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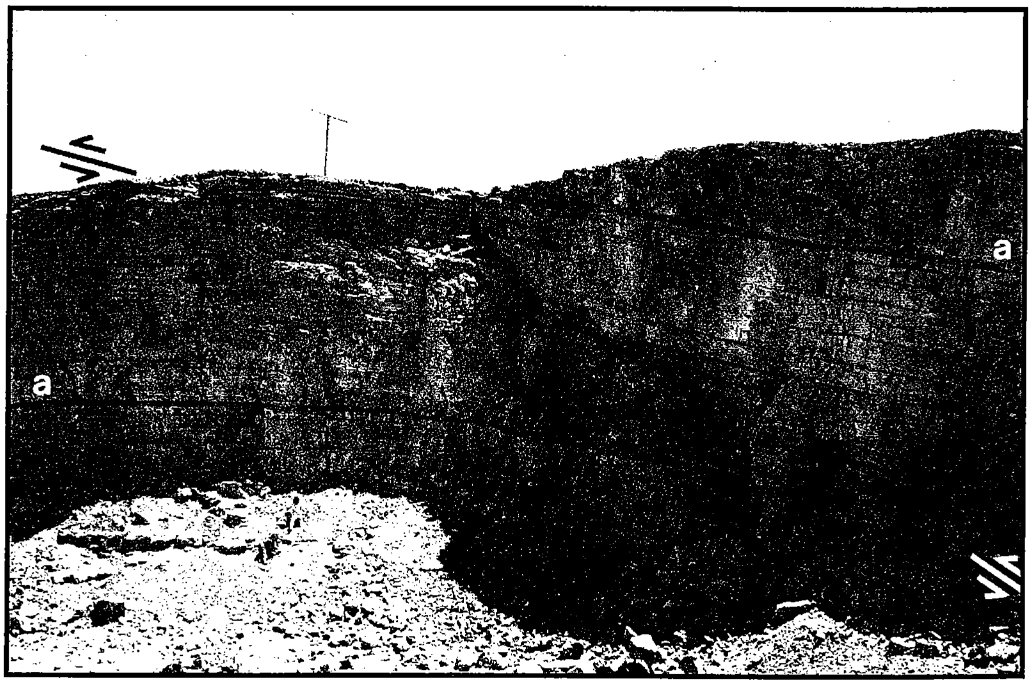
Hunton Group Core Workshop and Field Trip

Kenneth S. Johnson, *Editor*



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Hunton Group Core Workshop and Field Trip

Kenneth S. Johnson
Editor

Proceedings of a one-day core workshop (November 2, 1993) held in Norman, Oklahoma, and a one-day field trip (November 3, 1993) to the Arbuckle Mountains of southern Oklahoma.

Co-sponsored by:
Oklahoma Geological Survey
and
Bartlesville Project Office,
U.S. Department of Energy

Oklahoma Geological Survey
Charles J. Mankin, *Director*
The University of Oklahoma
Norman, Oklahoma

1993

SPECIAL PUBLICATION SERIES

The Oklahoma Geological Survey's Special Publication series is designed to bring timely geologic information to the public quickly and economically. Review and editing of this material has been minimized in order to expedite publication, and author-prepared illustrations have been used throughout.

Front Cover

Faulted anticline is beautifully exposed in Hunton anticline quarry (Stop 7 of field trip). Bois d'Arc strata are displaced by the low-angle thrust from south to north (right to left). Vertical offset across the thrust is about 20 ft, based on correlation of thin shale bed (a).

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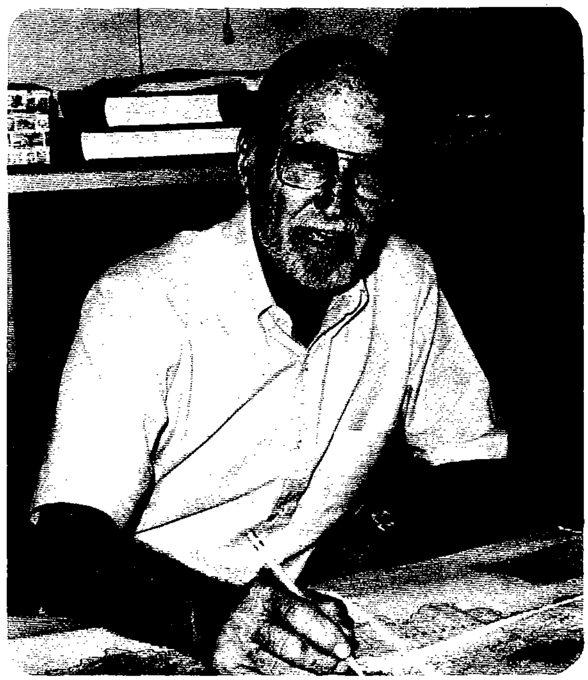

Thomas W. Amsden

DEDICATION

This publication is dedicated to Dr. Thomas W. Amsden in recognition of his many years of service to the Oklahoma Geological Survey, and especially for his devoted study of the Late Ordovician–Devonian Hunton Group in Oklahoma. He joined the Oklahoma Geological Survey in 1955, and worked almost exclusively on the Hunton Group until his retirement in 1985. Tom has long been recognized as a leading authority on Silurian–Devonian brachiopods, and on all aspects of geology of the Hunton Group. His detailed study of the biostratigraphy of the Hunton, based on outcrop and subsurface work, has established a wealth of data and understanding upon which other workers are now building. All the contributors to this volume have benefited greatly from Tom's work, as attested by their numerous references to his earlier publications. And so it is with profound thanks and appreciation for his highly professional work on the Hunton that his many colleagues and I dedicate this volume.

