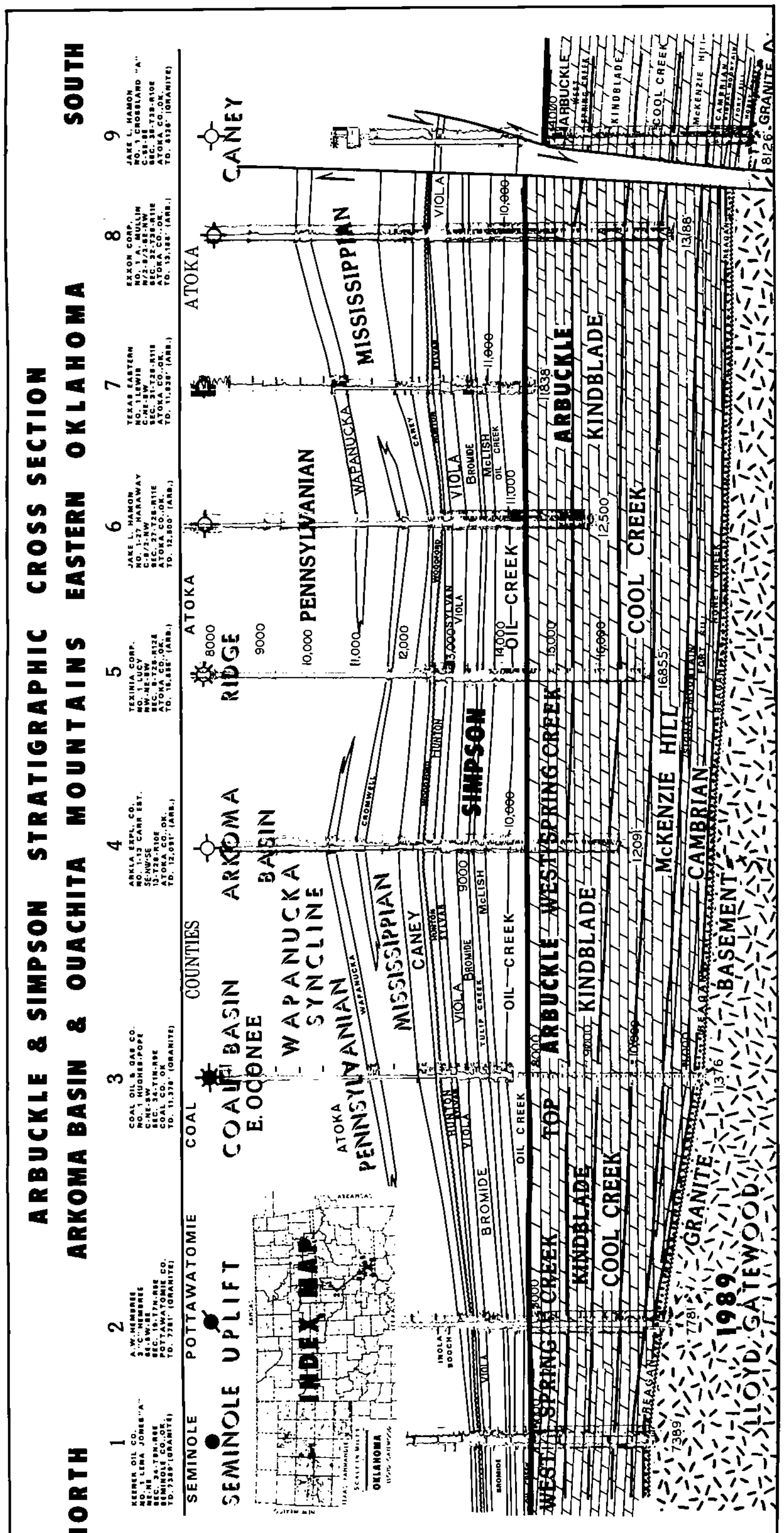

Figure 3. Cross section through Lehigh basin showing basement tests. Datum: top of Arbuckle. Constructed by Lloyd Gatewood.

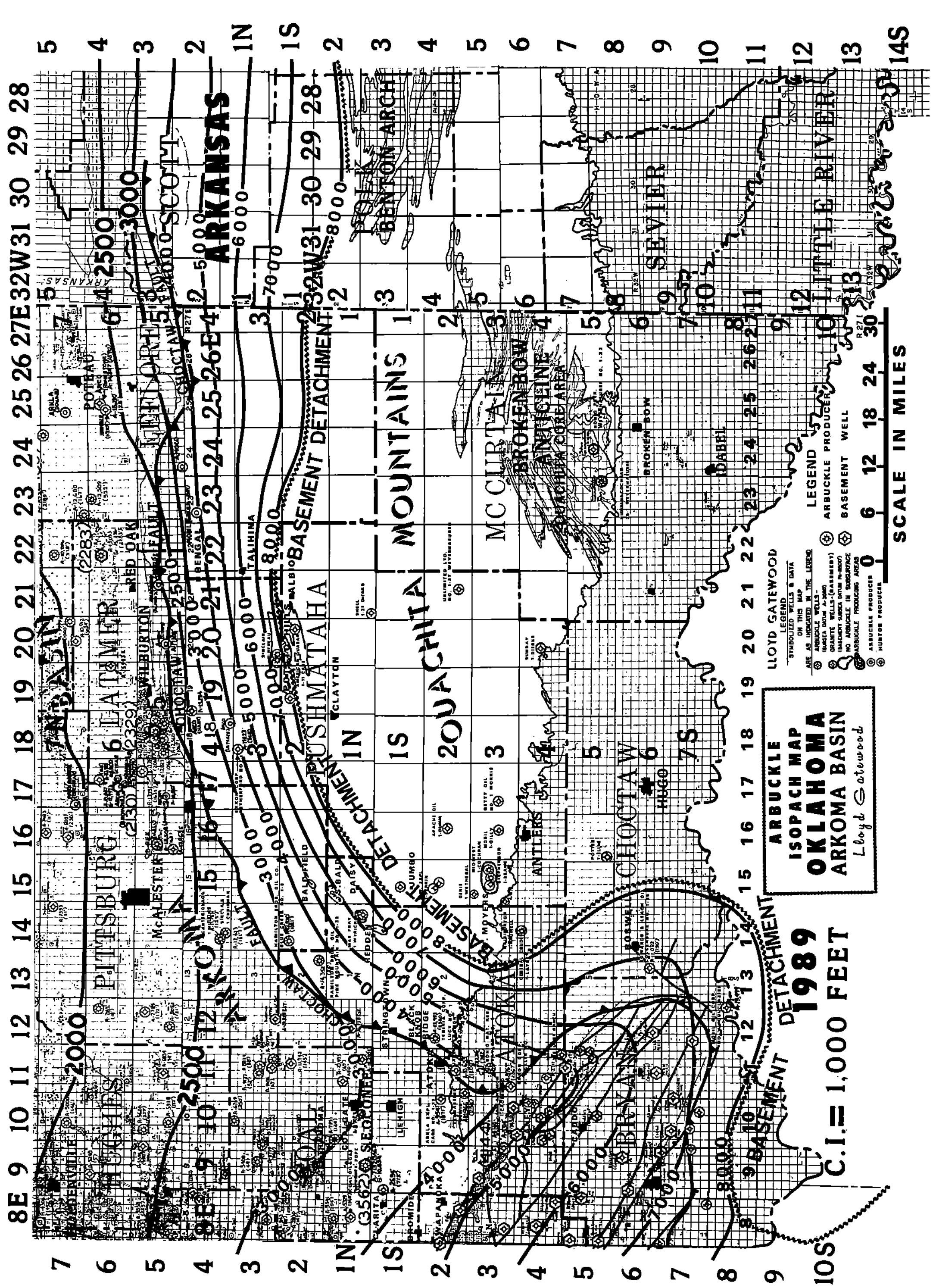

Figure 4. Isopach map of Arbuckle–Timbered Hills Groups beneath the Ouachita frontal belt. Constructed by Lloyd Gatewood.

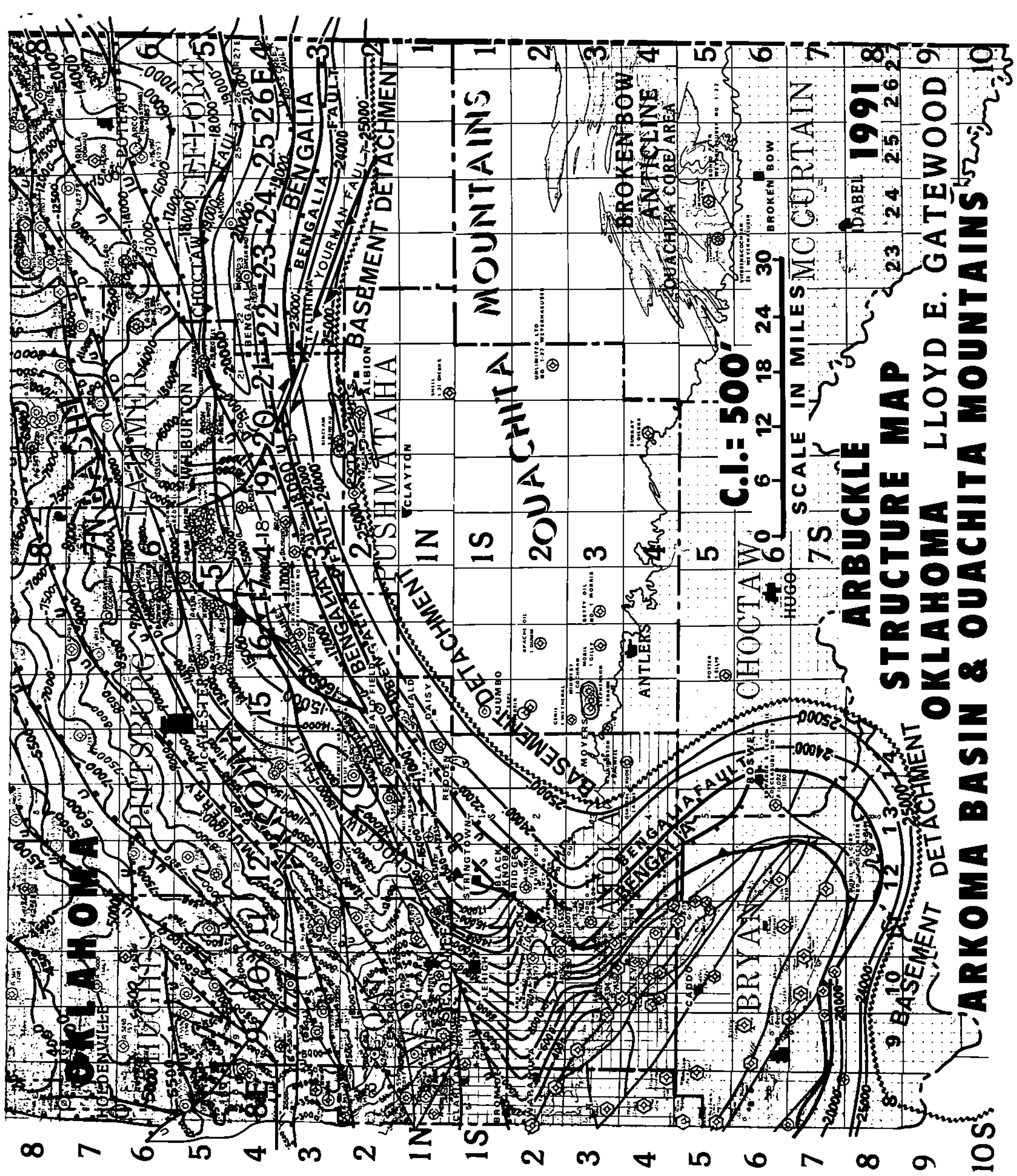

Figure 5. Structure map of top of Arbuckle Group beneath the Ouachita frontal belt. Constructed by Lloyd Gatewood.

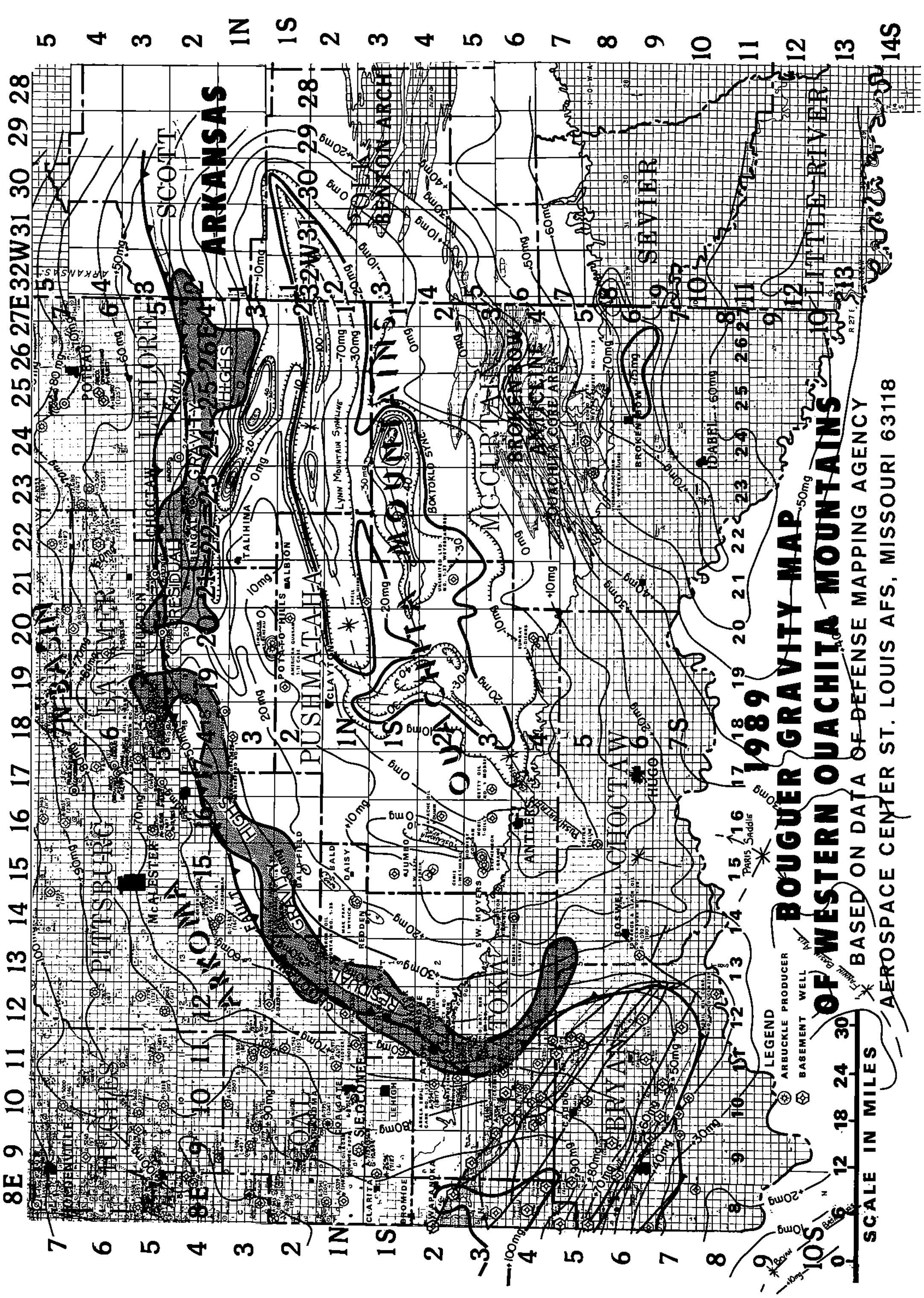

Figure 6. Bouguer gravity map of western Ouachita Mountains, with some residual high areas shown in the core area and Ti Valley area. Data from Defense Mapping Agency.

lites may also occur here, and they could be folded.

Magneto-telluric surveys have been made in parts of the Ouachita Mountains, but they have not been very reliable, because there is a gradual increase in density from Arbuckle to basement rocks.

In considering the Precambrian history of the Midcontinent structures, nearly all of the structures were probably formed along preexisting weak zones, which developed when the lower crust (sima) was formed. One theory is that the continents were formed from hot magma cooling, to form basalt or olivine basalt or "moon rock," and that large asteroids impacted this surface, forming large craters, causing lighter acidic magmas to extrude and cool along cracks, with rhyolite occurring at the surface and granite formed below. These rocks probably formed the shield areas of the continents, being thickest under the centers of the shields. Large circular weak zones or rills would have formed around the shields, with radial trenches extending outward from the centers. Continental accretion could occur later along the circular weak zones. Subsequent movements could occur at any time thereafter along the weak zones. With degassing of the magma, the atmosphere and oceans were formed.

The Gulf of Mexico could have been formed in early Precambrian time, as a small astrobleme, with circular weak zones, forming a proto-Ouachita belt, with radiating weak zones such as the Cuban trench, Appalachian belt, Mississippi River trend, Arbuckle alignment, Criner–Wichita alignment, and Rio Grande alignment (R. Donofrio, personal communications). The Sigsbee Deep could have been formed originally as an uplifted area, resulting from an impact and subsequent extrusion of lava, similar to Olympus Mons on Mars. The subsequent land mass has been called a Gulf plate by Beall (1973), and Llanoria by Miser (1921).

Weak zones under the Llano region of Texas could have been formed very early, with subsequent igneous rocks formed to build up a separate Texas plate (Beall, 1973). Much of the Midcontinent was probably formed from late Precambrian rhyolites extruding along preexisting weak zones, such as the Nemaha ridge, Spavinaw arch, proto-Ouachita Mountains, Arbuckle Mountains, and Wichita Mountains. The centers of active movements were the loci of subsequent granites in any weak zone, so that gradually the rhyolite would have been eroded away exposing granite along the early Precambrian weak zones. Low areas or basins would probably still have the rhyolite preserved, but the active weak zones would only have the granite equivalents preserved. The Washington County Volcanic Group of Denison (1982) probably represents the oldest Precambrian rhyolite in Oklahoma and Arkansas. This rock could be the basement rock beneath the Ouachita basin. The core area of the Ouachitas has quartz diorite, suggesting that this core trend was the locus of an early Precambrian weak zone.

A basal Cambrian sandstone rests upon the various Precambrian igneous rocks in the Midcontinent, variously named Reagan, Lamotte, and Hickory. The early active weak zones and later active tectonic zones would probably be the places where granitic rocks or intrusive rocks occur directly beneath the Cambrian sandstones.

The Arbuckles and Wichitas could have been formed by direct movements from below, coupled with movement of the Texas plate northeastward.

The Gulf plate could have collapsed in Desmoinesian time, creating the Ouachita belt. According to astrobleme theory, the Gulf plate would have been less dense than surrounding regions, and subsequently the volcanic land mass would collapse after a long erosional period. Thus, a subsequent movement along an early Precambrian weak zone could have formed the Ouachita Mountains. In middle Desmoinesian time and later, the Gulf plate would have been underwater, and the Moorehouse Formation and later rocks would have been deposited, with development of normal growth faults down-to-the-Gulf.

If such a Gulf plate or Llanoria existed, it could have been a land source for some of the sandstones such as the Blaylock and Hot Springs Sandstones in the Ouachitas. A platform (termed the Texarkana platform by Meyerhoff, 1967,1973) could have been formed around the plate. Cambrian–Ordovician carbonates could have been formed on this platform. Seismic sections across the Broken Bow core area of Oklahoma, extending into Lamar County, Texas, and into the Waco uplift of Hill County, Texas, 65 mi south of Dallas, show a smooth, arcuate reflection about 2–3 mi deep, with a broken sawtooth edge on the northwest side of the core area. A gravity high of +75 milligals occurs ~10 mi southeast of the Broken Bow core area, implying that this was the source detachment of the core area, and that the core area was probably thrust ~10 mi northwestward into the basin. Gravity anomalies are probably caused by differences in the sima, as explained by Nicholas and Rozendal (1975) and Kruger and Keller (1986).

In Lamar County, Texas, near Paris, just south of the Red River, the Hunt-Neely well was drilled in the core area of the Quachitas in 1975 to 18,721 ft. The sequence is similar to that of the exposed rocks in the Ouachitas as described by Honess (1923), Goldstein (1975), and Leander and Legg (1988). The section contains much black phyllite down to 14,130 ft, being mostly Womble and Mazarn. The ?Crystal Mountain Sandstone (or possibly Gunter Sandstone) occurs from 14,130 to

14,600 ft and is the reflective horizon on seismic. The next 4,121 ft is mainly carbonate rocks, with black phyllites, and with many veins of marble and quartz, belonging to the Collier Formation.

In Hill County, Texas, ~65 mi south of Dallas, on the Waco uplift, the Shell-Barrett well was drilled in 1967–68. The rocks in this well were like those in the Hunt-Neely well, with 9,691 ft of phyllite, slate, and quartzite, underlain by 6,050 ft of metamorphosed carbonates, with some black phyllite, and much muscovite mica. The basement rock was quartz diorite from 19,780 to 20,310 ft, considered to be Precambrian. The seismic reflector was the carbonate package.

In comparing these rocks in the wells to those of the outcrop, the thicknesses of the Womble–Mazarn below the Bigfork are similar to the thicknesses described in Arkansas, but do not correspond with the Broken Bow area of Honess, in Oklahoma. Is it possible that the Collier–Crystal Mountain–Mazarn–Blakely sequence of Honess is part of the lower Womble and Mazarn Shale only? It is quite possible that the quartzite–carbonate sequence in the Sohio-Weyerhaeuser well in sec. 22, T. 5 S., R. 24 E., McCurtain County, Oklahoma, at –11,058 ft, is actually the Crystal Mountain or Gunter and Collier, and that the overlying beds are Mazarn–Womble, thrusted and squeezed.

Oil, gas, and asphaltite occur in the Ouachitas in Ordovician through Pennsylvanian rocks. The black shales are excellent source rocks, and the fractured cherts, carbonates, and sandstones are good reservoir rocks.

Where the Ouachita–Arbuckle detachment occurs, rocks of the Arbuckle facies in the large anticlines and horst blocks are probably self-sourced, fractured, and capped by younger shales. Updip, these same rocks could be trapped against the fault face of Bengalia, against granite and younger rocks. Bengalia could be fractured, and where black shales are present, oil and gas could be present. No wells have been drilled on downside Bengalia, or in the detachment structures. The Sohio-Taylor well in sec. 15, T. 3 S., R. 11 E., and Texaco-Lucy in sec. 8, T. 2 S., R. 12 E., Atoka County, Oklahoma, are the only wells drilled into Bengalia, but not to basement. There is a need for rank wildcatting and large investments in the Ouachitas.

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PART II

ABSTRACTS AND SHORT REPORTS RELATED TO POSTER PRESENTATIONS

Oklahoma Geological Survey Circular 92, 1991

SPECIAL SEISMIC EXPLORATION TECHNIQUES FOR ARBUCKLE EXPLORATION

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INTRODUCTION

Seismic exploration in the Arkoma basin–Arbuckle trend is made more difficult by the geological environments above the Arbuckle. This brief report considers some techniques which can be used to improve the quality of seismic data over the Arbuckle trend. The goal of these techniques is the determination of the stratigraphy within the Arbuckle.