Kenneth S. Johnson

PREFACE

The Late Ordovician–Silurian–Devonian Hunton Group is a moderately thick sequence of shallow-marine carbonates deposited on the south edge of the North American craton. This rock unit is a major target for petroleum exploration and reservoir development in the southern Midcontinent. The workshop is being held to display cores, outcrop samples, and other reservoir-characterization studies of the Hunton Group and equivalent strata throughout the region.

The field trip is organized to complement the workshop by allowing examination of excellent outcrops of the Hunton Group in the Arbuckle Mountains.

The core workshop and the field trip cover such topics as: petroleum-reservoir characterization; deposition, diagenesis, facies, sequence stratigraphy, paleokarst, and petroleum production; and interpretation of Hunton Group geology based on petrologic, petrographic, and innovative subsurface studies. To facilitate the exchange of information on these important topics, the Oklahoma Geological Survey (OGS) and the Bartlesville Project Office of the U.S. Department of Energy (BPO–DOE) are co-sponsoring this workshop and field trip. The ultimate goal of this program is to increase activity in the search for, and the production of, our oil and gas resources.

Special thanks are expressed to the workshop presenters and field-trip leaders who contributed to this volume. Each of them was successful in synthesizing years of work and many data into the report. 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 organization and planning of this workshop include: Kenneth Johnson and Charles Mankin of the OGS; Tom Wesson, Michael Ray, and Edith Allison of BPO–DOE. Other OGS personnel who contributed include: Michelle Summers and Tammie Creel, Registration Co-Chairs; and Connie Smith, Publicity Chair. Field-trip planning was done by Zuhair Al-Shaieb (Oklahoma State University), Richard D. Fritz (MASERA Corp.), and James E. Barrick (Texas Tech University).

Kenneth S. Johnson
General Chairman

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PART 1

Core-Workshop Presentations

Overview of Hunton Facies and Reservoirs in the Anadarko Basin

Zuhair Al-Shaieb,
Geoff Beardall[1], Pat Medlock[2], Kathy Lippert[3], Felicia Matthews[4], and Frederica Manni[5]
Oklahoma State University
Stillwater, Oklahoma

ABSTRACT.—The Late Ordovician–Devonian-age Hunton Group was deposited in an epicontinental sea during a period of relatively slow subsidence. The Hunton Group consists of a series of shallow-water carbonates that are separated by disconformities in much of the region. The general stratigraphic divisions are the Chimneyhill Subgroup (Keel, Cochrane, and Clarita Formations), Henryhouse Formation, Haragan Formation, Bois d'Arc Formation, and Frisco Formation. The Henryhouse will be used to illustrate facies and reservoir characteristics that are generally analogous to all pre-Frisco Hunton units. The Frisco is genetically different from earlier Hunton rocks since it was deposited mainly as bryozoan/crinoid carbonate biohermal rocks on a post-epeirogenic unconformity surface.

The Upper Silurian Henryhouse Formation is a major oil and gas reservoir in the Anadarko basin, and a detailed examination of Henryhouse cores was conducted. Sedimentary structures, lithology, fossil content, and fabric relationships were used as criteria to recognize various depositional facies bands that parallel the paleoshoreline. Supratidal (I), intertidal (II), and subtidal (III) facies can be readily distinguished, and their spatial relationships consistently indicate shallowing upward sequences. Anhydrite occurs at the top of the sequence, suggesting that hypersaline conditions developed in supratidal environments.

Three stages of dolomitization are documented in the Henryhouse Formation. Petrographic, cathodoluminescent, and isotopic techniques were used to investigate the genesis and textural relationships of the various dolomite types. The following paragenetic sequence is discerned: (1) penecontemporaneous hypersaline dolomite occurring as brownish, hypidiotopic, 60–80 μm rhombs concentrated in the supratidal and intertidal facies; (2) marine and freshwater mixed dolomite occurring as white rims around pre-existing hypersaline dolomite and as anhedral white rhombs in vugs and molds; and (3) deep-burial vug, mold, and fracture-filling baroque dolomite.

The four types of porosity that developed are moldic, vuggy, intercrystalline, and fracture. Moldic and intercrystalline porosity are fabric selective, whereas vuggy and fracture porosity are nonfabric selective. Moldic and vuggy porosity are secondary, whereas intercrystalline can be either primary porosity preserved by dolomitization or late secondary dissolution porosity. Volumetrically, moldic porosity is the most important. Fracture porosity increases permeability in local areas, but is not an important widespread phenomenon. Dissolution is an important modifier of existing rock fabric and can greatly enhance moldic, intercrystalline, and fracture porosity. Since porosity is related directly to the depositional environment, it was best developed in shallow subtidal to upper intertidal facies. The characteristic features of these facies that caused them to preferentially develop porosity are burrowing and abundant pelmatozoan fossils.