PROBLEMS CAUSED BY COMPLEX OVERTHRUSTING

A major problem in seismic exploration of the Arbuckle is data quality beneath the complex overthrust zones. As illustrated in Figure 1, some of the faults reach the surface, while other imbricate systems of faults occur over the basement faults. *The problem is not the faulting, but the material which is faulted.* The Spiro and Wapanucka zones are high-velocity zones which are capable of critically refracting signals. The effect of the critical refraction is exaggerated in complex structural zones. As a result, these zones limit the aperture through which the seismic energy can penetrate. The term "seismic window" is used in Figure 1 to indicate that only selected recording windows can be used over such faults as the Choctaw in southern Oklahoma. Blindly laying a conventional pattern over this type of geology can lead to serious problems, since it is possible that much of the seismic energy is critically refracted out of the recording window, i.e., the range of source–receiver offsets that are being used. The solution is to use well control to get an approximate idea of where these critically refracting horizons are located, and to use ray-trace modeling to identify the best recording windows.

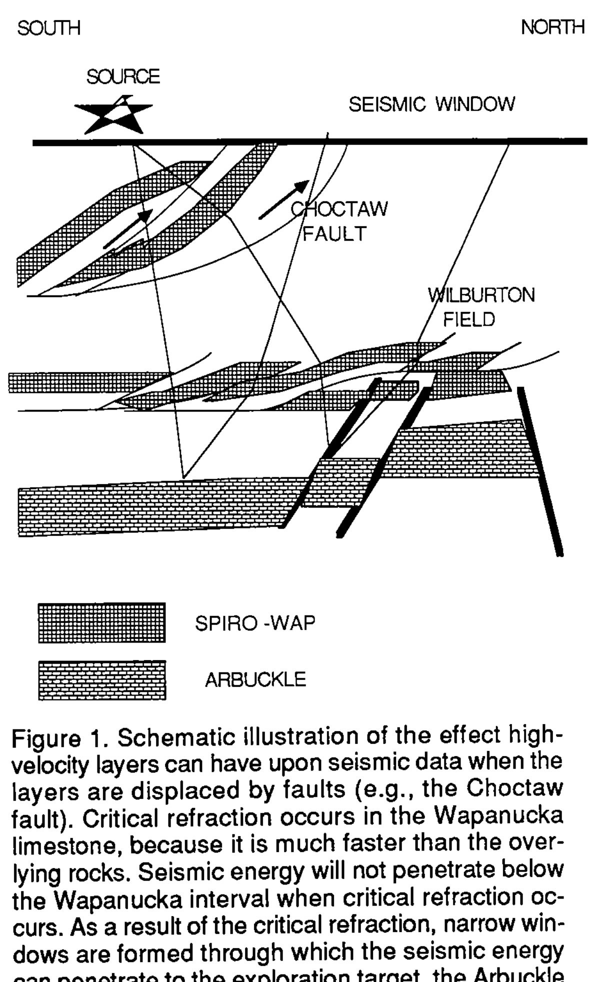

Figure 1. Schematic illustration of the effect high-velocity layers can have upon seismic data when the layers are displaced by faults (e.g., the Choctaw fault). Critical refraction occurs in the Wapanucka limestone, because it is much faster than the overlying rocks. Seismic energy will not penetrate below the Wapanucka interval when critical refraction occurs. As a result of the critical refraction, narrow windows are formed through which the seismic energy can penetrate to the exploration target, the Arbuckle Group. The window is indicated by the lines from the source to the reflector and back to the surface.

LOW-AMPLITUDE ARBUCKLE REFLECTION

Another problem facing seismic exploration in the Arbuckle trend is the relatively low amplitude of the Arbuckle reflection(s) compared to the high-amplitude reflections off the nearby Viola. The decrease in seismic amplitudes with increasing time is not smooth. Under these circumstances conventional gain functions will tend to emphasize the Viola at the expense of the Arbuckle reflector (Fig. 2). A proposed solution is to apply Automatic Gain Control (AGC) to selected windows below the Viola reflector and to reference those amplitudes to the Viola reflector. In this way, the "stratigraphic" variations within the Arbuckle can be mapped. The AGC technique can be used to map amplitude versus offset, which depends upon Poisson's ratio. Since the Poisson's ratios of dolomites and limestones are often different, the relative-amplitude studies should be useful for distinguishing limestones from dolomites. The ability to distinguish limestone from dolomite seismically would help to reduce the risks in Arbuckle exploration.

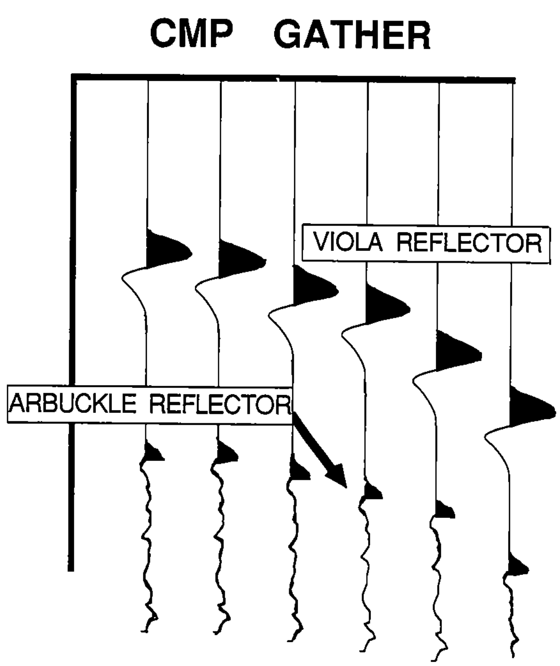

Figure 2. Schematic illustration of a problem with Arbuckle exploration. The amplitude of reflections from the Viola act to dominate the reflections from the Arbuckle. Standard approaches to amplitude gain will act to enhance the amplitude of the Viola reflection at the expense of the Arbuckle reflection. Automatic Gain Control (AGC) applied to a second window enclosing only Arbuckle reflectors can eliminate this problem.

LOW-FREQUENCY MAPPING OF BASEMENT

Fractured environments can scatter the high-frequency energy out of seismic signals. Low-frequency signals can sometimes be used to make a reliable interpretation of basement structures. Special techniques can be used with vibrators to extend the low-frequency end of the bandwidth. Simple filtering and processing of the low-frequency data yield a picture which is not as heavily influenced by the high-frequency scattering.

USING ALL THE WELL LOGS TO ESTIMATE SEISMIC VELOCITIES

In many geologic environments the quality of sonic logs can be undesirable. If sonic logs are deemed unreliable or are not available, other logs can often be used to estimate the velocities. In some lithologies the resistivity log can be used. In other environments the neutron-density crossplot along with the gamma ray can be used to estimate seismic velocities. These techniques can be used to "edit" logs and improve the match of synthetics to seismic sections. The Arbuckle exploration trend is an example where logs other than the sonic log have proven to be useful for seismic studies.

SUMMARY

The Arbuckle trend in Oklahoma is made more difficult by a number of geologic factors. For example, in the Ouachita mountains the Spiro and Wapanucka have considerable dip. These units can critically refract seismic signals and effectively create "windows" through which the target (e.g., the Arbuckle) can be imaged. The seismic "windows" caused by the overthrust sections require special recording geometries with restricted source–receiver offsets. A priori knowledge of the structures will have to be used to improve seismic imaging under these circumstances.

When high-amplitude reflectors are close to the reflecting horizon being explored, the reflecting horizon can be barely visible. Localized application of AGC can solve this problem. For mapping of basement, low-frequency signals offer advantages over broad-band signals. The sacrifice of bandwidth may be worth the gain in signal strength. Application of sonic logs to estimation of synthetics can be facilitated by using other logs.

AULACOGEN COLLAPSE: A MODEL FOR THE COMPRESSIVE STAGE OF AULACOGEN EVOLUTION

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INTRODUCTION

Aulacogens represent the scars on the continental crust where continental breakup did not succeed. Arguments are presented in this brief summary of our research to suggest that the relaxation response of the lower crust to an intruded body with higher density ultimately leads to *aulacogen collapse*, i.e., *a circumstance where the intruded mass begins to drop and the crust acts to fill in the void.* The process is very much like a ball dropping in a fluid, where the lower crust and upper mantle represent the fluid and the ball represents the intruded body; however, the shallow crust responds more elastically and exhibits faulting and mountain-building. We therefore propose that the collapse of the southern Oklahoma aulacogen produced the compression responsible for building the Arbuckle and Wichita mountain systems. The plate collision responsible for the Ouachita mountain system cannot easily be used to explain the timing and sense of faulting observed in the Arbuckle and Wichita Mountains. This paper suggests that the active lifetime of aulacogens can be very long (tens to hundreds of million years). As a result, the stratigraphic record over ancient aulacogens can be used in a study of the rate of compression, fault activity, and earthquake potential over active aulacogens (e.g., the southern Oklahoma aulacogen and the Reelfoot rift, respectively). The rate of aulacogen collapse will be determined by the initial conditions and the effective viscosity of the lower crust.

WHEN DOES AN AULACOGEN CEASE TO BE ACTIVE?

Garner and Turcotte (1984) argued that the amount of subsidence observed in the Anadarko basin is more than would be predicted by the uplift of the Wichita mountains or thermal cooling of the lithosphere. Instead, they suggested that a heavy mass must be attached to the bottom of the crust/lithosphere in order to cause the current configuration of the Anadarko basin. Weissel and Karner (1989) suggested that aulacogen topography varies over a period of 100 m.y. due to the temperature history of the lithosphere. Applying ideas proposed by McKenzie (1978), both the Weissel and Karner (1989) and the Garner and Turcotte (1984) papers related change of flexural rigidity of the lithosphere to the temperature history of the lithosphere. Another factor affecting aulacogens over the same time scales (10–100 million years) will be erosion. The time required for the tectonic activity of an aulacogen to cease is unknown. However, there are indications that aulacogens have compressive stages long after their initial rift and subsidence stages (Hoffman and others, 1974). If these later events are caused by the aulacogens, then it would appear that aulacogens can be active for 100 million years or more!

DETERMINATION OF AULACOGEN HISTORY

To add to the confusion over aulacogens, subsequent tectonic events can overlap the evolutionary history of aulacogens. The southern Oklahoma aulacogen is a case in point. Most studies (e.g., Perry, 1989) of aulacogens assume that the compressional events and further subsidence recorded in the overlying sediments are due to other tectonic events. For the southern Oklahoma aulacogen, the "other tectonic event" was the collision with the African–South American plate, which ultimately is responsible for the Ouachita and Appalachian thrust systems.

Figure 1 is a schematic diagram of the plate boundary between North America and the African–South American plate. The actual location of the plate boundary was most likely south and southeast of the location indicated. The indicated subduction direction is based upon evidence of volcanism south of the present Ouachita thrust (Perry, 1989). The timing of events and nature of the tectonics are not easily explained by plate collision. For example, the Wichita orogeny began during the Mississippian. The Arbuckle orogeny followed after the Wichita orogeny, since there is evidence that the Ardmore and Marietta basins were a single basin before Pennsylvanian time (Ham and others, 1964). The Ouachita orogeny represented the closing stage of plate collision.

The timing of the Wichita and Arbuckle orogenies could be explained by separate detachments in the lower crust and upper mantle; however, as illustrated in Figure 1, the geometry of the plate boundary would place the Wichita–Arbuckle system under tension, rather than compression. The incorrect principal stress direction, i.e., tension rather than compression, appears to be the strongest argument against plate collision as an explanation of the Wichita and Arbuckle orogenies.

AULACOGEN COLLAPSE

Much of the tectonic deformation in the Wichitas and Arbuckles can be explained in terms of what is thought to occur when an aulacogen is formed. Initially, mantle material is injected into the lithosphere and crust by a mantle plume. The injection must lift and thin the lithosphere while acting to stretch the thin, elastic layer of the crust. Rifting typically forms an asymmetric graben pattern such as that shown in Figure 2 (Rosendahl, 1987). The dark lines are used in Figure 2 to indicate the outline of material which has been intruded into the lower crust. The half-grabens represented the future sites of the Wichita and Arbuckle Mountains. As illustrated in the schematic cross sections shown in Figure 3, the half-grabens that formed over the incipient rift represented vessels for collecting the magmas. Although the term half-graben was not used, the work of Ham and others (1964) lends observational support to the initial rift configuration illustrated in Figure 3. The body of mantle material injected into the lower crust is identified in the figure as an intrusive body. The main fault for each half-graben is indicated near opposite edges of the body intruded at depth. Based upon a gravity model by Pruatt (1986), the lower crust under the Anadarko basin has an injected body with a density anomaly that approximates the outline given in Figures 2 and 3.

When solidified, the intruded bodies are the strongest parts of the crust, i.e., the igneous rocks act as a buttress. The faulting due to later tectonic events will be controlled by the emplacement of these intruded bodies into the shallow crust. However, as suggested by the cartoon in Figure 3, most of the injected material is below the center of the Anadarko and Ardmore–Marietta basins. The faults associated with the half-grabens probably acted as conduits for the magmas which accumulated in the half-grabens.

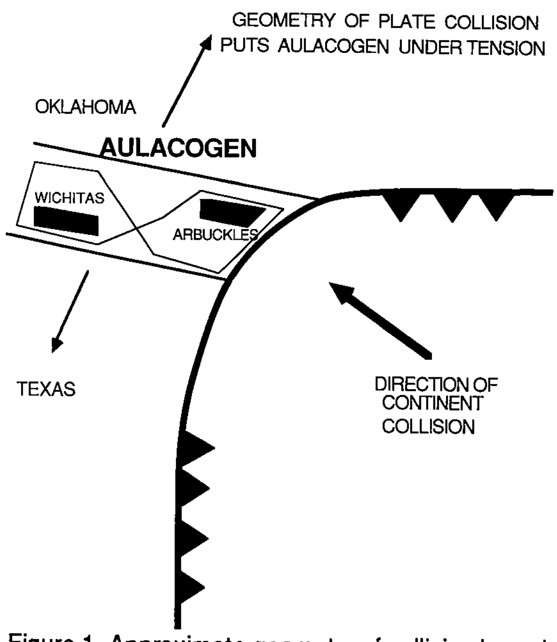

Figure 1. Approximate geometry of collision boundary between North America and the African–South American plate. Actual plate boundary was farther south and east than illustrated in this figure. The plate collision along the boundary shown would have placed the Oklahoma aulacogen under tension. The compression responsible for mountain-building over the aulacogen must have been caused by forces associated with the aulacogen, rather than the plate collision.

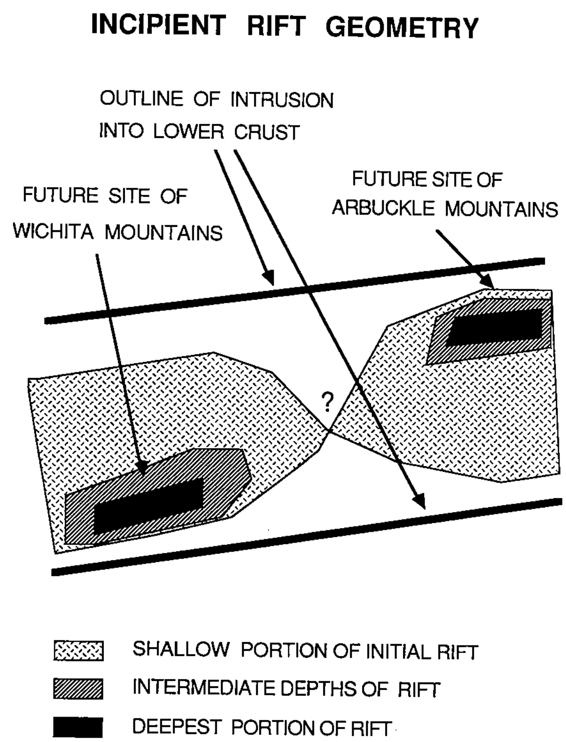

Figure 2. Asymmetric half-graben rift pattern responsible for the birth of the Wichita and Arbuckle Mountains. The deeper parts of the rift zones accumulated igneous units which became the cores of the respective mountain systems.

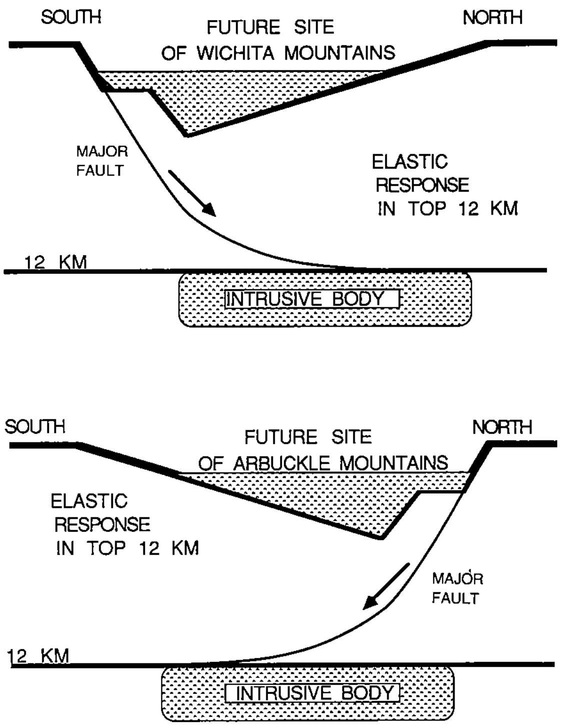

Figure 3. Cross sections of the half-grabens shown in Figure 2. Each of these grabens is associated with a major normal fault which may have acted as a conduit from a deeper intrusive body (inferred from gravity studies) and from the near-surface igneous suites. The shaded bodies represent the igneous units involved. The exact geometry of the faulting is unknown. See the paper by Weissel and others (1989) for a discussion of possible rift fault geometries.

Once the abnormal pressure at the base of the lithosphere declines, i.e., the mantle plume responsible for the rift dies, the extended crust and lithosphere fall back into isostatic equilibrium with the asthenosphere. The flexural response of the lithosphere acts to exaggerate the half-graben topography (Weissel and Karner, 1989). Further modification of the rift topography occurs over a 100-m.y. period due to temperature effects upon the flexural rigidity of the lithosphere (Weissel and Karner, 1989).

The injected mass of mantle material will be supported for a period of time which depends upon the rheological properties of the crust and upper mantle; however, as the injected body cools and the lower crust and upper mantle begin to relax, the intruded mass will begin to drop. The time for this process depends upon the effective viscosity of the lower crust and upper mantle. In order to estimate the effective viscosity of the lower crust–upper mantle system, the simple model given by Vening Meinesz (see Heiskanen and Vening Meinesz, 1958) for the response of a viscous half-space to a surface load can be used. The relaxation or response time of their model may be computed using

$$\eta = \frac{\rho g t}{2\alpha}$$

where η = viscosity of half-space;
g = acceleration due to gravity (980 cm/sec/sec);
t = relaxation time

and

$$\alpha = \sqrt{\left(\frac{2\pi}{L_x}\right)^2 + \left(\frac{2\pi}{L_y}\right)^2}$$

where L_x = length scale of surface load in x-direction; and
L_y = length scale of surface load in y-direction.

Using a surface load with dimensions of 200 mi by 50 mi and relaxation times on the order of 100–200 m.y. yields an order-of-magnitude estimate of the effective viscosity of the lower crust of 10^{24} poises.

In Figure 4 the basic idea of aulacogen collapse is presented. The intruded mass is held for a finite period of time until the viscous behavior of the

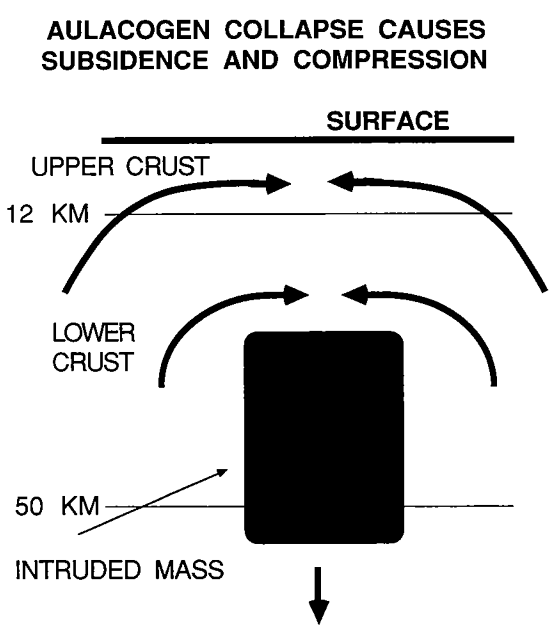

Figure 4. Aulacogen collapse. The igneous mass intruded into the crust under the aulacogen eventually begins to sink as the lower crust responds to the additional load. The part of the crust above the mass acts to fill the void. Cataclastic flow will occur in the lower crust, while faulting and mountain-building occur in the elastic upper crust (top 12 km).

lower crust–upper mantle system begins to respond to the extra load. For the southern Oklahoma aulacogen, the mass appears to have been supported for a period of 170 m.y. or more (Middle Cambrian–Early Mississippian). As the mass drops toward equilibrium at greater depths, the lower crust and upper crust act to fill the void above the dropping mass; thermal contraction could play a role in exaggerating the effect. If the injected mass is *close to the surface* of the earth, the initial response should be subsidence, followed by a combination of vertical and horizontal forces. This type of force system is thought to be responsible for the birth of the Wichita Mountains (Evans, 1979). As the injected mass moves deeper below the surface, the forces acting on the surface weaken and become more horizontal. This is the type of force system which can cause the folding observed in the Arbuckles.

Thus, aulacogen collapse can be used to explain the different tectonic environments of the Wichita and Arbuckle systems if the injection was deep beneath the Arbuckles and shallow below the Wichitas. This hypothesis is supported by the cooler temperatures in the Ardmore–Marietta basin determined by vitrinite reflectance (Cardott and Lambert, 1985). The aulacogen collapse model proposed in this paper can be quantified using our current understanding of the stratigraphy (e.g., Gatewood, 1978a,b).

SUMMARY

The common explanation for the Wichita–Arbuckle system, i.e., plate collision, has some obvious problems with timing and principal-stress directions. This paper suggests instead that as the mass of mantle material injected into the crust drops, the crust moves to fill in the void. The process is called *aulacogen collapse.* Aulacogen collapse can be used to explain the tectonic sequence and principal-stress directions responsible for the formation of the Wichita–Arbuckle system. The real test of the model will be a quantitative comparison of the structure and stratigraphy over the aulacogen with the timing and principal-stress directions predicted by the model.

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OIL AND GAS IN ARBUCKLE GROUP RESERVOIRS IN OKLAHOMA: AN NRIS PERSPECTIVE

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INTRODUCTION

The Natural Resources Information System of Oklahoma (NRIS) is an integrated system of computerized information related to the geological resources of Oklahoma. For petroleum resources, it includes oil, associated gas, non-associated gas, and condensate production data, and producing reservoirs. These data can be accumulated, mapped, and graphed by lease (township, range, and section), field, county, any of five regional subdivisions, or the entire state. Well-history data are being accumulated for the state, including such information as drilled and completed footage and depths, formation tops, drill-stem tests, and initial production, among others. Data integration provides for the representation of information in a variety of tabulated and graphic formats. Computer graphics have been reduced variously from page-size illustrations for compatibility with the format of this publication.

ARBUCKLE GROUP RESERVOIR OCCURRENCE

Production from the Arbuckle Group is employed to demonstrate some of the capabilities of NRIS. Arbuckle reservoirs produce oil and/or gas in all five major petroleum provinces of Oklahoma: the Anadarko, Ardmore, and Arkoma basins, the Cherokee shelf, and the Nemaha uplift. Arbuckle reservoirs also produce hydrocarbons from 129 fields located in 42 of Oklahoma's 73 oil- and gas-producing counties (Fig. 1). Arbuckle

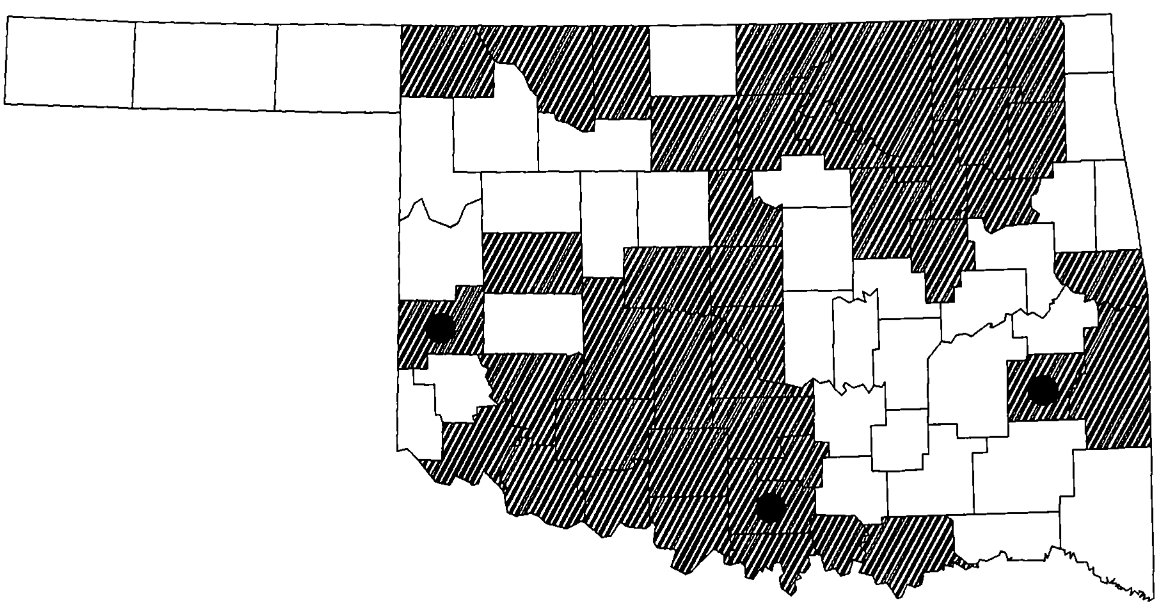


Figure 1. Oklahoma counties having oil and gas production from the Arbuckle Group. Hachures indicate production. For counties identified here, but not listed on Tables 1 or 2, production is from outside an assigned field limit (Alfalfa, Cotton, Mayes, and Sequoyah Counties). Dots identify Beckham, Carter, and Latimer Counties, respectively, west to east. Source: NRIS data files.

TABLE 1. — FIELDS PRODUCING FROM THE ARBUCKLE GROUP

(IF 1983–88 ARBUCKLE PRODUCTION >1,000 BBL OR 10,000 MCF)

(Liquids production in bbl; gas production in Mcf)

COUNTIES	FIELD CODE AND NAME		1983-88 LIQUIDS PRODUCTION			1983-88 GAS PRODUCTION		
			ARBUCKLE	ARB+OTHER	FIELD TOTAL	ARBUCKLE	ARB+OTHER	FIELD TOTAL
OSAGE	012657	ALMEDA	.	.	65,199	11,748	.	475,023
COMANCHE	021903	APACHE SOUTH	.	.	9,444	220,387	.	322,990
CADDO	021970	APACHE TOWNSITE	9,431	.	10,806	1,976,566	.	2,667,652
OSAGE	029599	ATLANTIC	258,266	233,104	2,577,898	.	.	555,927
OSAGE	031335	AVANT	.	45,708	1,538,743	37,992	.	1,145,361
OSAGE	031366	AVANT WEST	.	41,552	608,242	142,127	346,532	1,266,832
CRAIG	038304	BANZET NW	2,767	.	2,767	.	.	6
OSAGE	040001	BARNSDALL	.	3,989	607,098	92,759	142,985	1,240,755
OSAGE	040063	BARNSDALL WEST	.	.	239,895	372,638	.	519,775
OKMULGEE	052239	BEGGS DISTRICT	17,227	.	953,087	11,417	.	2,948,066
NOBLE	064508	BILLINGS	27,374	60,962	443,244	22,859	30,913	193,462
OSAGE,TULSA	065407	BIRD CREEK	1,239	.	2,063,404	.	.	1,248,012
KAY	088178	BRAMAN NORTH	40,961	46,439	162,207	.	.	33,694
GARFIELD	095091	BROWN	22,009	.	61,949	.	.	469,111
KAY,OSAGE	103663	BURBANK	40,939	.	9,658,645	.	.	.
OSAGE	118328	CANYON CREEK	16,107	.	70,008	.	.	.
ROGERS	125644	CATALE DISTRICT	2,560	.	3,438	.	.	.
CARTER	164280	COTTONWOOD CREEK	493,663	.	493,663	366,929	.	366,929
BRYAN,MARSHALL	174325	CUMBERLAND	.	.	972,397	558,646	14,240	16,489,891
CREEK,PAYNE	175162	CUSHING	811,607	811,100	12,353,219	396,910	297,621	9,939,001
CUSTER	175472	CUSTER CITY NORTH	.	.	93,909	1,801,215	.	29,906,110
MURRAY	182615	DAVIS WEST	34,291	.	34,291	15,162	.	15,162
KAY	192479	DILWORTH DISTRICT	45,793	208,754	1,153,285	9,359	103,792	4,781,351
OSAGE	196518	DOMES-POND CREEK	3,290	.	3,649,514	18,705	38,771	3,715,421
LOVE	226359	ENVILLE SW	22,719	.	801,451	7,219	.	14,785,261
LOVE	226390	ENVILLE WEST	179	.	29,516	18,830	.	2,472,133
GARVIN,STEPHENS	226514	EOLA-ROBBERSON	1,085	6,271	6,315,474	.	.	34,468,766
OSAGE	244017	FLAT ROCK	182,606	124,105	783,563	.	277,204	590,167
OSAGE,WASHINGTON	251814	FORTY FIVE	.	.	173,017	145,569	.	487,159
OSAGE	273161	GILLILAND	42,748	24,970	1,181,508	.	.	328,765
GARVIN,GRADY,MCCLAIN	278587	GOLDEN TREND	65,796	7,921	25,666,199	581,019	405,233	213,875,107
KIOWA	281966	GOTEBO DISTRICT SOUTH	114	.	20,881	25,881	.	50,708
OSAGE	304315	HAPPY HOLLOW NE	30,085	.	30,085	3,221	.	3,221
CARTER,JEFFERSON	312799	HEALDTON	951,855	.	11,217,087	1,627,726	.	2,978,321
CARTER	319464	HEWITT	13,591	.	13,936,613	.	.	596,423
MCINTOSH,OKMULGEE	326340	HOFFMAN	3,659	.	22,746	4,832	.	1,516,409
CARTER	329874	HOMER	72,528	.	85,094	.	.	.
OSAGE	330246	HOMINY	51,506	14,274	113,641	.	.	14,508
OSAGE	330277	HOMINY EAST	8,036	.	113,529	.	.	.
OSAGE	330385	HOMINY WEST	61,899	.	93,259	18,708	.	18,708
GARVIN,MURRAY	331486	HOOVER SE	48,740	.	51,327	158,956	.	158,956
OSAGE	000666	HULAH	.	.	.	55,689	.	55,689
CARTER,JEFFERSON	362665	JOINER CITY	19,098	.	864,629	24,446	.	6,566,265
CARTER,LOVE	362696	JOINER CITY SE	32,996	.	296,524	48,115	.	947,537
CREEK,OSAGE,PAWNEE, TULSA	377362	KEYSTONE	1,602	.	686,689	.	.	703,396
KIOWA	385563	KOMALTY DISTRICT	34,158	748	311,911	501	.	155,284
LOGAN	401407	LANGSTON	5,610	.	10,817	.	.	.
COMANCHE	405623	LAWTON DISTRICT	2,507	33,292	146,877	6,716	22,804	198,079
JEFFERSON,STEPHENS	422040	LOCO DISTRICT	12,471	8,171	2,420,622	.	10,394	352,669
HARPER,WOODWARD	429697	LOVEDALE	614	13,171	57,417	20,539	490,944	12,802,208
HARPER,WOODS	429790	LOVEDALE NW	1,411	.	417,525	44,179	.	17,179,932
OSAGE	439535	MADALENE EAST	5,874	.	104,777	.	.	74,351
MARSHALL	439876	MADILL	1,161	.	411,358	.	.	9,168,353

TABLE 1. — *Continued*

COUNTIES	FIELD CODE AND NAME		1983-88 LIQUIDS PRODUCTION ARBUCKLE	ARB+OTHER	FIELD TOTAL	1983-88 GAS PRODUCTION ARBUCKLE	ARB+OTHER	FIELD TOTAL
BECKHAM	455562	MAYFIELD WEST	913	.	255,653	13,515,331	.	101,452,142
OSAGE	460933	MCINNIS	20,826	.	20,826	.	.	.
MCINTOSH,OKMULGEE	485836	MORRIS DISTRICT	31,470	.	614,082	.	.	3,963,544
CLEVELAND	511905	NORMAN NORTH	55,633	.	1,193,648	42,917	.	3,763,295
CLEVELAND,OKLAHOMA	522270	OKLAHOMA CITY	34,670	.	3,487,972	19,759	.	8,880,058
OSAGE	529098	OSAGE CITY DISTRICT EAST	47,678	.	256,090	.	.	.
OSAGE	544328	PAWHUSKA NW	2,202	.	2,202	.	.	.
OSAGE	549312	PENN CREEK	28,619	1,990	37,772	42,701	.	52,676
OSAGE	578988	QUAPAW	.	.	605,580	18,753	.	942,993
WAGONER	592289	RED BIRD DISTRICT	.	10,839	81,629	33,454	59,698	634,322
OSAGE	590057	RED BLUFF DISTRICT NORTH	8,221	.	8,221	.	.	.
CREEK,OSAGE,TULSA	590832	RED FORK DISTRICT	9,573	.	568,848	.	.	408,537
LATIMER,LE FLORE	591700	RED OAK-NORRIS	.	.	.	159,717	.	167,016,216
CREEK,TULSA	630455	SAND SPRINGS DISTRICT	11,892	.	141,017	.	.	1,370,536
OSAGE	651818	SHIDLER NE	1,088	.	1,088	.	.	.
CARTER,GARVIN, STEPHENS	653086	SHO-VEL-TUM	46,849	.	109,209,115	188,315	.	236,309,473
JEFFERSON	670795	SPRING	36,026	.	236,072	.	.	.
CARTER	671508	SPRINGER	30,671	.	54,854	.	.	875,198
MUSKOGEE,TULSA, WAGONER	681909	STONEBLUFF DISTRICT	21,884	19,130	480,538	.	.	665,515
PAWNEE	701728	TERLTON NORTH	70,788	1,949	596,399	18,104	9,332	1,499,339
JACKSON,TILLMAN	709525	TIPTON	1,206	.	230,020	.	.	.
OSAGE,TULSA	721958	TURLEY DISTRICT	5,841	38,779	327,858	46,812	117,260	417,111
LATIMER,PITTSBURG	767826	WILBURTON	.	.	.	17,931,566	.	169,227,653
OSAGE	768508	WILDHORSE	3,160	.	742,075	.	.	49,541
OSAGE	768539	WILDHORSE NE	3,061	.	25,463	.	.	.
OSAGE	768601	WILDHORSE SOUTH	23,556	.	43,205	120,681	.	263,115
OSAGE	768632	WILDHORSE WEST	19,873	.	19,873	.	.	.
CANADIAN	789064	YUKON NW	6,826	4,230	519,781	279,828	74,667	24,212,566

Note: Production from Arbuckle does not necessarily occur in all counties in which the field is located. Arb + other is comingled production; field code from OGS GM-28, map of Oklahoma oil and gas fields.

production by field is shown in Tables 1 and 2. Table 1 lists cumulative Arbuckle production (1983–88) by field, for all fields in which there are some wells that produce only from Arbuckle reservoirs. Table 2 identifies all other fields that produce from Arbuckle, all of which comingle Arbuckle production with that from shallower reservoirs.

Three counties (Beckham, Carter, and Latimer) are provided as examples for greater detail of Arbuckle hydrocarbon production (Fig. 1).

ANADARKO BASIN

Beckham County is located in the central part of the deep Anadarko basin, where the Arbuckle Group commonly lies at depths exceeding 24,000 ft. Virtually all of the Arbuckle production in the county is from the Mayfield West field, although Arbuckle also produces in the Carter East field (Fig. 2). For Mayfield West, gas and condensate production from Desmoinesian sandstone and arkosic sandstone, Hunton, Simpson and Viola, and Arbuckle Group reservoirs are shown graphically for 1983 through 1988 (Figs. 3,4), and as a diagram for the six-year cumulative, showing proportional distribution of production (Figs. 5,6).

ARDMORE BASIN

Cottonwood Creek and Healdton fields are among nine fields that produce oil and associated gas from Arbuckle strata in Carter County, in the Ardmore basin (Fig. 7). Cottonwood Creek field was discovered in November 1987 and had grown to 12 producing leases in 9 quarter sections by

TABLE 2. — ADDITIONAL FIELDS WITH REPORTED ARBUCKLE PRODUCTION, 1983–88

COUNTIES	FIELD CODE AND NAME	OIL &/OR GAS
CARTER	023864 ARDMORE SW	OIL
OSAGE	037645 BANDWHEEL	OIL & GAS
OSAGE	039257 BARKER	OIL
OSAGE,WASHINGTON	041458 BARTLESVILLE-DEWEY	OIL & GAS
CREEK,TULSA	067396 BIXBY-JENKS DISTRICT	OIL
OSAGE	068182 BLACK DOG	OIL
OSAGE	068213 BLACK DOG DISTRICT SOUTH	OIL & GAS
KAY	069608 BLACKWELL	OIL
OSAGE,PAWNEE	082613 BOSTON	OIL
GARVIN	087651 BRADY	OIL
HARPER	101524 BUFFALO NORTH	OIL & GAS
GARVIN	107786 BUTTERLY NE	OIL & GAS
WASHINGTON	116623 CANARY	OIL
BECKHAM	122092 CARTER EAST	OIL & GAS
CRAIG	003973 CENTRALIA NE	OIL
OSAGE,PAWNEE	148237 CLEVELAND	OIL & GAS
ROGERS,TULSA	154205 COLLINSVILLE	GAS
JEFFERSON	162244 CORNISH SW	OIL
OSAGE	213811 EDGEWOOD	OIL
OSAGE	213837 EDGEWOOD SOUTH	OIL
OSAGE	226049 ENTERPRISE	OIL
OSAGE	245009 FLESHER EAST	OIL
COMANCHE	250822 FORT SILL NORTH	OIL & GAS
TILLMAN	256201 FREDERICK SE	OIL
GARFIELD	264596 GARBER	OIL & GAS
MURRAY	316382 HENNEPIN EAST	OIL
OSAGE	330308 HOMINY FALLS	OIL
KAY	336354 HUBBARD	OIL
OSAGE	355610 JAVINE DISTRICT	OIL
GARVIN	411940 LEWIS	OIL
OSAGE	443751 MANION	OIL & GAS
COMANCHE	472745 MIDWAY	OIL
OSAGE	499758 NAVAL RESERVE	OIL & GAS
OSAGE	499851 NAVAL RESERVE SOUTH	OIL
OSAGE	501797 NELAGONEY	GAS
CLEVELAND	509797 NOBLE NW	OIL & GAS
NOWATA	513982 NOWATA-CLAGGETT	OIL & GAS
OSAGE	529284 OSAGE-HOMINY DISTRICT	OIL
JEFFERSON	529377 OSCAR	OIL
OSAGE,TULSA	535555 PAGE DISTRICT	OIL
MCCLAIN	545010 PAYNE	OIL & GAS
OSAGE	553063 PETTIT	OIL
WAGONER	671756 SPRINGHILL WEST	OIL & GAS
OSAGE	690093 SUNSET	OIL
LOVE	703759 THACKERVILLE NORTH	OIL & GAS
OSAGE	707324 TIDAL OSAGE	OIL & GAS
MCCLAIN	750271 WASHINGTON DISTRICT	OIL
OSAGE	778871 WOOLAROC	OIL & GAS

Note: Field code from OGS GM-28, map of Oklahoma oil and gas fields.

June, 1989. Figure 7 illustrates only the discovery quarter section, as nomenclature lags behind development. Field development is illustrated through geographic extension and growth in production from the last quarter of 1987 through the second quarter of 1989 (Figs. 8,9). The discovery well flowed freely prior to the second quarter of 1988, thus the field production was unusually high until that well was brought under control.

Healdton field has produced from the Arbuckle since 1960, although shallow production was discovered there in 1913. The field presently produces from at least 14 reservoirs, which are separated into Cisco, Hoxbar, Deese, Simpson, and Arbuckle Groups for the purposes of graphic illustration of proportional oil and gas production. Only gas is produced from sandstones in the Cisco Group. The proportional distribution of 1988 oil production by reservoir is displayed in Figure 10, with variation for the period 1983–88 shown in Figure 11. Figures 12 and 13 display the comparable data for gas production by reservoir.

ARKOMA BASIN

The Arbuckle was established as a natural-gas reservoir in the Wilburton gas field (Fig. 14), in the deep part of the Arkoma basin in late 1987, although production was first discovered there in the Hartshorne Sandstone (Desmoinesian) in 1929. It is the only production from the Arbuckle Group in the basin. Production for the Latimer County part of the field is separated into sandstones within the Atoka Formation, the Spiro sandstone at the base of the Atoka Formation,

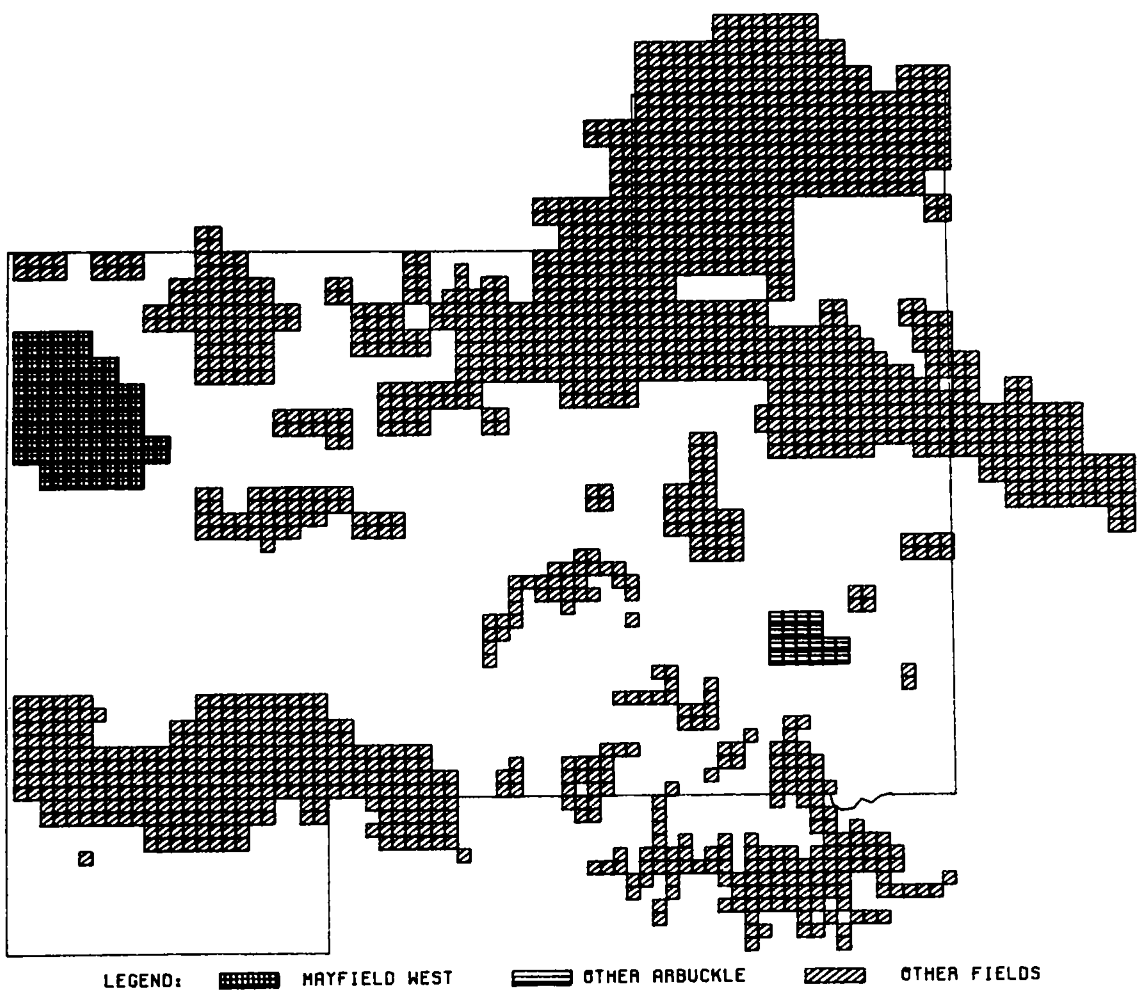


Figure 2. Oil and gas fields in Beckham County and vicinity, Oklahoma. Each map unit is ¼ section. Source: NRIS data files.

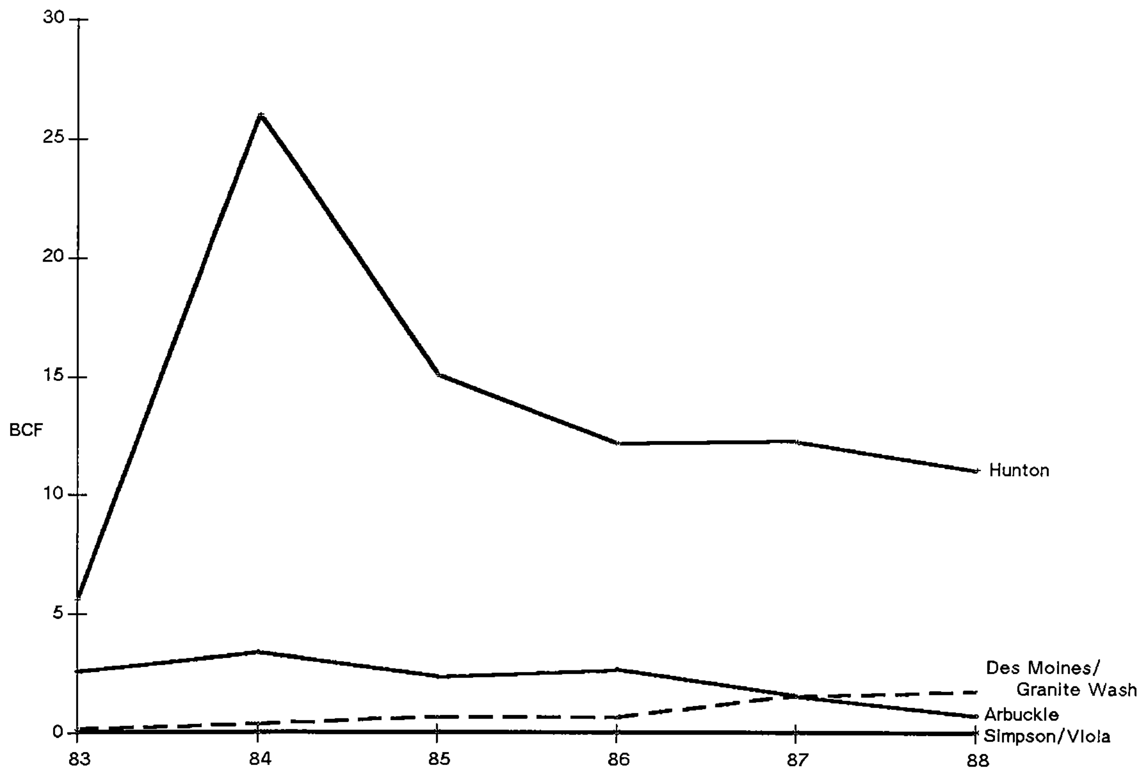


Figure 3. Mayfield West field: Natural-gas production by reservoir, 1983–88. BCF, billions of cubic feet.

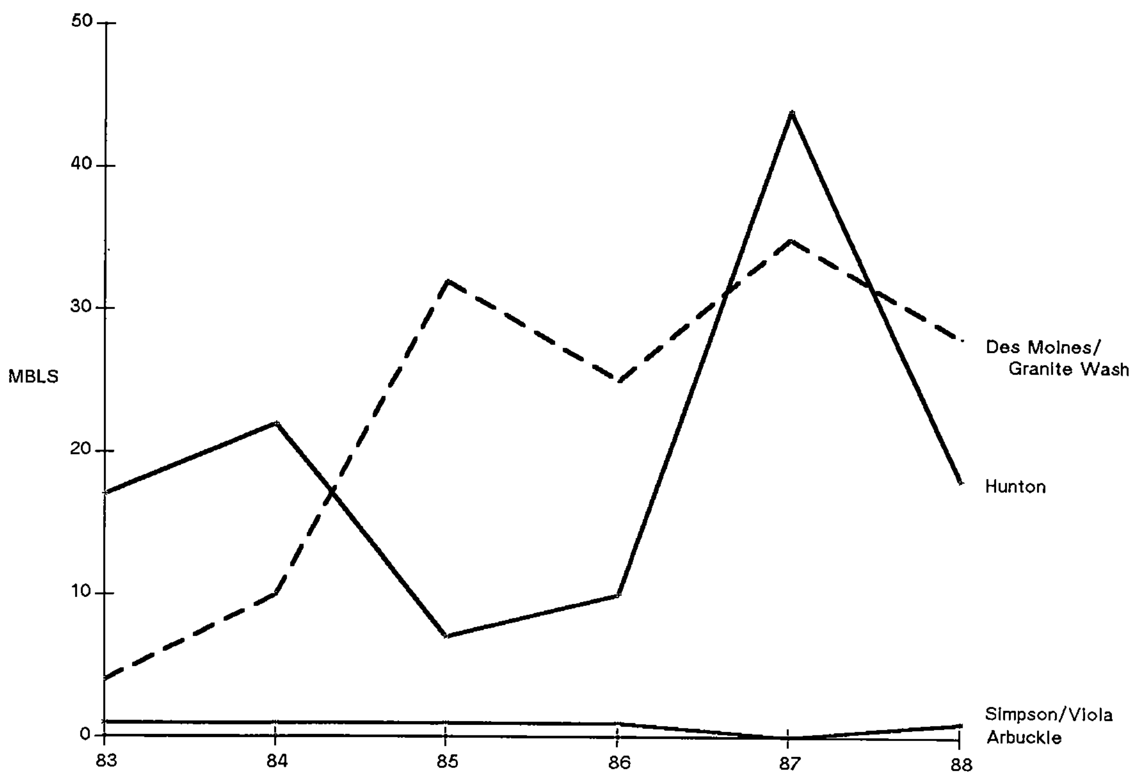


Figure 4. Mayfield West field: Petroleum-liquids production by reservoir, 1983–88. MBLS, thousands of barrels.

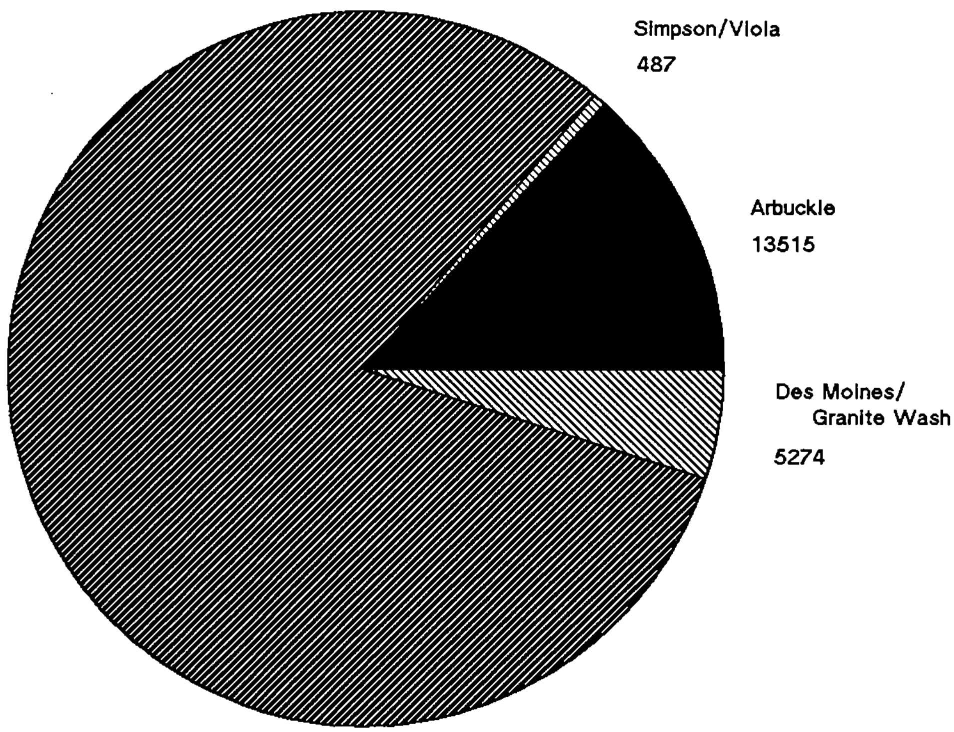


Figure 5. Mayfield West field: Relative proportions and amounts of cumulative natural-gas production by reservoir, 1983–88. MMCF, millions of cubic feet.

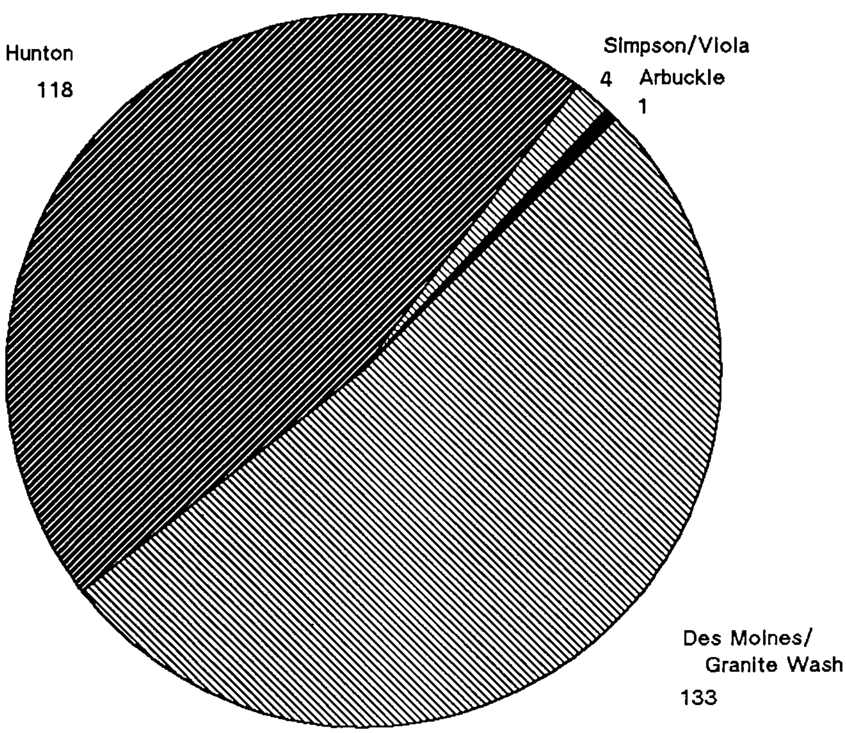


Figure 6. Mayfield West field: Relative proportions and amounts of cumulative petroleum-liquids production by reservoir, 1983–88. MBLS, thousands of barrels.

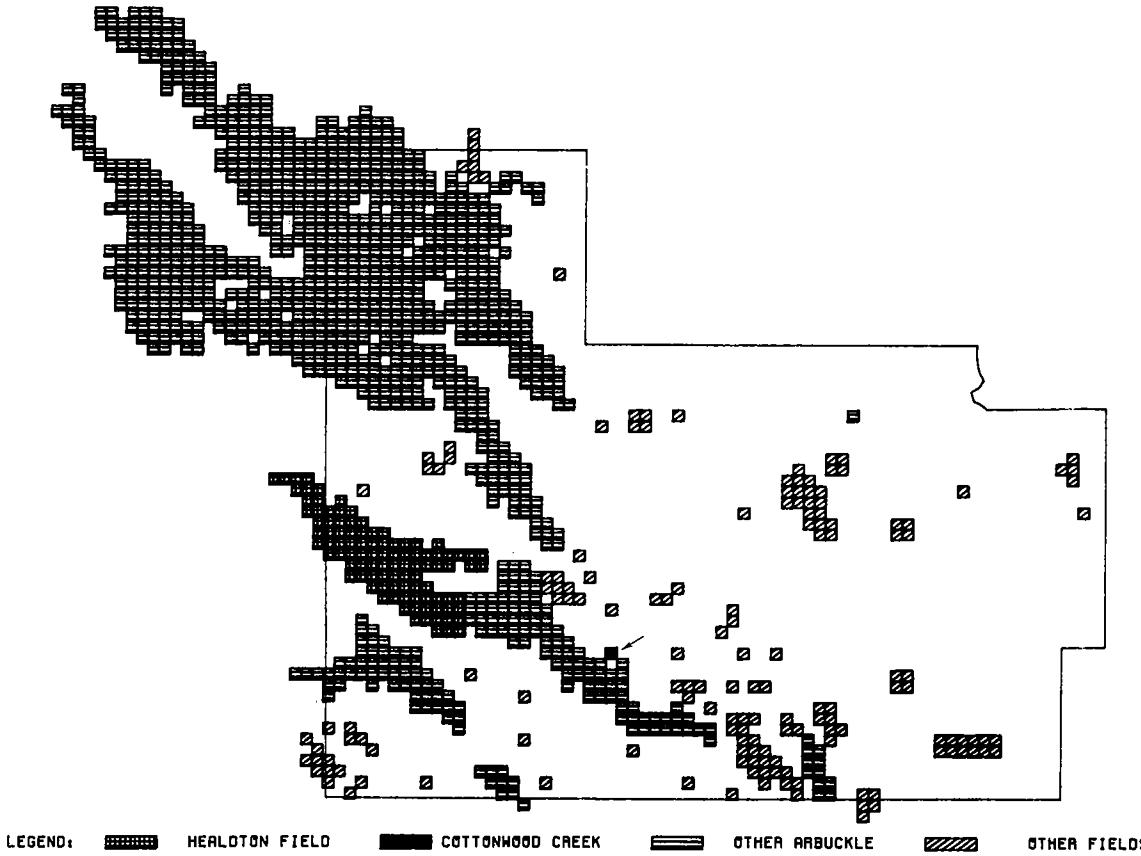


Figure 7. Oil and gas fields in Carter County and vicinity, Oklahoma. Each map unit is ¼ section. Arrow identifies Cottonwood Creek field. Source: NRIS data files.

Atoka and Spiro comingled, the Spiro sandstone and Wapanucka Limestone comingled, Spiro comingled with Morrowan and Atokan reservoirs, and the Arbuckle Group. Production of gas by reservoir is illustrated for the period 1983–88 (Fig. 15) and for 1988 (Fig. 16). Operators in the Wilburton area are departing from conventional wisdom of Arbuckle exploration by drilling 1,000 ft or more into the Arbuckle. The discovery well (ARCO Yourman No. 2) penetrates the entire Arbuckle section, 2,262 ft thick at that location. Prior to this, the norm in most of Oklahoma had been to test only the uppermost few hundred feet of the Arbuckle Group in exploratory wells. The deeper drilling of, and completions in the Arbuckle are shown graphically in Figure 17.

ACKNOWLEDGMENTS

The authors appreciate the assistance of members of the GIS staff in the preparation of data and computer graphics for this report: Joe Cusimano, Ann Gray, Kathy Hines, Anne Mycek-Memoli, and Terry Rizzuti.

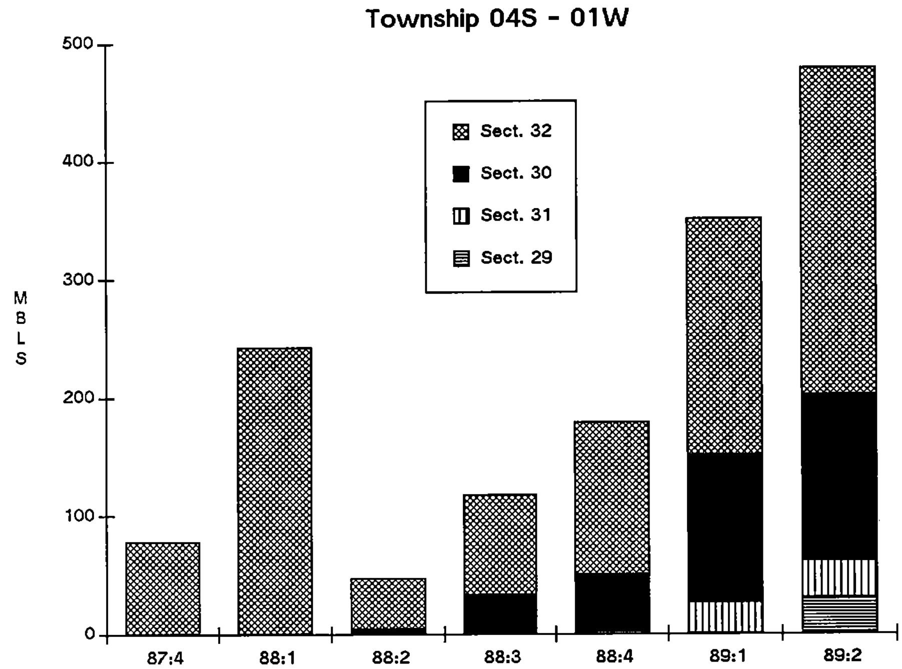


Figure 8. Cottonwood Creek field: Quarterly oil production, by section, from fourth quarter, 1987 (discovery), through second quarter, 1989. MBLS, thousands of barrels.

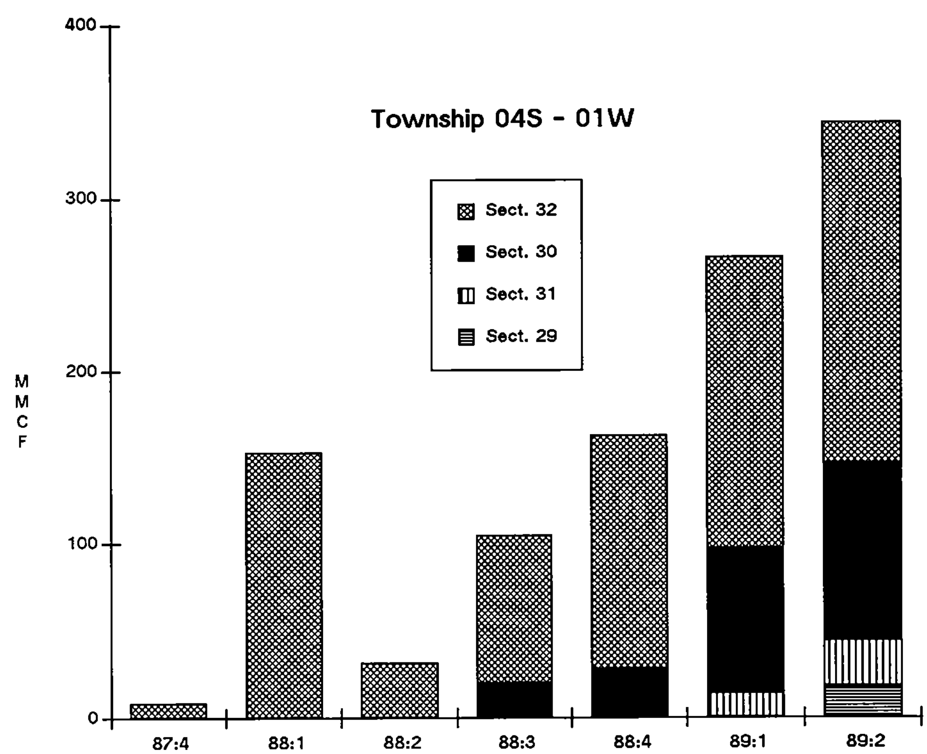


Figure 9. Cottonwood Creek field: Quarterly natural-gas production, by section, from fourth quarter, 1987 (discovery), through second quarter, 1989. MMCF, millions of cubic feet.

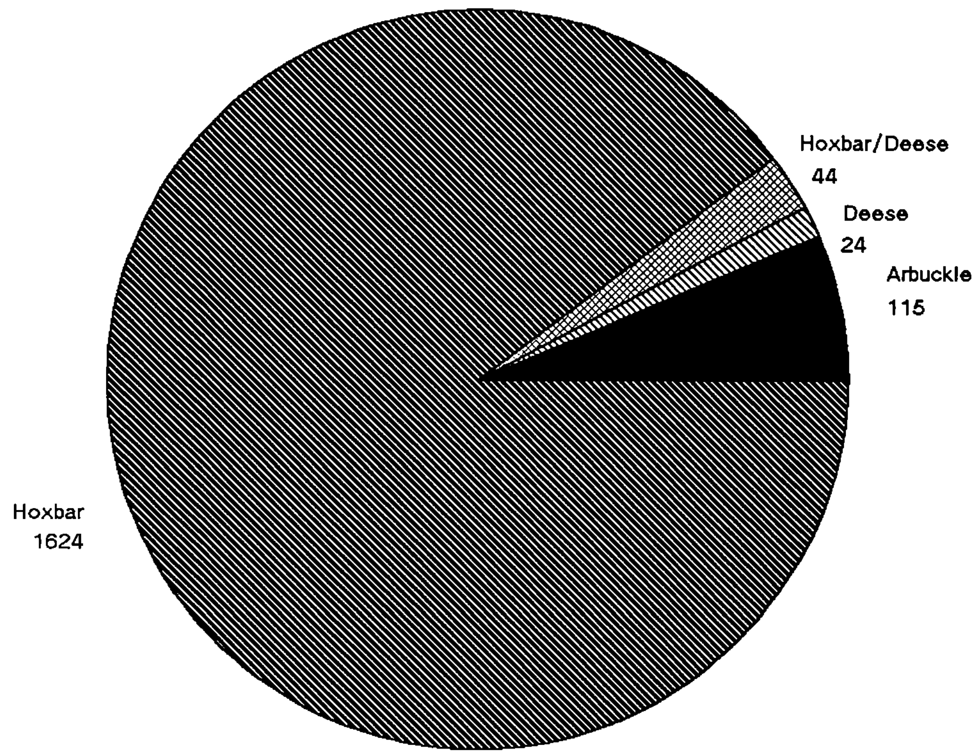


Figure 10. Healdton field: Relative proportions and amounts of petroleum-liquids production by reservoir in 1988. MBLS, thousands of barrels.

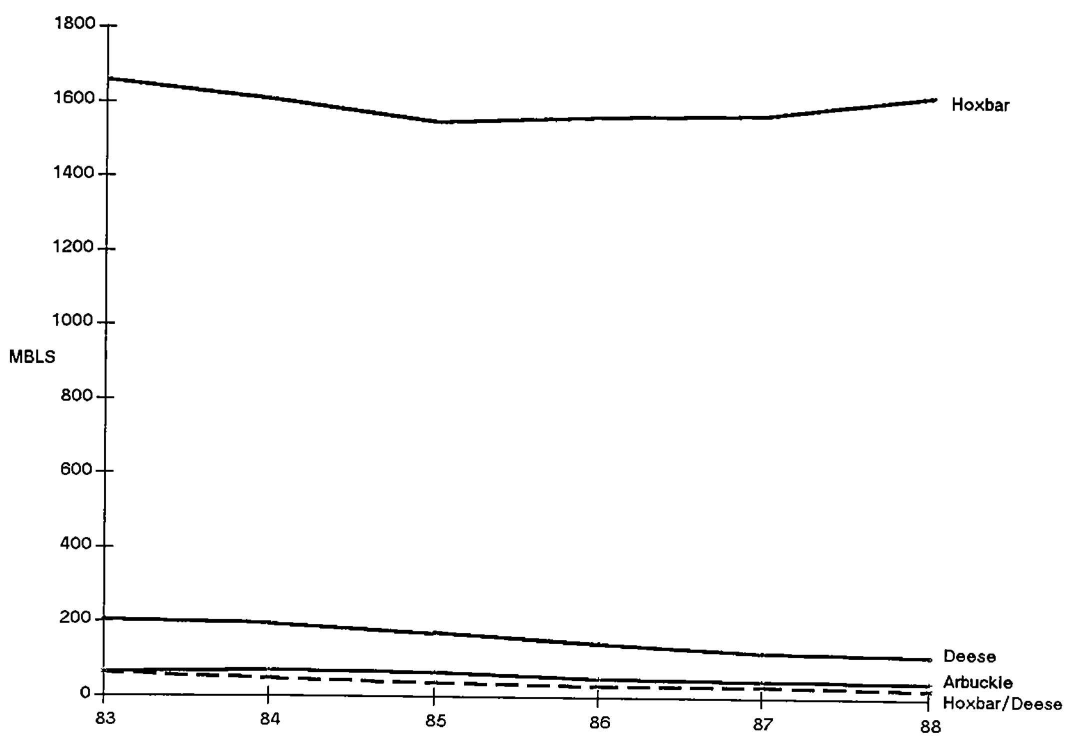


Figure 11. Healdton field: Petroleum-liquids production by reservoir, 1983–88. MBLS, thousands of barrels.

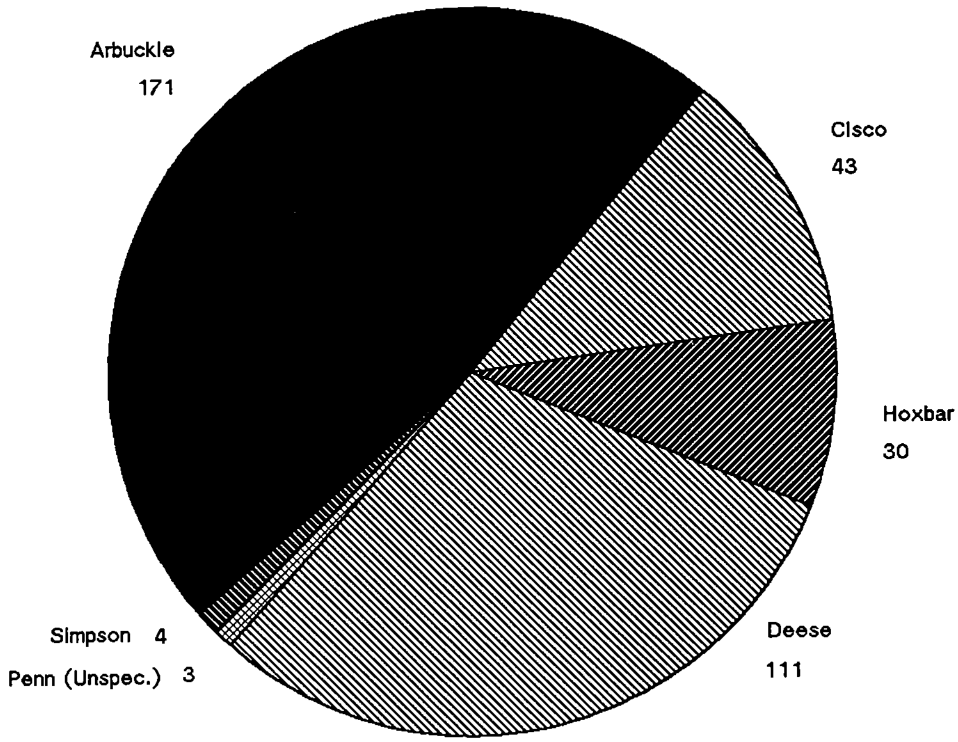


Figure 12. Healdton field: Relative proportions and amounts of natural-gas production by reservoir in 1988. MMCF, millions of cubic feet.

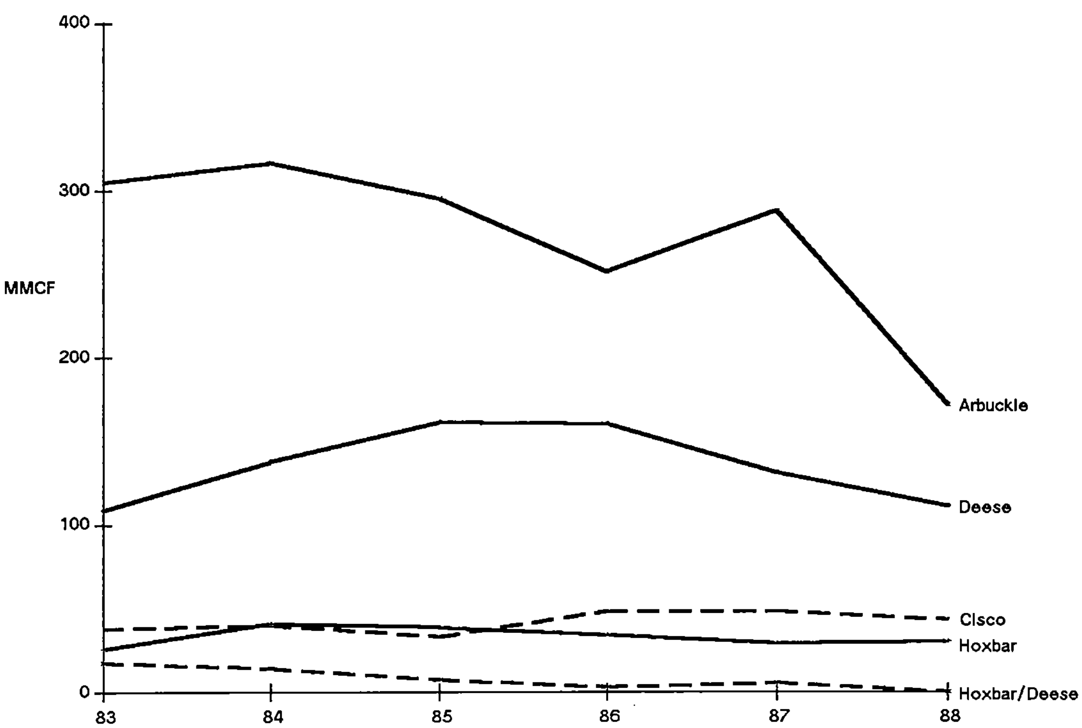


Figure 13. Healdton field: Natural-gas production by reservoir, 1983–88. MMCF, millions of cubic feet.

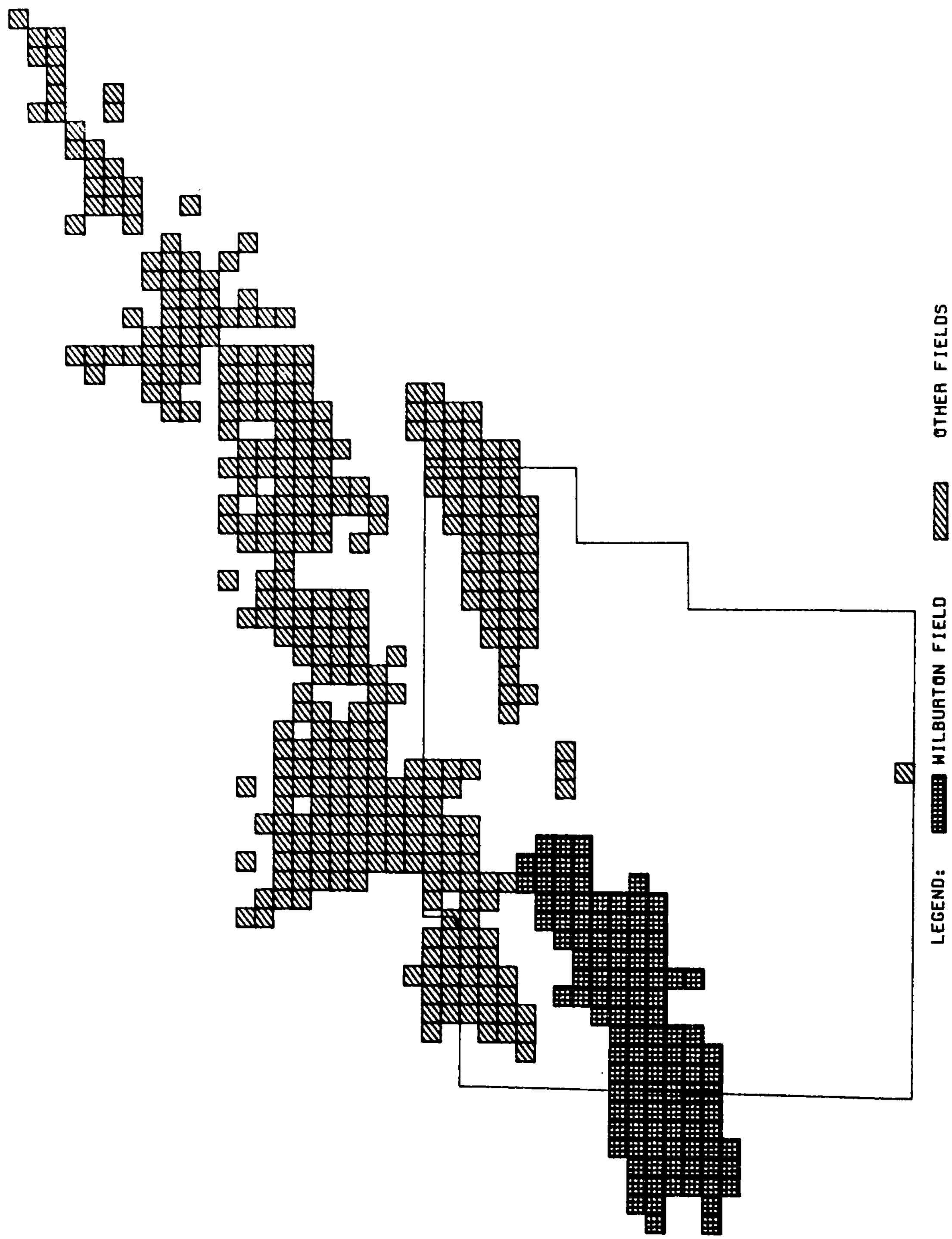


Figure 14. Gas fields in Latimer County and vicinity. Each map unit is one section. Source: NRIS data files.

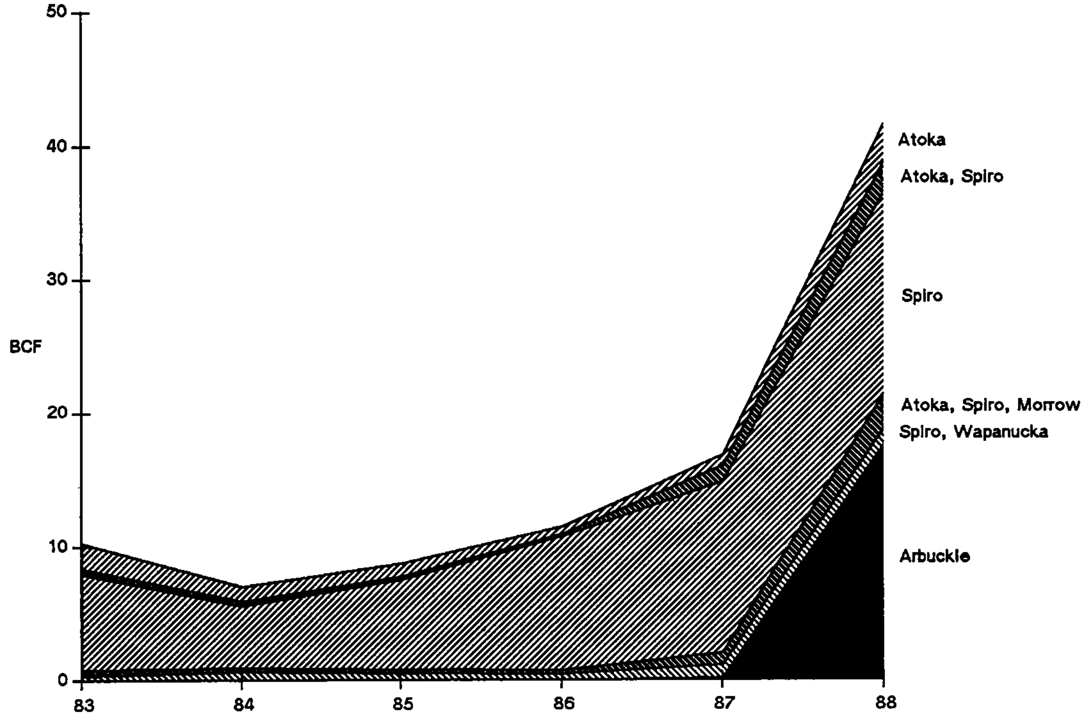


Figure 15. Wilburton field: Natural-gas production by reservoir, 1983–88 (Latimer County only). BCF, billions of cubic feet.

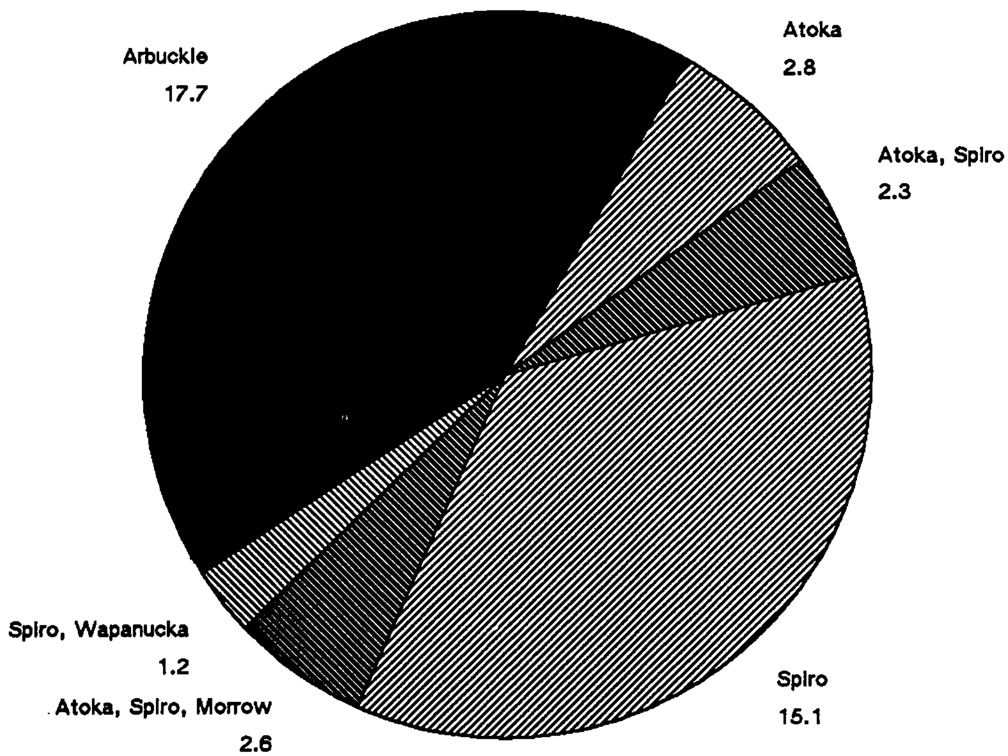
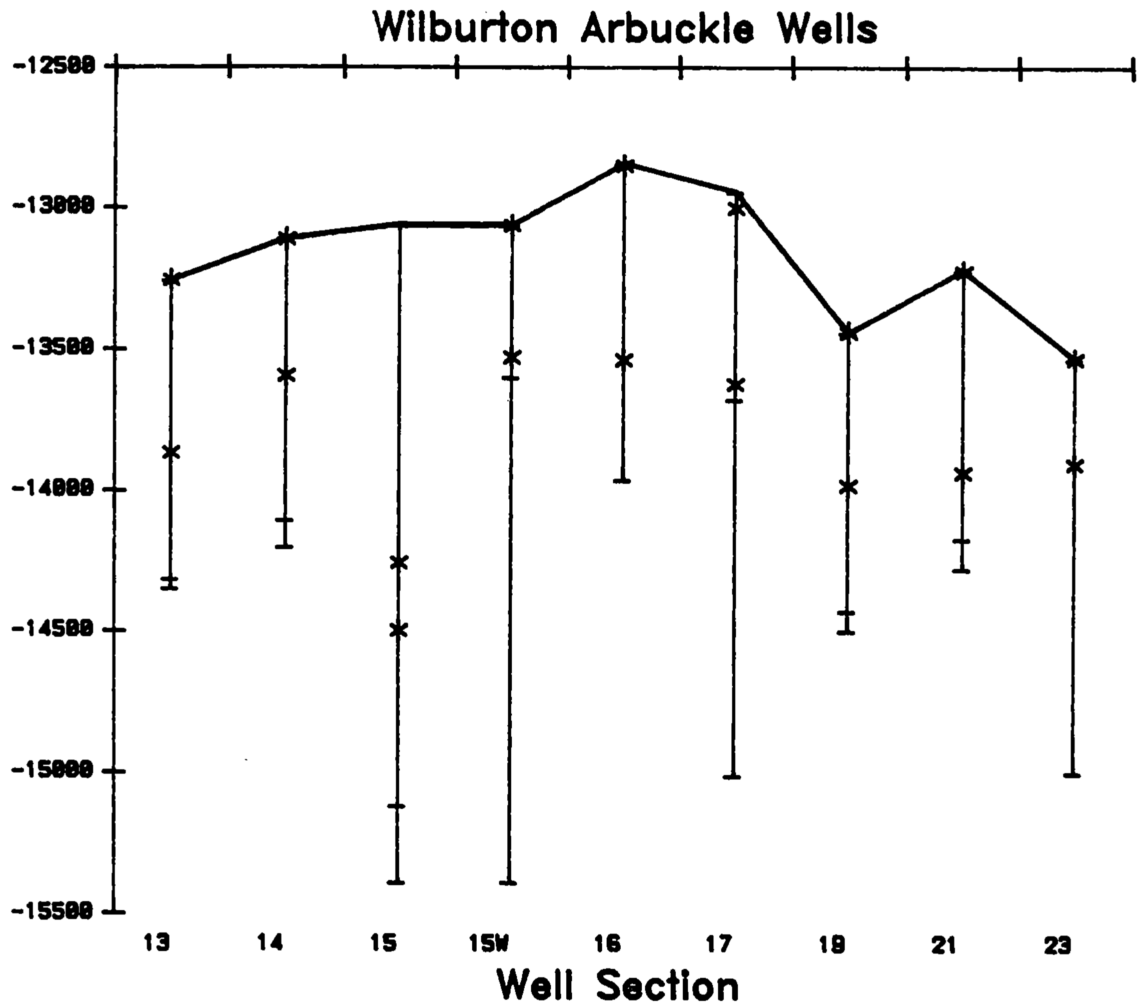


Figure 16. Wilburton field: Relative proportions and amounts of natural-gas production by reservoirs in 1988 (Latimer County only). BCF, billions of cubic feet.



Wilburton Arbuckle Wells
Township 05N-18E

Section	Lease Name	Well #	Completion Date
13	Austin	2	04/14/89
14	Costilow	3	06/28/88
15	Yourman	2	12/21/87
15(W)	Yourman (Workover)	2	06/05/88
16	Kilpatrick	2	06/23/88
17	Fazekas	2	04/21/88
19	Bennett State	2	12/17/88
21	Paschall	2	12/11/88
23	Williams A	3-23	04/05/89

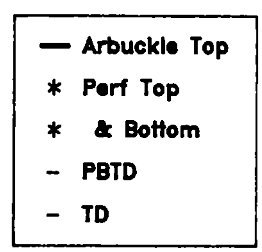


Figure 17. Wilburton field: Arbuckle reservoir completions: perforated interval, total depth, and plugged-back total depth.

GRAPTOLITE REFLECTANCE AS A POTENTIAL THERMAL-MATURATION INDICATOR

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INTRODUCTION

Graptolites (vernacular term for Class Graptolithina) are an extinct class of colonial marine invertebrates found in carbonate and clastic rocks of Cambrian to Pennsylvanian age (Bulman, 1970). Graptolites are an important index fossil for Lower Ordovician to Lower Devonian strata. The Class Graptolithina comprises six orders: Dendroidea, Tuboidea, Camaroidea, Stolonoidea, Crustoidea, and Graptoloidea. Of the two most common graptolite orders, Dendroidea and Graptoloidea, the Graptoloidea are more important for graptolite-reflectance studies in Ordovician to Lower Devonian strata because of thicker walls, whereas the Dendroidea are important for post-Silurian strata.

Beginning in 1976, organic petrologists identified graptolites as appearing "vitrinite-like" in reflected white light. Graptolite-reflectance analysis was adapted from vitrinite-reflectance analysis used on the woody tissue of vascular plants in post-Silurian strata. Graptolite reflectance is a measurement of the percentage of incident white light reflected from the polished surface of graptolite periderm (organic skeleton). Graptolite reflectance was assumed by the earliest workers to follow the same maturation trend as vitrinite reflectance. Subsequent work has shown that the graptolite maturation curve varies from the vitrinite maturation curve. Graptolite reflectance can presently be used as a qualitative measure of thermal maturation in interpretation of the thermal history of pre-Devonian strata. Future work will further correlate the graptolite-reflectance scale with the vitrinite-reflectance scale for use as a quantitative measure of thermal maturation.

The objectives of this report are (1) to introduce the potential use of graptolite reflectance as a thermal-maturation indicator for pre-Devonian strata; (2) to illustrate the characteristics of graptolites in reflected white light; and (3) to apply the graptolite-reflectance technique to three formations in southern Oklahoma.

PREVIOUS WORK

Graptolite-reflectance measurements have been applied to thermal-maturation studies in Australia, West Germany, Turkey, the British Isles, Poland, Sweden, and Canada. Kidwai (1986) and Malinconico (1989) have applied graptolite reflectance analysis to Ordovician strata in the United States.

Kurylowicz and others (1976) were the first to measure the maximum reflectance of coalified graptolite fragments in Ordovician strata to indicate the level of thermal maturation in the Amadeus basin, Australia. They noted that the bireflectance (maximum reflectance–minimum reflectance, %) and optical appearance of graptolites equated them more closely with the vitrinite-maceral group of coal than with either the liptinite- or inertinite-maceral groups. However, they assumed that graptolite reflectance follows the same maturation trend as vitrinite reflectance. Burne and Kantsler (1977) applied the graptolite-reflectance technique to Lower Ordovician strata of the Canning basin, Australia.

Teichmüller (1978) described the finely lamellar structure of graptolite cortical periderm in high-maturity Ordovician strata of West Germany. Clausen and Teichmüller (1982) discussed further the finely lamellar structure of the cortex and measured the maximum and minimum or random reflectance on graptolites of Germany and Sweden. Teichmüller and Teichmüller (1982) applied graptolite-reflectance measurements to level of maturation in West Germany.

Jackson and others (1984) measured maximum reflectance on graptolite zooclasts (denoted as vitrinite-reflectance equivalent) from Ordovician strata in the Amadeus basin, Australia. Gorter (1984) supplemented the graptolite-reflectance data of Kurylowicz and others (1976) for Ordovician strata of the Amadeus basin, Australia, and was the first to correlate graptolite reflectance to the conodont-color-alteration index (CAI). Gorter (1984) stated that graptolite and vitrinite reflectance values are nearly equivalent at a CAI of 1.5–2, whereas graptolite reflectance is higher than vitrinite reflectance at a CAI of ≥3.

Goodarzi (1984) described the dispersion of maximum and minimum reflectance, bireflectance, and morphology of graptolite fragments from Ordovician–Silurian strata of Turkey; therein were first described the two types of surface morphology of graptolites in reflected white light: granular and non-granular. Non-granular fragments were shown to have higher reflectance and higher bireflectance than granular fragments.

Goodarzi and others (1985) measured the maximum and minimum reflectance of graptolites, acritarchs, chitinozoans, and scolecodonts of Ordovician–Silurian samples from the Grand Banks of Newfoundland, noting the strong anisotropy of graptolite fragments.

Goodarzi and Norford (1985) described several optical properties of graptolites (maximum and minimum reflectance, bireflectance, refractive index, absorptive index) from Ordovician–Silurian strata of Canada. Maximum graptolite reflectance was related to maximum vitrinite reflectance from CAI values. Graptolites in shaly matrices were found mostly to have non-granular surface morphology, while those in carbonate matrices mostly have granular texture. Studied graptolites from Orders Graptoloidea and Dendroidea were identified to the genus level.

Goodarzi (1985) measured the dispersion of optical properties (maximum reflectance, refractive index, absorptive index) of graptolite periderm in the visible spectrum (450–650 nm) from Ordovician–Silurian strata of Canada. The spectral trends of reflectance, refractive index, and absorptive index of graptolite periderm were reported to vary with increasing maturity (increasing aromaticity), similar to vitrinite in coal. Granular surface morphology of graptolites was present in low-maturity periderm in a carbonate matrix, and non-granular surface morphology was present in mature to post-mature periderm in a shale matrix.

Kemp and others (1985), Oliver (1988), and Parnell (1989) measured the maximum graptolite reflectance in Ordovician–Silurian strata of the British Isles to assess low-grade metamorphism.

Bertrand and Héroux (1987) measured the random reflectance of chitinozoans, graptolites, and scolecodonts in Ordovician–Silurian strata of Anticosti Island, Canada. The graptolite maturation curve was reported to be subparallel to and above the vitrinite maturation curve (higher reflectance).

Goodarzi and others (1988) determined thermal-maturation level from optical properties of graptolites from core samples of Ordovician strata in Sweden. They concluded that reflectance measurements from graptolite specimens associated with carbonate or pyrite should not be used for maturation assessment.

Goodarzi and Norford (1989) reported the maximum and minimum reflectance and bireflectance of graptolites from Upper Cambrian to Upper Silurian shales of Poland, Sweden, and Canada. Maximum graptolite reflectance was reported to be higher than random vitrinite reflectance at the equivalent level of thermal maturity based on the CAI.

Bustin and others (1989) measured optical properties (maximum, minimum, and random reflectance and bireflectance) and chemical properties (Rock-Eval pyrolysis and infrared spectroscopy) of graptolite periderm and vitrinite macerals following laboratory-simulated maturation. Maximum and random graptolite reflectance were directly correlated with vitrinite of the same maturation level.

DISTINGUISHING CHARACTERISTICS OF GRAPTOLITES

Photomicrographs illustrating characteristic features of graptolite periderm in reflected white light are found in Kurylowicz and others (1976), Teichmüller (1978), Clausen and Teichmüller (1982), Stach and others (1982), Goodarzi (1984, 1985), Goodarzi and Norford (1985,1989), Goodarzi and others (1988), and Bustin and others (1989).

Distinguishing characteristics of graptolites (reflected white light, 200× to 500× magnification) are (1) finely lamellar structure of cortical and fusellar periderm, with alternating bright and dark lamellae, often resembling a fingerprint pattern (Fig. 1); (2) lobate shape of cortical and fusellar periderm (Fig. 2); (3) thin, elongate walls (Fig. 3); (4) two parallel walls separated by a secondarily filled space (Fig. 4); and (5) granular and non-granular surface texture.

APPLICATION TO OKLAHOMA

It is well documented that Oklahoma contains graptolite-rich rocks. Decker (1959; see bibliography by Huffman, 1958) published more than 20 articles concerning graptolite-bearing strata of Oklahoma. Figure 5 summarizes the Ordovician–Silurian graptolite-bearing strata of Oklahoma.

Graptolite reflectance was measured on samples from eight locations in southern Oklahoma (Fig. 6, Table 1): six samples of the Viola Springs Formation (Upper Ordovician), one sample of the Bigfork Chert (Upper Ordovician), and one sample of the Polk Creek Shale (Upper Ordovician).

One core and six outcrop samples of graptolite-bearing rocks from southern Oklahoma were obtained from the collections of S. C. Finney; one outcrop sample of the Polk Creek Shale was collected by B. J. Cardott (Table 1).

Figure 1. Photomicrograph showing finely lamellar structure of cortical and fusellar periderm (Viola Springs Formation, sample 2; reflected white light, oil immersion, 200×, diameter of field 140 μm).

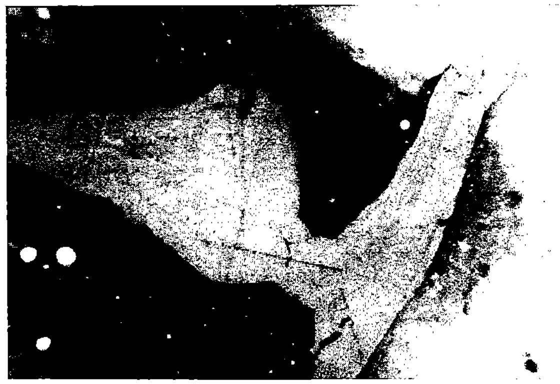

Figure 2. Photomicrograph showing lobate shape of cortical and fusellar periderm. The right edge of the far right side of the specimen is the outer wall of the rhabdosome (Viola Springs Formation, sample 2; reflected white light, oil immersion, 80×, diameter of field 320 μm).

Figure 3. Photomicrograph showing thin, elongate walls of the graptolite periderm; the spaces between the walls housed the zooids (Athens Shale, Middle Ordovician, Alabama; reflected white light, oil immersion, 200×, diameter of field 140 μm).

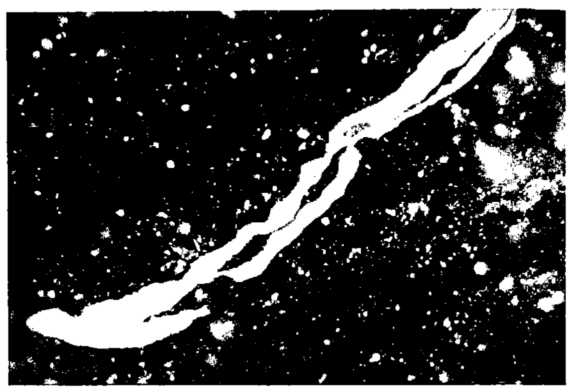

Figure 4. Photomicrograph showing two parallel walls separated by a secondarily filled space that once housed a zooid. The specimen has suffered postdepositional flattening (Viola Springs Formation, sample 2; reflected white light, oil immersion, 80×, diameter of field 320 μm).

Several methods of sample preparation were evaluated: polished thin section; polished pellets of isolated graptolite specimens (isolated by acid digestion of rock matrix); and polished whole-rock pellets. The primary problems with polished thin sections and isolated graptolite pellets were limited control of specimen orientation and plucking during polishing of the delicate, brittle, carbonized graptolite specimens. Polished whole-rock pellets were preferred because of ease of preparation and limited plucking and orientation problems. Both oriented (graptolite specimens oriented parallel to bedding) and non-oriented whole-rock pellets were used. Kerogen-concentrate pellets were used for the Polk Creek Shale sample. Pellets were polished to a scratch-free, relief-free surface.

Graptolite reflectance was measured following the standard procedures of vitrinite-reflectance analysis (ASTM, 1988, D2798): 500× magnification; plane-polarized, monochromatic green light (546 nm); immersion oil $N_e = 1.5180$. Mean graptolite reflectance values (Table 1) are based on 20 to 60 measurements. Mean random graptolite reflectance (plane-polarized light, stationary stage) was determined on samples with small graptolite fragments. CAI values were determined by W. C. Sweet and S. M. Bergström, and reported to S. C. Finney (personal communication, 1985).

Figure 6 and Table 1 summarize the location and petrographic data of the Viola Springs, Bigfork Chert, and Polk Creek Shale samples from southern Oklahoma. Both the graptolite-reflectance and

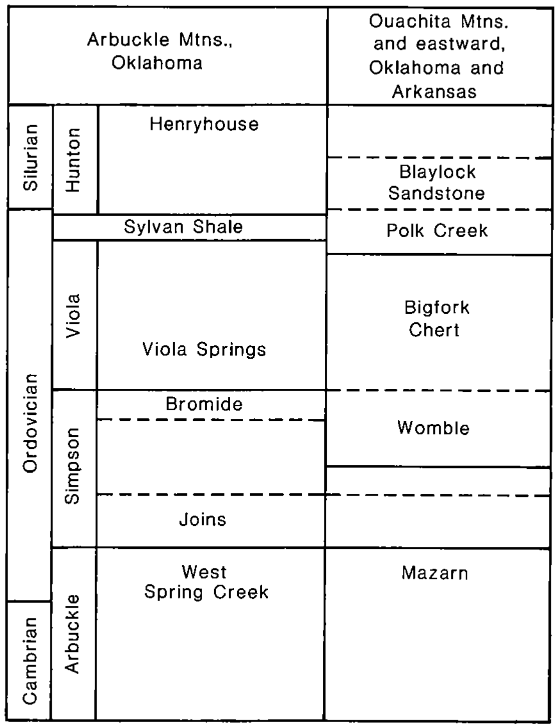

Figure 5. Graptolite-bearing strata of Oklahoma. Adapted from Decker (1936a,b;1959) and Ethington and others (1989).

CAI data indicate that the samples are of low thermal maturity.

Mean maximum graptolite-reflectance values of the Viola Springs in the Arbuckle Mountains are 0.36–0.69% (samples 1–3), which compares with mean random vitrinite reflectance of the Woodford Shale (Upper Devonian–Lower Mississippian) in the Arbuckle Mountains of 0.35% to 0.77% (Cardott and others, 1990). The 0.39% mean maximum graptolite reflectance of the Viola Springs in the Criner Hills (sample 5) compares with the 0.43% mean random vitrinite reflectance of the Woodford Shale in the Criner Hills (sec. 36, T. 5 S., R. 1 E.) (Cardott, unpublished data). Mean maximum graptolite reflectance of the Viola Springs in the northwest extension of the Marietta basin (sample 6), at a depth of 5,628 ft, is 0.81%. The 0.66% mean random graptolite reflectance of the Polk Creek Shale (sample 7) compares with the 0.47% mean random vitrinite reflectance of the Arkansas Novaculite (Devonian–Lower Mississippian; sec. 13, T. 2 S., R. 11 E.) (Cardott, unpublished data). Mean random graptolite reflectance of the Bigfork Chert in the Ouachita Mountains (sample 8) is 0.82%. The 0.34% mean random bitumen reflectance for sample 8, based on 50 measurements, correlates with an equivalent vitrinite reflectance value of 0.61% (Jacob, 1985).

Direct correlation of graptolite reflectance with another thermal-maturation indicator has not yet

TABLE 1.—SAMPLE INFORMATION AND PETROGRAPHIC DATA

Sample no.	Formation	Location	Sample type	Lithology	Mean maximum R_o(%)	Mean minimum R_o(%)	Graptolite texture[a]	CAI[b]
1	Viola Springs	22–2S–1W	outcrop	limestone	0.36	0.32	G	1
2	Viola Springs	25–2S–1E	outcrop	limestone	0.69	0.65	G	1
3	Viola Springs	28–1S–2E	outcrop	limestone	0.43	0.38	G	1
4	Viola Springs	12–2S–6E	outcrop	limestone	0.49[c]	NA	G	1
5	Viola Springs	35–5S–1E	outcrop	limestone	0.39	0.33	G	1
6	Viola Springs	20–4S–5W (8–20 Dillard, depth of 5,628′)	core	limestone	0.81	0.66	G	1.5
7	Polk Creek	14–2S–11E	outcrop	shale	0.66[c]	NA	NG	NA
8	Bigfork Chert	9–1S–12E	outcrop	limestone	0.82[c]	NA	NG	1

[a]G = granular; NG = nongranular.
[b]Conodont-color-alteration index.
[c]Mean random graptolite reflectance (plane-polarized light, stationary stage).

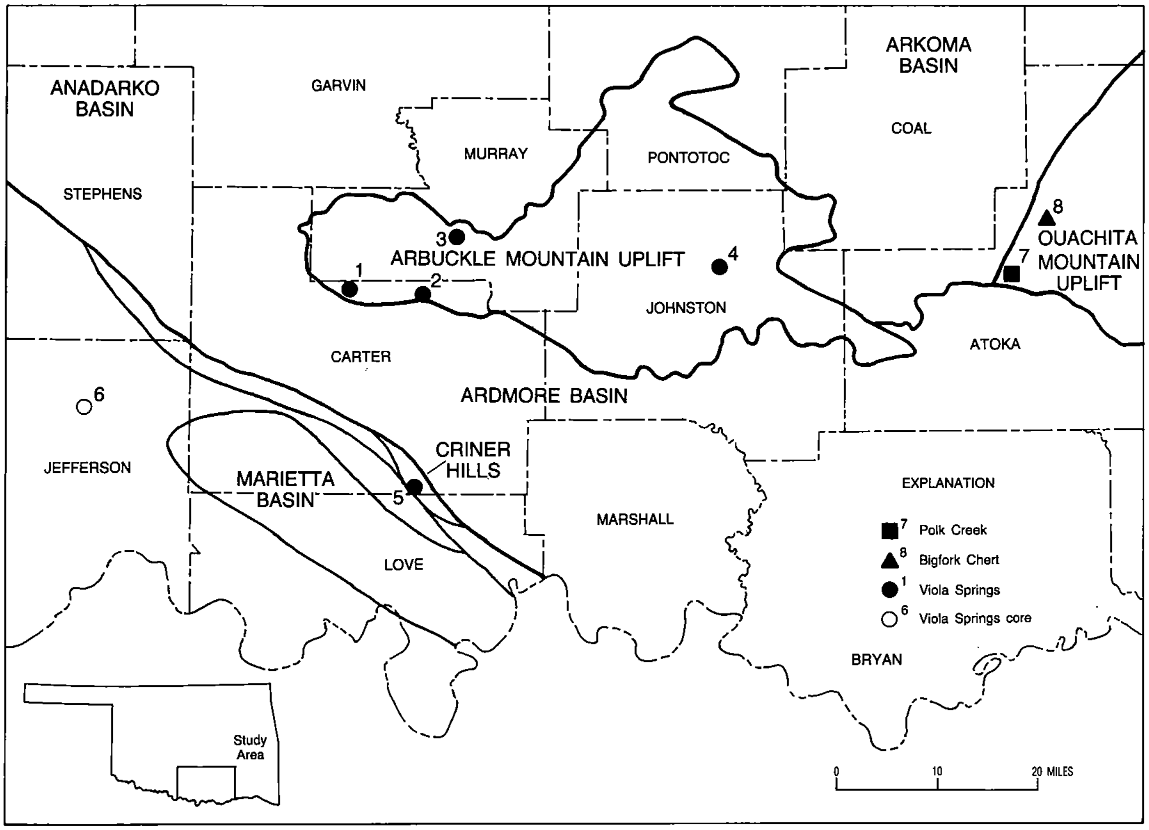

Figure 6. Major geologic provinces of southern Oklahoma, showing sample locations of graptolite-bearing strata. Refer to Table 1 for sample locations.

been achieved. Therefore, the graptolite-reflectance data are compared with other thermal-maturity data and interpreted qualitatively. Delineation of the graptolite maturation curve has been complicated by measurements of both maximum and random reflectance on granular and nongranular graptolites, as used in this study.

Goodarzi and Norford (1989) emphasized the importance of measuring maximum graptolite reflectance, due to the biaxial optical character of the graptolite periderm, exhibiting anisotropy. Anisotropy is measured as bireflectance (maximum R_0–minimum R_0, %). Bireflectance increases with increasing maturity, the greatest bireflectance being observed on graptolites sectioned perpendicular to bedding; development of bireflectance in graptolites starts at lower maximum reflectance than in vitrinite (Goodarzi and Norford, 1985,1989).

Maximum reflectance is usually higher than random reflectance, the spread between the values becoming greater with increasing maturity. The practical application of measuring random graptolite reflectance from dispersed graptolite fragments in shales, as used for samples 7 and 8, suggests that random graptolite reflectance in shales, rather than maximum reflectance, should be correlated with random vitrinite reflectance, as applied by Bertrand and Héroux (1987).

Goodarzi (1984) and Goodarzi and Norford (1985) indicated that reflectance is lower for graptolites with granular surface texture, found mostly in carbonate matrices, than for graptolites with nongranular morphology, found mostly in shaly matrices. Goodarzi and Norford (1989) and Bustin and others (1989) restricted their studies to graptolites from shales.

Measurement of random graptolite reflectance in three samples and the occurrence of granular-textured graptolites in most of the samples (Table 1) may have resulted in lower reflectance values than if maximum reflectance were measured from nongranular specimens. However, the graptolite reflectance would not have changed significantly, because of the low maturity and low anisotropy (bireflectance <0.06% for surface samples, Table 1) of these samples.

In this study, the lower graptolite-reflectance values may be similar to vitrinite reflectance values. Gorter (1984) stated that maximum graptolite reflectance was roughly equivalent to maximum vitrinite reflectance at reflectances less than

~1.3%. Bertrand and Héroux (1987) and Goodarzi and Norford (1989) indicated that graptolite reflectance is higher than vitrinite reflectance at the equivalent level of thermal maturity. Goodarzi and Norford (1989) further stated that graptolites found in a carbonate matrix have a reflectance similar to vitrinite reflectance, whereas graptolites preserved in a shale matrix have a higher reflectance than vitrinite.

Comparison of graptolite reflectance of Ordovician samples with vitrinite reflectance of Devonian–Lower Mississippian samples from southern Oklahoma, both of low thermal maturity, indicates that graptolite reflectance is a very good qualitative indicator of thermal maturity.

ACKNOWLEDGMENTS

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Oklahoma Geological Survey Circular 92, 1991

STUDY OF THE ORDOVICIAN ARBUCKLE–SIMPSON AQUIFER IN THE SULPHUR AREA OF SOUTH-CENTRAL OKLAHOMA

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ABSTRACT.—Because of concern about decreasing flow from springs near Sulphur, Oklahoma, in the Travertine District of the Chickasaw National Recreation Area (CNRA), a hydrogeologic study was conducted. In the CNRA, Ordovician Arbuckle Group and Simpson Group rocks are the primary aquifers and generally are overlain by a surficial cover of Pennsylvanian-aged conglomerate. Because the hydrogeology of the area is influenced largely by geologic structure, a major part of the study entailed determination of subsurface structure of the Arbuckle–Simpson aquifer. Preliminary geologic sections were constructed based on available surface-geology maps, reconnaissance of surface geology, and available borehole geologic and geophysical data. These sections were refined by analysis of gravity and magnetic data obtained for 180 data stations in a 125-km^2 area that included the CNRA. These data were analyzed using a Fast Fourier Transform computer program. Computations included second vertical derivatives and incremental downward continuation of the gravity and magnetic fields.

Utilizing geologic data from various sources, density model inversion of gravity data was computed using a least-squares matrix-inversion program. For areas of limited geologic data, geologically feasible models of the structural geometry were developed from north-south Bouguer gravity profiles.

Interpretation and synthesis of these data sets resulted in the interpolation of fault locations beneath the Pennsylvanian-age conglomerate. The structural geometry of the study area appears to be consistent with that of a left-lateral wrench-fault model. Faults with geometries best explained as en echelon extension features, transtensional normal faults, and antithetic faults are present in the study area. The subsurface structure below the Travertine District of CNRA is one of convergence and termination of several faults.

STRATIGRAPHY AND FACIES IN THE MIDDLE ORDOVICIAN SIMPSON GROUP, SOUTHWESTERN KANSAS AND WESTERN OKLAHOMA

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ABSTRACT.—The Middle Ordovician Simpson Group in southwestern Kansas and western Oklahoma consists of a complex series of sandstones, green sandy shales, and sandy carbonate rocks. Within this area, the Simpson lies entirely in the subsurface. Borehole logs, interpreted sample logs, and cores were used to study the lithofacies and their relationships to the conventional stratigraphic nomenclature established from outcrops in the Arbuckle Mountains.

In south-central Kansas, in Kiowa County and adjoining counties, the Simpson can be subdivided into six informal stratigraphic units. The lowermost two, a sandstone and a green sandy shale, correlate with the McLish Formation. The upper four—a sandstone; a sandy, cherty dolomite; a green, sandy shale; and a thin, sandy limestone—correlate with the Bromide Formation.

To the south, in western Oklahoma, the Simpson becomes thicker and more of its formations can be distinguished as the southern Oklahoma aulacogen is approached. Where fully developed, the Simpson is divided, in ascending order, into the Joins, Oil Creek, McLish, Tulip Creek, and Bromide Formations. The Oil Creek Formation extends north approximately to the Kansas state line, but the Joins and Tulip Creek Formations are found only farther south.

A notable feature within the Bromide Formation is a well-defined area of carbonate rocks ~50 mi across east-west and 150 mi across north-south, half of which is in Kansas and half in Oklahoma. This carbonate facies is primarily a sandy, cherty dolomite averaging ~35 ft thick. The boundaries of this facies are very sharp, and its thickness can increase from 0 to 25 ft within a mile.

In southwestern and west-central Kansas, the Simpson is considerably thinner, and the informal six-unit stratigraphy cannot be correlated into this area. Toward the northwest, the Simpson tends to be reduced to a thin, sandy dolomite.

STRATIGRAPHIC AND DEPOSITIONAL FRAMEWORK, SUPRA-EVERTON MIDDLE AND UPPER ORDOVICIAN STRATA, NORTHERN ARKANSAS

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INTRODUCTION

The current stratigraphic nomenclature (Fig. 1) of supra-Everton Ordovician units in northern Arkansas was established by Miser (1922). These strata crop out in a narrow east-west belt from the Mississippi embayment near Batesville, Independence County, on the east, to Ponca, Newton County, on the west. Between Ponca and eastern Oklahoma, supra-Everton rocks are covered by Carboniferous strata.

The cumulative thickness of these rocks is slightly in excess of 200 m; however, in no one locality is this maximum attained. The sequence is thickest near Batesville in the east, but it thins markedly westward through the thinning and disappearance of individual units. In the westernmost part of the outcrop belt, the entire supra-Everton succession is composed of an unevenly distributed Fernvale Limestone that ranges in thickness from 0 to 7.5 m, plus a few scattered exposures of thin Plattin Limestone and Cason Shale. Details on the petrology and stratigraphy of the succession are found in Craig (1975a,b) and Craig and others (1988).

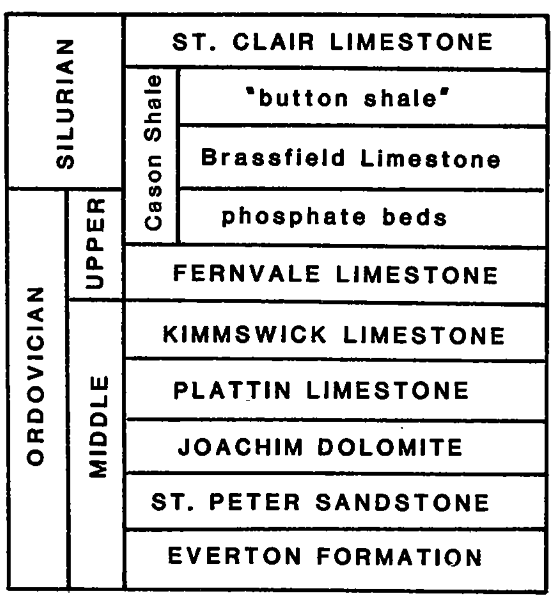

Figure 1. Stratigraphic column, supra-Everton Ordovician strata, northern Arkansas.

DEPOSITIONAL FRAMEWORK

Supra-Everton deposition in northern Arkansas began with transgression of a strandline sand, now the St. Peter Sandstone, over the eroded surface of the Everton Formation. Directly following transgression, deposition of the Joachim Dolomite was initiated, rapidly aggrading near-shore parts of the sandy shelf into a broad tidal flat. The passage from the St. Peter into the Joachim is gradational. The basal Joachim is composed of burrow-mottled, sandy, fossilferous dolomudstone of shallow subtidal to low intertidal origin. The middle and upper Joachim are characterized by hemicycles composed of a basal sandy intraclastic dolograinstone; a middle, faintly laminated dolomudstone; and an upper, prominently mud-cracked, laminated dolomudstone that contains abundant laminoid and irregular fenestrae (birdseye structure), calcite pseudomorphs after halite, and in places distinct stromatolitic heads. The bulk of the Joachim is interpreted as the product of the intertidal and supratidal zones. The hemicycles are believed to represent shallowing-upward sequences that start with grainstone of the low intertidal zone and terminate with mud-cracked, intraclastic layers of the supratidal zone.

In contrast to the thick-bedded dolostone of the Joachim, the Plattin is dominantly slabby-bedded limestone with only subordinate dolostone. The most common Plattin lithic type is composed of millimeter-scale laminations of lime mudstone alternating with peloidal packstone/grainstone. This rock possesses abundant mud cracks, fenestral fabric, and calcite pseudomorphs after gypsum and halite. Evaporites grew as porphyroblasts in Plattin sediment, in places becoming laterally linked to form irregular fenestrae. All these features unequivocally identify the bulk of the Plattin as a product of the supratidal environment. The base of the Plattin in the eastern part of the outcrop belt contains layers of mollusk wackestone of low faunal diversity, suggesting subtidal influence. Skeletal wackestone/packstone possessing a relatively diverse fauna characterized by *Tetradium* (tabulate coral) and *Hedstroemia* (calcareous alga) occurs near the top of the Plattin, marking a distinct subtidal incursion over the Plattin tidal flat. Above the *Tetradium* limestone, the youngest Plattin preserved records renewed tidal-flat progradation.

The Kimmswick Limestone is composed of interbedded bioturbated skeletal wackestone, packstone, and poorly washed grainstone. It contains the most diversified fauna of all the supra-Everton Ordovician units and is interpreted as the product of a low-energy, subtidal environment. Protection was probably afforded by shoals of crinozoan sands south of the outcrop belt, as suggested by subordinate crinozoan grainstone interbedded with more common lower-energy wackestones and packstones.

The Fernvale Limestone is thick-bedded, commonly cross-bedded, coarse-grained crinozoan grainstone interpreted to have accumulated in a high-energy, open, subtidal environment as shifting subaqueous dunes.

The final phase of Ordovician sedimentation in northern Arkansas is assigned to the Cason Shale, a unit with a complicated depositional and nomenclatural history. Only the basal beds of the Cason are Ordovician; the remainder belong to the Lower and Middle Silurian and will not be discussed here. Details of the Cason nomenclatural problem are given in Craig (1975a,1984). In eastern outcrops, the Ordovician part of the Cason consists of phosphatic sandstone and shale overlain by crinozoan grainstone that grades upward into sporadically occurring oolitic–intraclastic grainstone and fenestral lime mudstone, the record of which has been mostly removed by post-Ordovician erosion. The phosphate grains, consisting of crinozoan parts, rock fragments, and internal molds of clams and snails, are almost certainly reworked from below and represent a transgressive lag deposit that was distributed unevenly on the underlying erosional surface.

To the west, the phosphate content of the Ordovician Cason decreases significantly, and the phosphatic sandstone and shale are replaced by greenish, dolomitic, silty shale that resembles the Sylvan Shale of Oklahoma.

STRATIGRAPHIC FRAMEWORK

Although there is relatively little disagreement on the interpretation of environments of deposition of these supra-Everton units, geologists have been divided on the historical meaning of some of the formational contacts. Most agree that the St. Peter rests unconformably on lower strata and is conformably overlain by the Joachim. Most are in agreement also that the lower Cason is unconformable with the underlying Fernvale. Miser (1922) placed unconformities between the carbonate units, even though their planes of separation appeared "even" to him. Some later geologists have interpreted the deepening-upward carbonate succession and smooth contacts as a result of landward migration of adjacent, coexisting lithotopes through a period of time that involved no significant breaks in deposition (Young and others, 1972). However, evidence now available strongly suggests that Miser's original interpretation is correct, and that periods of subaerial exposure and erosion separate these carbonate units.

Contacts between the carbonates are "welded"; that is, they occur within a single bed and can be collected. They are universally sharp, show small-scale truncation features, and can show small-scale scalloping (relief measured in millimeters) of the upper surface of the underlying unit. In addition, patches of the overlying rock commonly occur within the underlying unit up to several centimeters below the contact. Freeman (1966) applied the term "microkarst" to these features.

Slight angular relationships occur between the Joachim and Plattin and between the Plattin and Kimmswick; where these occur, the removal of several meters of rock can be demonstrated.

Thickness distributions also suggest that the units were separated by periods of erosion. Significant thickness variations, which bear no apparent relation to facies patterns, occur over short distances. On a regional scale, all units thin to the west, and several eventually disappear from the sequence. The Joachim and Kimmswick do not occur west of western Stone County, where the Plattin is in sharp contact with the St. Peter below and the Fernvale above. In the westernmost part of the outcrop belt, the entire interval is represented by an unevenly distributed Fernvale Limestone, which in most exposures rests directly on the Everton Formation, and a few scattered, thin occurrences of Plattin and Cason. Where units are thin, their occurrence is sporadic; that is, occurrences of a few centimeters to a few meters can be present in some sections, and the unit may be altogether absent in other sections. Such occurrences are best explained as erosional remnants. To interpret this westward thinning of formations as depositional pinchout seems unreasonable in light of the distribution pattern at their distal edges.

It is probable that the unconformities separating these units in outcrop disappear into the Arkoma basin to the south and the Mississippi embayment to the east, where gradational contacts would be expected. The outcrop observations noted above can be explained best by regression, which resulted from epeirogenic uplift, followed by erosion, during which some rock was removed. Erosion on the broad, gentle folds formed during uplift thinned units across structural highs and produced the variable and unpredictable thicknesses of the formations. There is no indication that significant topography was developed on these erosional surfaces in northern Arkansas. Uplift was consistently greater to the west, as indicated by the thinning and disappearance of units in that direction. Subsequent transgression brought offshore facies over near-shore ones, producing the deepening-upward succession of units separated by sharp, smooth contacts observed in outcrop.

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CONODONT BIOSTRATIGRAPHY OF LOWER ORDOVICIAN ROCKS, ARBUCKLE GROUP, SOUTHERN OKLAHOMA

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INTRODUCTION

The Arbuckle Group of southern Oklahoma displays the only complete exposure of the shallow-water carbonate rocks that characterize the Lower Ordovician of interior North America. Trilobites have been described from the lower formations of this Ordovician sequence (Stitt, 1971, 1977, 1983), and sporadic occurrences of other fossil invertebrates are known, but much of the section is sparingly fossiliferous. As a consequence, these magnificent exposures have not contributed notably to continuing efforts toward development of a comprehensive biostratigraphic scheme for the Lower Ordovician of North America.

In a project supported by a grant from the National Science Foundation, we collected samples at stratigraphic intervals averaging 25 ft apart through the Arbuckle from the upper part of the Signal Mountain Formation to the highest exposed units of the West Spring Creek Formation along Interstate Highway 35 and in adjacent ranch fields on the south flank of the Arbuckle Anticline near Ardmore, Oklahoma. Approximately 95% of these samples were productive of conodonts in abundances ranging from a few tens to >1,000 elements per kilogram.

Preservation of the specimens is good to excellent; CAI values of slightly over 1 show that these fossils have been exposed to paleotemperatures <100°C. The stratigraphic ranges of these conodonts show that the Arbuckle can be divided into a succession of biostratigraphic intervals; work in progress on the equivalent section in the Wichita Mountains in southwest Oklahoma indicates that the same succession of conodonts is represented there, and supports the conclusion that this sequence of faunas can be used to correlate the Arbuckle rocks of the subsurface. In addition, this biostratigraphic continuum, based on collections from an uninterrupted section, offers a standard for analysis of the numerous geographically localized and stratigraphically limited outcrops of Lower Ordovician strata in the Ozark Mountains and Upper Mississippi Valley.

SIGNAL MOUNTAIN, BUTTERLY, AND McKENZIE HILL FORMATIONS

A low-diversity association of conodonts is present in the upper Signal Mountain and through the Butterly. The lineage of species of *Cordylodus* Pander is a dominant component of all of the productive samples in this interval. Depending on which species is selected by international agreement for definition of the base of the Ordovician System, the Cambrian/Ordovician boundary will be placed at the base of the range of *C. proavus* in the upper Signal Mountain, or within the range of *C. intermedius* high in the Butterly. The conodonts from the Butterly provide biostratigraphic control for a unit that previously had produced only a few invertebrate fossils. The conodont population diversified during the time of deposition of the McKenzie Hill Formation, which documents the first major evolutionary radiation of these organisms. The distinctive assemblage of conodonts that is present in all but lowest McKenzie Hill has been reported widely from North America and has been interpreted as representing a unique zonal interval in the biostratigraphy of the Lower Ordovician of this continent (*Rossodus manitouensis* Zone; Landing and others, 1986).

COOL CREEK FORMATION

The Cool Creek Formation is the least-fossiliferous of the Ordovician formations within the Arbuckle Group. Other than the brachiopod *Diaphelasma oklahomense* Ulrich and Cooper and the lithistid sponge *Archaeoscyphia*, the fossils recorded from this unit are stromatolites of limited stratigraphic value.

The boundary between the Cool Creek and McKenzie Hill Formations is marked by an abrupt change in the conodont faunas. The diverse assemblage of the McKenzie Hill is replaced by a low-diversity fauna dominated by species of *Oneotodus* Lindström. This marked boundary in the conodont succession has been recognized at many localities across the North American craton;

it offers a unique event in the development of Early Ordovician conodont faunas that can be used for accurate correlations over long distances.

The occurrence of *Macerodus dianae* Fåhraeus and Nowlan in the Cool Creek is of significance for intercontinental correlation of Ordovician strata. This species occurs in Newfoundland in association with the graptolite *Tetragraptus approximatus*, a species that has been used to correlate rocks in North America with the lower part of the Arenigian Series of the classic British Ordovician succession. Thus, the presence within the Cool Creek of *M. dianae* probably indicates Tremadocian/Arenigian boundary beds at this stratigraphic level in the southern Midcontinent.

Four species—aff. *Drepanoistodus inaequalis sensu* van Wamel, *Eucharodus parallelus (Branson and Mehl), Glyptoconus quadraplicatus (Branson and Mehl),* and *Ulrichodina abnormalis (Branson and Mehl)*—appear near the top of the Cool Creek and range through the overlying Arbuckle strata to near the base of the Middle Ordovician. These species are present in and are commonly the dominant components of the conodont faunas of medial and upper Lower Ordovician strata throughout North America.

KINDBLADE FORMATION

The Kindblade Formation contains a more diverse association of conodonts than the underlying Cool Creek. Many species are long-ranging, some with their lowest occurrence in the Cool Creek below, and do not offer biostratigraphic resolution. Additional species appear through the Kindblade section, and some of these have ranges that allow them to be used to identify parts of the formation. The lowest occurrence of the albid, heavy-ribbed *Oneotodus costatus* is 100 ft above the base of the formation; the bottom of its range indicates a position low in the Kindblade. *Acodus deltatus* is abundant through the middle third of the Kindblade, and is joined in the upper half of its range by *Protoprioniodus russoi*. These species are supplanted up-section by a fauna that includes *Oneotodus carlae, Baltoniodus oepiki,* a species of *Protopanderodus,* and *Protoprioniodus simplissimus.* The presence of *Diaphorodus delicatus, Oepikodus communis,* and *Protoprioniodus papiliosus* in the top 60 ft of the Kindblade provides criteria for relating this part of the Arbuckle to equivalent strata across North America. These species, which are characteristic of the *O. communis* Zone, are the dominant elements in many of the samples collected in the overlying West Spring Creek Formation.

WEST SPRING CREEK FORMATION

Almost half of the taxa from the West Spring Creek range through all but the upper 100 ft of the formation. Many of these species are present in the Kindblade, and several range down to the upper Cool Creek. Nevertheless, a variety of species appear at successive levels within the West Spring Creek and allow correlations with other sections via the bases of their ranges. A marked change in the faunas occurs near the top of the exposures of the West Spring Creek along I-35 (the upper 45 ft of the formation are not available there). Almost all of the long-ranging species disappear at this level, and a new conodont population is introduced. These new taxa include hyaline forms that dominate lower Middle Ordovician faunas of the lower part of the Simpson Group. Their presence in the upper West Spring Creek reaffirms the earlier interpretation of Derby (1969) that the Lower Ordovician/Middle Ordovician boundary is ~100 ft beneath the top of the Arbuckle Group in the Arbuckle Mountains. Most of the samples taken in the Middle Ordovician part of the Abuckle contain specimens that show significant abrasion, indicating quite turbulent water and probably very shallow conditions at the time of deposition.

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REGIONAL ENVIRONMENTS OF DEPOSITION DURING CAMBRIAN AND ORDOVICIAN TIME IN WESTERN ARKANSAS

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ABSTRACT.—During Late Cambrian through Ordovician time, the Ozark shelf of northwestern Arkansas was a subsiding, shallow-water carbonate platform bordered on the east by the graben of a major rift and on the south by the deep-water, craton-edge, proto-Ouachita basin.

On a much larger scale, Middle Cambrian open seas began advancing from both the east and west onto the stable North American craton—at about the same time that the sea of the proto-Ouachita basin began to advance northward along the subsiding floor of the Mississippi Valley graben of eastern Arkansas, where thick lenses of arkose (mostly red) had been accumulating. The eastern open sea and Ouachita restricted sea entered the graben from opposite directions and, as a unit, continued transgressing westward across the Midcontinent. After that transgression, deposition in the graben probably was nearly continuous through Ordovician time, whereas deposition on the shelf of northwestern Arkansas was cyclic during the Late Cambrian and Early Ordovician (minor transgressions/regressions). During the Middle and Late Ordovician, major regressions periodically interrupted deposition in the shelf area, allowing up-slope truncation. Intermittently, fine- to medium-grained (locally coarse-grained) rounded-and-frosted quartz sand from the north spread across the shelf to form sandstone units that divide the predominantly carbonate sequence (locally cherty as a result of diagenetic changes, and locally shaly) into regional stratigraphic units.

To the south in the Ouachita region, neither the earliest Cambrian deposits nor the underlying crust have been reached by drilling. Speculations concerning their character stem from geophysical data and conceptual models. However, Late Cambrian and Ordovician sediments deposited in this protobasin are now exposed in the allochthonous assemblage that makes up the core of the Ouachita Mountains. These (and older Ouachita strata insinuated by concept) accumulated along the passive southern edge of the North American craton, probably in a Cambrian rift zone that was flanked on the outboard side by an orogenic landmass. In spite of the lack of any known beds older than mid-Late Cambrian, sediments of the "Ouachita facies" must have begun to accumulate as soon as rifting started, probably at least by the beginning of Cambrian time.

The known rock types from the Ouachita protobasin, in decreasing order of abundance, are shale, siltstone, sandstone, chert, limestone, and conglomerate (locally containing metamorphic and igneous clasts). This sequence is made up of both (1) deep-water clastics from multiple external sources, and (2) indigenous pelagic or mostly indigenous hemipelagic deposits. Thick units—mainly medium- to coarse-grained sandstone, but including some olistostromal intervals in the Early Ordovician Crystal Mountain Sandstone and the Middle Ordovician Blakely Sandstone—were derived from the shelf and submarine scarps to the north. Detritus that makes up the siltstone of the Lower Ordovician Mazarn Shale and wacke of the Middle Ordovician upper Womble Shale likely was derived from sources to the east and south. Masses of metagabbro (Precambrian) and serpentinite (of unknown age) occur at three sites in the upper Womble, offering a glimpse of the earlier history of this region.

SUBSURFACE STRUCTURE MAP OF THE ARBUCKLE GROUP, SOUTH-CENTRAL OKLAHOMA

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The future role of the Cambrian–Ordovician Arbuckle Group is one of the most important questions in the development of hydrocarbon resources in Oklahoma. The Arbuckle attains a thickness of >7,000 ft in southern Oklahoma where it occurs in outcrop and at depths of 20,000 ft or more in the subsurface. It has been penetrated in at least 1,900 wells in the area; however, the average depth drilled into the Arbuckle is <380 ft. The recent discovery of significant hydrocarbons at the Cottonwood Creek field has renewed interest in the Arbuckle. A structure contour map of the Arbuckle Group was created for Bryan, Caddo, Carter, Commanche, Cotton, Garvin, Grady, Jefferson, Johnston, Love, Marshall, Murray, and Stephens Counties (Fig. 1).

The database used for the map was created from the commercially available Petroleum Information Well History Control System[1]. This information was analyzed to determine the locations of probable faults: the data were then contoured, manually and by computer, to create Figure 1. The program, Interactive Surface Modeling by Dynamic Graphics Inc.[1] was used for computer processing of these data.

The map shows a strong W–NW trend of relief on the Arbuckle surface. Well-data used for this map, though not uniformly distributed, are dense and deep enough to yield a fairly detailed interpretation of the upper surface of the Arbuckle in this area. Formal publication of this map at a scale of 1:500,000 is planned through the Oklahoma Geological Survey.

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[1]The use of trade names or names of corporations is for descriptive purposes only and does not constitute an endorsement by the U.S. Geological Survey.

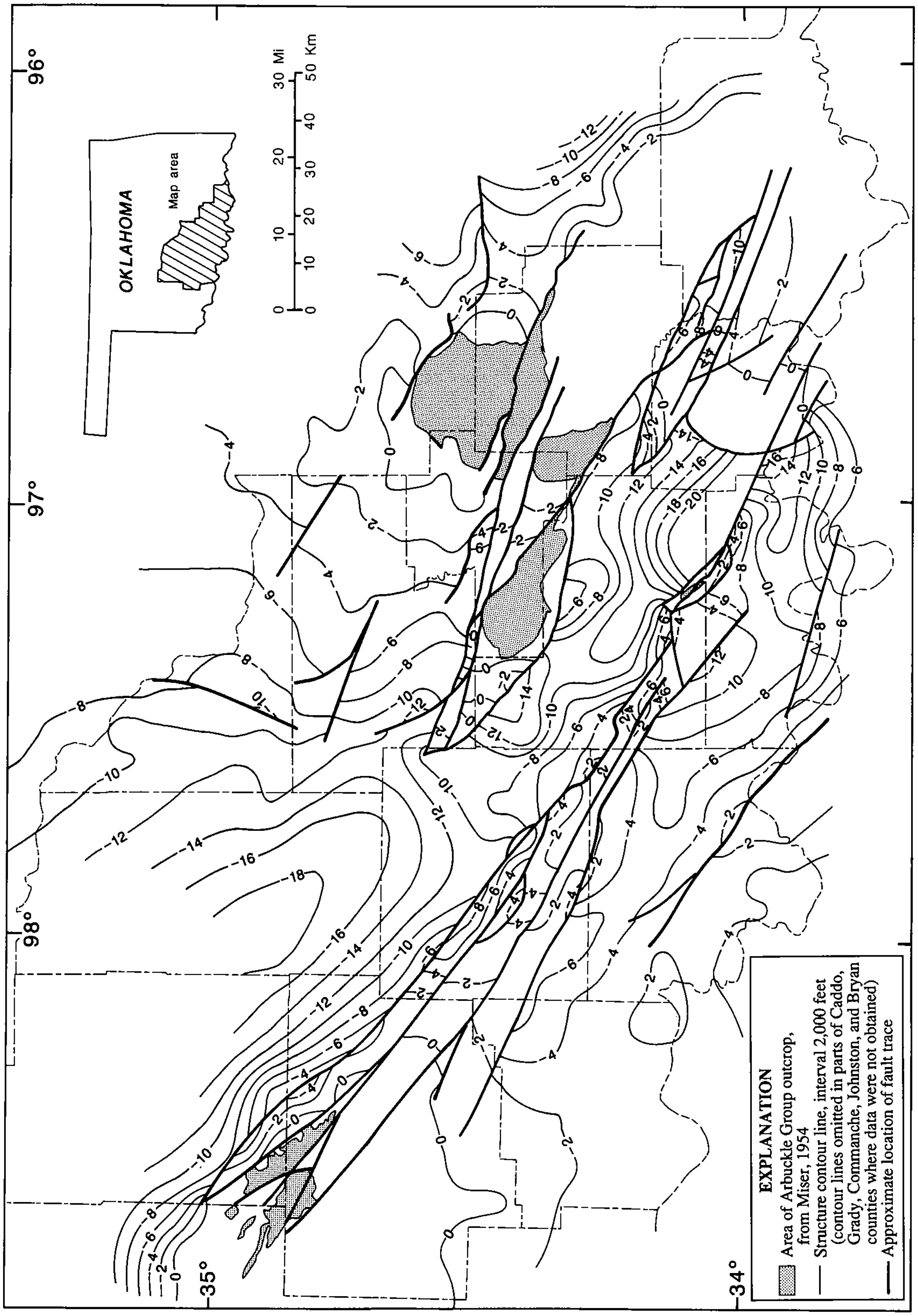

Figure 1. Structure contour map of the top of the Arbuckle Group.

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NONFUEL MINERAL RESOURCES IN LATE CAMBRIAN AND ORDOVICIAN ROCKS OF OKLAHOMA

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ABSTRACT.—Principal nonfuel mineral resources in Late Cambrian and Ordovician rocks of Oklahoma include limestone, dolomite, silica sand, and rock asphalt. Other resources include shale, hematite, limonite, manganese, zinc, lead, chert, and quartz veins. The largest area of outcrops of these mineral deposits is in the Arbuckle Mountains, although important outcrops are present in the Wichita and Ouachita Mountains and the Criner Hills, and small exposures are scattered in the Ozark uplift.

Enormous reserves of limestone, mainly in the Arbuckle and Viola Groups, are quarried in the Arbuckle and Wichita Mountains for aggregate (crushed stone). Quarries producing millions of tons of stone each year are the main source of aggregate for the Oklahoma City metropolitan area, as well as the southern and western parts of the State. Abundant reserves of high-purity dolomite are present in the Royer and Butterly Dolomites of the Arbuckle Group; they are mined at three localities in the Arbuckle Mountains for fluxing stone, glass manufacture, refractories, animal feeds, and conventional aggregate uses.

Large reserves of high-purity silica sand in the Simpson Group are mined hydraulically at three places in the Arbuckle Mountains. Crude sand consisting of 98% silica is upgraded to 99.8% silica, and the product is marketed for glass-making, foundry sands, ceramics, and the manufacture of sodium silicate. Additional small resources are present in the Ozarks region. Outcropping Simpson sandstones and Viola limestones locally are impregnated with petroleum and constitute large reserves of natural rock asphalt. Deposits of rock asphalt in the Arbuckle Mountains supported a major part of the local road-surfacing activities from 1891 to 1960, but all natural rock-asphalt quarries are now inactive.

Cement is manufactured from Sylvan Shale and the Viola and basal Hunton (Keel oolite) limestones south of Ada in the Arbuckle Mountains. In the Wichita Mountains, small deposits of low-grade hematite in the Reagan Formation have been used as a paint pigment, whereas in the Arbuckle Mountains a number of small deposits of limonite (derived by intense weathering of iron-bearing parts of the Arbuckle Group) have been mined as a source of high-grade iron ore. Arbuckle strata in the Arbuckle Mountains host small deposits of manganese oxide, as well as low-grade vein deposits of zinc and lead. Chert and veins and crystals of colorless and milky quartz are present at scattered locations in Cambrian and Ordovician strata of the Ouachita Mountains, but these have not been developed commercially.

A more thorough discussion of Oklahoma's nonfuel mineral resources is given by Johnson elsewhere in this volume.

CONODONT BIOSTRATIGRAPHY OF LOWER AND MIDDLE ORDOVICIAN ROCKS IN THE BENTON UPLIFT, WEST-CENTRAL ARKANSAS

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INTRODUCTION

The Lower and Middle Ordovician rocks of the core of the Ouachita Mountains in west-central Arkansas comprise a thick succession of three black, graptolitic shale units that alternate with two intervals of sandstone; the upper shale unit is overlain by a succession that consists primarily of cherts. These units were defined as the Collier Shale, Crystal Mountain Sandstone, Mazarn Shale, Blakely Sandstone, Womble Shale, and Bigfork Chert (in ascending order) early in this century through the work of Purdue, Miser, and Ulrich. This sequence is a classic example of graptolite facies, and the accepted ages of these units for the past 75 years have been based on interpretation of occurrences of graptolites in them. Occasional reports of other fossils in these rocks have not contributed significantly to biostratigraphic evaluation of the Ouachita sequence.

All of the units in the Benton uplift contain limestones, samples of which have been processed for recovery of conodonts. Thin, micritic ribbon limestones typically are only a few inches thick and are transitional lithologically with the enclosing shales. Such limestones usually give voluminous silty residues when digested in acetic acid and yield no more than a few tens of conodont elements per kilogram of rock; preservation of the specimens is moderate to poor. Most of the conodonts recovered from the Collier and the Mazarn were found in limestones of this type. Grainstones up to 1 ft thick are present in the Womble, in addition to ribbon limestones like those in the older units. The grainstones produce acid residues that contain an abundance of highly spherical, vitreous quartz grains that commonly show frosted surfaces as a result of quartz overgrowths. Recovery of hundreds of conodont elements per kilogram of rock from these grainstones is common; preservation of the specimens ranges from moderate to excellent. In addition to conodonts, silicified invertebrates (trilobites, ostracodes, bryozoans) have been found in residues from the Collier and Womble. In all cases, the conodonts are dark gray to opaque black, indicating exposure to paleotemperatures up to as high as 300°C.

Large collections of conodonts have been assembled from the Womble in exposures east and south of Lake Ouachita, and somewhat smaller numbers have been obtained from the Collier and Mazarn. In addition, these fossils have been found in limestone interbeds and in clasts in debris flow deposits in the Crystal Mountain and Blakely sandstones. Study of the Bigfork Chert is less advanced, but collections to date show that limestones near the base of that formation contain well-preserved conodonts in moderate abundance. The Ouachita conodonts include species that have been used to establish a succession of biostratigraphic zones in thick, continuous sequences of Ordovician rocks elsewhere in North America. It is possible to place samples collected from geographically isolated and stratigraphically limited outcrops in the Benton uplift in proper stratigraphic order by referring the conodonts recovered from them to this external standard, thereby achieving a greater degree of biostratigraphic resolution than previously has been possible. Conodonts, where available, will greatly assist in regional geologic analyses of the structurally complicated Ouachita region.

COLLIER SHALE

The conodonts from the Collier reported by Repetski and Ethington (1977), as well as most of those recovered subsequently, are forms typical of the *Rossodus manitouensis* Zone. This assemblage of species is widely distributed across North America and has been found in the Lower Ordovician of the Siberian platform, but is not known from western Europe. Its presence in the Collier demonstrates that the upper part of that formation is correlative with the McKenzie Hill Formation in Oklahoma and with the Gasconade Formation in Missouri. The Collier locally exposes strata as old as those of the Upper Cambrian *Elvinia* Zone (Hart and others, 1987); protoconodonts characteristic of the Upper Cambrian have been found in these outcrops. Collier conodonts are sparse in the productive samples; preservation commonly is moderate to poor, and size is very small.

MAZARN SHALE

At least half of the samples collected from the Mazarn have been barren of conodonts. Many of the productive samples have yielded only a few elements per kilogram of dissolved rock, but these few specimens are sufficiently distinctive to demonstrate that they came from the Mazarn. Fortunately, some samples have provided abundant, moderately diverse, and reasonably well-preserved collections. These collections contain representatives of the warm-water species that have been attributed to the North American Midcontinent Province, as well as others that are characteristic of the cold-water North Atlantic Province. Some samples are dominated by one of these assemblages to the near exclusion of the other, but other samples contain representatives of the two faunas in about equal numbers. We conclude that the cold-water conodonts represent an indigenous population that inhabited the deep waters in which the Ouachita facies were deposited, and that the warm-water forms were carried into the basin in slurries of lime mud derived from a shallow shelf where those species lived. Two conodont zones have been recognized in the Mazarn, an older one characterized by *Acodus deltatus* Lindström, and a younger one with *Oepikodus evae* Lindström. The *O. evae* Zone is present in highest Lower Ordovician elsewhere in North America, so that the top of the Mazarn may be as young as early Whiterockian. Additional collections are being assembled in an effort to establish the stratigraphic level of the lower and upper boundaries of the Mazarn.

BLAKELY SANDSTONE

Conodonts have been found in limestone beds and in clasts in debris flow deposits in outcrops of Blakely west of Crystal Springs, Arkansas, and along the eastern shores of Lake Ouachita. These outcrops are interpreted to represent the middle to upper part of the formation. The conodonts include species such as *Histiodella holodentata* Ethington and Clark and *"Cordylodus" horridus* Barnes and Poplawski, forms that elsewhere in North America are characteristic of the middle Whiterockian Fauna 4 of Ethington and Clark (1971). The Blakely thus seems to represent the pre-Chazyan part of the Whiterockian Series; it has equivalents in the Joins and Oil Creek Formations (lower Simpson Group) of southern Oklahoma.

WOMBLE SHALE

Conodonts are more abundant and diverse in the Womble than in the older units of the Ouachita facies; over three-fourths of the processed samples have been productive, and the specimens are more robust and better preserved than their older counterparts from the Benton uplift. Three distinct associations have been identified. The oldest of these is an assemblage of species that Bergström (1971 and subsequently) has considered to be characteristic of the *Pygodus serra* Zone. This zone has been subdivided into subzones based on the ranges of species of *Eoplacognathus,* some of which have been found in our studies. They offer the possibility of eventually applying a high-resolution biostratigraphy to the interpretation of outcrops of the lower Womble. This zone is present in the road cuts near Walnut Creek west of Crystal Springs, and has been recognized along the shore in the eastern part of Lake Ouachita and near Jessieville. Outcrops of the Womble that yield this association of conodonts can be correlated with the McLish–Tulip Creek interval in Oklahoma and with lower Chazyan rocks elsewhere in North America.

The Zone of *Pygodus anserinus* is present in the massive limestone ledges of the quarry at Mountain Pine, Arkansas, but has not been found as widely distributed as that of *P. serra.* This part of the Womble is of late Chazyan age; the nearest correlative unit is the lower part of the Mountain Lake member of the Bromide Formation in Oklahoma.

Upper Womble contains an assemblage with typical species of the Midcontinent Province including *Phragmodus undatus, Plectodina* spp., *Ansella robusta, Dapsilodus nevadensis,* and *Curtognathus* sp. Also present in this interval are North Atlantic forms such as *Prioniodus gerdae, Periodon aculeatus,* and species of *Eoplacognathus* and *Amorphognathus.* The latter genera have been used by Bergström for a high resolution biostratigraphy in the southern Appalachians, and allow correlations of Ouachita formations with other stratigraphic units in the Appalachian–Ouachita orogen. Unfortunately the large platform elements commonly have suffered severely during the deformation of the Ouachita rocks so that recovery of complete, easily identified material requires processing large samples. The youngest conodont zone so far recognized in the Womble is that of *Prioniodus gerdae* which has been found along the Caddo River west of Norman, Arkansas. Collections that may indicate even younger Womble have been made west of Caddo Gap and near Crystal Springs. Each of these collections was made in outcrops believed to be close to the boundary with the overlying Bigfork Chert. Efforts are underway to recover conodonts by hydrofluoric acid treatment of Bigfork cherts in order to clarify and stabilize the position of this stratigraphic boundary.

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POROSITY DEVELOPMENT IN QUARTZ-RICH, OOLITIC LIMESTONES OF THE UPPER ARBUCKLE GROUP—A RESPONSE TO UNLOADING

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The Early Ordovician Arbuckle Group of southern Oklahoma is composed of a thick sequence of carbonates that were deposited in shallow, epicontinental seas (Gatewood, 1978a,b; Ragland and Donovan, 1986). Several concentrations of quartz-sandy detritus, two of which are significant markers in outcrop, are found in the Cool Creek and Kindblade Formations. The Thatcher Creek Sandstone Member at the base of the Cool Creek Formation represents a significant increase of quartz detritus into the Oklahoma aulacogen. The Herringbone sand is a distinctive, cross-bedded, quartz-rich carbonate unit occurring ~3m above the base of the Kindblade Formation.

Several types of porosity modified by cementation and/or dissolution are found in the Arbuckle carbonates in the subsurface (Gatewood, 1978b; Shirley, 1989). Porosity is less prominent in the surface exposures of the Slick Hills (northeast of the Wichita Mountains in southwestern Oklahoma), but includes interparticle, intraparticle, and intercrystalline porosity. All types are modified by compaction, cementation, replacement, dissolution, and fracturing (Cloyd and others, 1986; Donovan, 1987).

Much of the porosity in the quartz-sandy units results from megascopic and microscopic fracturing. Although many of the fractures are pervasive and non-fabric-selective, three types of fabric-controlled fractures occur (Ragland and Matthews, 1989):

1) Fractures resulting in voids between different minerals (Fig. 1). Quartz grains with syntaxial overgrowths are separated from carbonate grains and cements by fractures tracing the outermost edges of the overgrowths. Typically, the width of the pore is nearly uniform around the grain and faithfully records the shape of the syntaxial overgrowth. Overgrowths are not separated from the detrital quartz grains, which reflects the strength of the bond between the grains and quartz cement. Where grains are devoid of overgrowths, nearly circular cross-sectional voids surround rounded quartz grains separating the grains from micritic matrix and carbonate allochems.

2) Fractures resulting in voids between different textures with the same mineralogy (Fig. 1). Ooids in the sandstones have large nuclei relative to the thickness of the cortices. Most nuclei are composed of micrite, although an occasional grain of quartz silt or fine sand was utilized as a nucleus. The radial cortices pulled away from the dense micritic nuclei, leaving circular or near-circular cross-sectional voids. The voids occur at or near the outer edges of the nuclei. The cortices appear to have bonded tightly to the outermost rim of the nuclei and, when fracturing occurred, pulled away a few micrite crystals. Many nuclei appear to have "dropped" when support was removed, suggesting that the voids are completely encompassing. Fracturing also occurred between adjacent radial

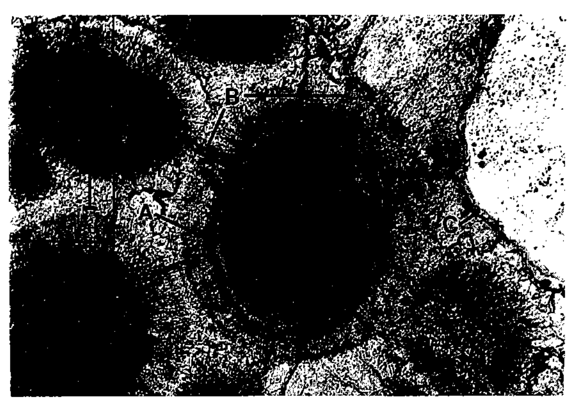

Figure 1. Fabric-controlled fracture porosity (A) surrounds nucleus of an ooid. Enclosure is incomplete; the nucleus appears to have "dropped" to the lower right. Radial fractures (B) connect the inner ring with other fabric-controlled fractures and interparticle porosity. Syntaxial quartz overgrowth on a detrital quartz grain in upper right is encompassed by fabric-controlled porosity (C). Scale bar equals 250 μ; ordinary light.

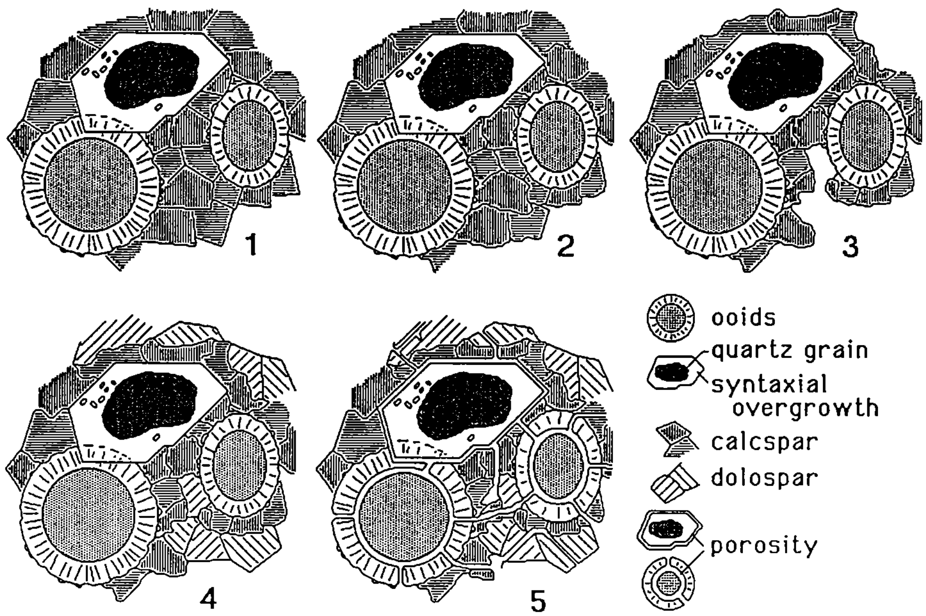

Figure 2. Stages in the development of fabric-controlled porosity in quartz-sandy, oolitic units of the Cool Creek and Kindblade Formations. Each diagram represents ~4 mm². See text for discussion.

crystals in the cortices. The radially oriented fractures connect the voids encasing the nuclei with interparticle and intercrystalline porosity.

3) Fractures resulting in voids between crystals. Porosity in calcite and dolomite cements occurs along crystal boundaries. The fractures provide interconnecting conduits between other inter- and intraparticle pore spaces.

Fabric-controlled fracture porosity appears to be the result of pressure release following the removal of overburden. The development of the fractures may have followed the course of events outlined below and illustrated in Figure 2:

1) Diagenesis of sediments began immediately after deposition. Carbonate grains and isopachous rims of calcite cement were slightly etched as syntaxial quartz overgrowths were formed around quartz detritus. Pore spaces were partially filled with the overgrowths; some carbonate grains were partially replaced by the silica. As burial continued, sparry calcite filled the remaining pores.

2) Deep burial resulted in pressure solution and distortion of calcite-crystal boundaries.

3) After uplift and deformation, quartz-rich oolites were exposed to near-surface and surface dissolution. Vuggy porosity was formed.

4) Sparry dolomite filled vuggy porosity during reburial.

5) Removal of overburden without concurrent dissolution resulted in fracture porosity. Non-fabric-selective fractures cross-cut mineralogies and textures. Fabric-selective fractures formed around nuclei, between cortical fibers, between crystals, and around the outer edges of syntaxial overgrowths. Fractures were enlarged through dissolution.

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PRELIMINARY REPORT ON THE PALEONTOLOGY AND STRUCTURAL GEOLOGY OF NEWLY DISCOVERED EXPOSURES OF COLLIER SHALE, BENTON UPLIFT, EASTERN OUACHITA MOUNTAINS, ARKANSAS

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The Collier Shale is the oldest formation known to crop out in the Ouachita Mountains of Arkansas and Oklahoma. Prior to 1977, most workers assigned a Cambrian age to the Collier, based on its stratigraphic position below strata which, in part, contained Early Ordovician graptolites. Diagnostic Early Ordovician conodonts representative of the *Rossodus manitouensis* Zone were reported from limestones in the upper Collier Shale in the type area in Montgomery County, Arkansas, and in McCurtain County, Oklahoma (Repetski and Ethington, 1977).

During recent mapping and biostratigraphic investigations in northern Garland and western Saline Counties, Arkansas, several previously unrecognized areas of Collier Shale were discovered near Lena Landing on Lake Ouachita (Focht, 1981), near Jessieville (Hart and others, 1986), and near Buckville on Lake Ouachita and south of Paron (Stone and Ethington, this paper). These exposures of Collier are complexly deformed and occur in two or more thrust plates. The Collier Shale determinations are strongly supported by faunal evaluations made by Stitt and Hohensee (trilobites) and Ethington (conodonts) from specimens obtained from thin intervals of dark limestones.

The largest, most studied of the new exposures is near Jessieville, where ~30 km^2 of Collier occurs (Hart and others, 1987). Several thin limestone beds near the top of the Collier yield conodonts of the *Rossodus manitouensis* Zone like those from the type area. Near the base of the exposed Collier, trilobites characteristic of the *Elvinia* and *Taenicephalus* Zones (Franconian Stage) of the Upper Cambrian were recovered from clasts and discontinuous beds of dark, pyritic limestone at 13 localities (Hohensee and Stitt, 1989). Abundant species include *Aphelotoxon lumaleasa, Cliffia lataegenae, Comanchia amplooculata, Dellea suada, Housia vacuna, Irvingella major, Kindbladia wichitaensis, Kymagnostus harti, Neoagnostus? dilatus, Pulchricapitus fetosus,* and *Parabolinoides* spp. Many of the same species of Upper Cambrian trilobites have also been recovered from limestone clasts in the Collier near Lena Landing. Most of these trilobite species are abundant in Upper Cambrian shallow-water limestones in Missouri, Oklahoma, and Texas.

Most of the trilobites are very small (1–2 mm in length), disarticulated, unsorted, and unabraded. Many specimens representing early growth stages are present alongside adult forms. The fossils appear to have accumulated near where they were shed as molts, and the fauna has the high species diversity and style of preservation characteristic of shelf trilobites. Some features of the fauna suggest deposition in a deep-water, outer-shelf setting, rather than in a shallow-water location. The fossils are found in lenses and thin beds of black to dark-gray, pyritic, peloidal to micritic limestone, rather than the light-colored limestone characteristic of shallow-water Cambrian shelf carbonates. The many immature forms present attest to the lack of sorting in this environment, and suggest deposition in quiet, perhaps deep water. Agnostid trilobites are very abundant, and Robison (1976) pointed out that agnostid trilobites were most abundant in outer-shelf areas that faced open oceans. The absence of olenid trilobites argues against a slope environment. We believe that these Upper Cambrian trilobites lived in a deep-water, outer-shelf setting.

However, this was not the final resting place for these fossils. The lower Paleozoic strata of the Ouachitas are believed by most geologists to have been deposited in deep water on the continental slope or in a deep ocean basin. Most of the Collier consists of gray to black, diagenetically and sometimes metamorphically altered shale that is inferred to have a hemipelagic origin. Several distinctive types of gravity-flow deposits can be distinguished in the trilobite-bearing beds in the Collier, including channel-form debris-flow deposits, sheet-form debris-flow deposits, carbonate turbidity-flow deposits, and grainstone gravity-flow deposits. The first three types of deposits contain clasts that were lithified before they were eroded and transported, but the grainstone gravity-flow deposits represent redeposition of previously unconsolidated materials.

We believe that the limestone clasts containing Late Cambrian trilobites from the *Elvinia* and *Taenicephalus* Zones were eroded from their original depositional site on the outer shelf and redeposited in deeper water on the adjacent continental slope or basinal plain as channel and sheet-form deposits and in carbonate turbidites. These clast beds are interbedded with trilobite-bearing, bioclastic limestones that are interpreted as gravity-flow deposits of previously unconsolidated materials. All of these beds were emplaced at approximately the same time, and thus the deposition of the Collier began in the Late Cambrian and, based on the conodont evidence, continued into the Early Ordovician.

Additional Early Ordovician conodonts from the *Rossodus manitouensis* Zone have been recovered from previously unrecognized exposures of the Collier near Buckville and south of Paron. A few conodonts from the Upper Cambrian (Franconian) *Proconodontus tenuiserratus* Zone occur with *Elvinia* Zone trilobites at one of the Jessieville trilobite localities.

The eastern Benton uplift is a large structural culmination with folds and thrust nappes comprising mostly pre-middle Mississippian strata. A complex structural history is indicated by the rocks exposed in this part of the uplift, with several phases of deformation overprinting one another. The two dominant phases are (1) major N-directed thrust faults and nappes; and (2) later, S-verging chevron folds and some thrust faults. This later phase caused refolding of the earlier structures, and generated low-rank metamorphism and quartz veins.

At the Jessieville and Lena Landing localities, the Collier Shale occurs mostly along the crest of a broadly arched, doubly plunging, thrust-faulted anticlinorium that trends NE and verges SE (Hart and others, 1987). It appears that these Collier outcrops were originally part of a large, N-directed thrust, and that the anticlinorium is mostly a later, superposed feature. At the Paron locality, the Collier Shale represents a small part of the overriding strata of the Alum Fork thrust-fault system. The Alum Fork thrust is one of a series of very large, N-directed thrust faults that have telescoped deeper basinal sequences many miles compared to the strata that occur immediately to the west. The Collier Shale at the Paron locality represents more-distal basinal deposits than those exposed at Jessieville, Lena Landing, or Buckville.

Several important interpretations are based on the present studies: (1) the Collier Shale accumulated in a deep-water marine environment, and contains some limestones deposited as sediment gravity flows from the nearby outer shelf; (2) the diverse fauna of *Elvinia* Zone trilobites recovered from the Collier can be correlated with the middle and upper parts of the *Elvinia* Zone in Oklahoma, Texas, Missouri, and elsewhere in North America; (3) the occurrence in the Collier Shale of Late Cambrian trilobites of North American affinities establishes that the Benton uplift of the eastern Ouachita Mountains of Arkansas is not an exotic terrane; (4) paleontologic evidence indicates that deposition of the Collier was slow, and that it began in the Late Cambrian and continued into the Early Ordovician; (5) mostly via northward thrusting and intense compression, a series of complexly deformed allochthonous plates, some containing Collier Shale, were thrust northward out of the basin flanking the North American continent and stacked one on another in the eastern Benton uplift; (6) the shelf edge that existed during deposition of the Collier Shale now outcropping near Jessieville was probably originally located some tens of miles south along the southern flank of the Benton uplift; (7) additional exposures of Collier Shale await discovery in the eastern Benton uplift.

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