Burrowing enhanced the porosity and permeability of the sediment, allowing dolomitizing fluids to more effectively operate, which preserved the enhanced porosity and permeability. Subsequent dissolving fluids then were able to attack the nondolomitized fossils, creating moldic porosity. The supratidal facies (I) is often extensively dolomitized, but failed to develop significant porosity without burrowing and fossils. Subtidal Henryhouse carbonate (facies III) was never highly dolomitized, due to both relative impermeability (never highly burrowed) and increased distance from the source of dolomitizing fluids. Moldic porosity is rare in the subtidal facies as fossils were never dissolved.

Porosity was also enhanced by karstic processes during epeirogenic uplifts. However, unconformity surfaces were not required for porosity development.

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Al-Shaieb, Zuhair; Beardall, Geoff; Medlock, Pat; Lippert, Kathy; Matthews, Felicia; and Manni, Frederica, 1993, Overview of Hunton facies and reservoirs in the Anadarko basin, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 3–39.

INTRODUCTION

The Hunton Group was named by J. A. Taff (1902) in his work on the geology of the Atoka Quadrangle. In 1910, C. A. Reeds defended his doctoral dissertation, "The Stratigraphy of the Hunton Formation." In this stratigraphic and paleontological study, Reeds proposed a fourfold arrangement of the rock units and assigned them formation rank. These were, in ascending order, the Chimneyhill Limestone, Henryhouse Shale, Haragan Shale, and Bois d'Arc Limestone.

Further stratigraphic and paleontologic work was done by Amsden (1957) and Shannon (1962). Shannon stated that the Hunton Group was "an essentially conformable succession of strata that represents a sequence, or body of rock bounded by time-transgressive interregional unconformities." Subsurface investigations made by Shannon indicate that the physical relationships of formations within the Hunton Group illustrate continuous sedimentation within the unit from Middle Silurian through Early Devonian. Shannon's subsurface work strengthened Reeds' four-unit stratigraphic divisions and provided recognition of the units on wire-line logs (Menke, 1986).

Amsden extensively discussed the existence of unconformities in the Hunton (Amsden, 1960, and subsequent publications) and established the stratigraphic and paleontological framework that is recognized today (Fig. 1).

The Hunton Group overlies the Sylvan Shale. The basal Chimneyhill Subgroup is a medium- to coarse-crystalline limestone that may be further subdivided into three formations: the Keel, Cochrane, and Clarita Formations. The Henryhouse Formation unconformably overlies the Chimneyhill, and is separated from the overlying Haragan Formation by lithologic and paleontological differences (Amsden, 1957,1960,1975,1980). The Bois d'Arc Formation overlies the Haragan and commonly is separated from the Haragan by the cherty beds in its base. The uppermost unit in the Hunton Group is the Frisco Formation. The Frisco was deposited on the unconformable surface that developed as a result of Early Devonian epeirogenies (Amsden, 1975).

Investigations conducted by Beardall (1983), Medlock (1984), Manni (1985), and Menke (1986) addressed specific Hunton stratigraphic units and established depositional models for the reservoir facies. Morgan (1985) showed the importance of oolitic facies and certain aspects of reservoir development. Matthews (1992) illustrated the importance of karstic processes on Hunton reservoir evolution. Other important studies that addressed various aspects of Hunton geology include Morgan (1922), Ballard (1930), Posey (1932), Decker (1935), Maxwell (1936), Anderson (1939), Swesnik (1948), Tarr (1955), Oxley (1958), Maxwell (1959), England (1965), Kunsman (1967), Harvey (1969), Isom (1973), Hollrah (1977), Borak (1978), and Throckmorton and Al-Shaieb (1986).

REGIONAL GEOLOGIC HISTORY AND STRUCTURAL SETTING

The Anadarko basin is highly asymmetric, with a steep faulted southern margin and a gently sloping northern shelf. The basin is bounded on the east by the Nemaha uplift and the west by the Cimarron arch.

The Anadarko basin is a part of what was described as the southern Oklahoma aulacogen by Shatski (1946), Burke and Dewey (1973), and Hoffman and others (1974). During Precambrian and Early and Middle Cambrian time, intrusive and extrusive igneous rocks were emplaced in the failed rift. These rocks consist of Cambrian-age granites, rhyolites, gabbros, anorthosites, and basalts that intruded and covered Precambrian granitic rocks. From Late Cambrian through the Ordovician, carbonates dominated deposition in the aulacogen. Subsidence was relatively rapid and a thick sequence (>6,000 ft along the depocenter) of shallow-water carbonates accumulated (Johnson, 1989). The depositional cycles recorded by these rocks reflect sea-level fluctuations in a broad epicontinental sea.

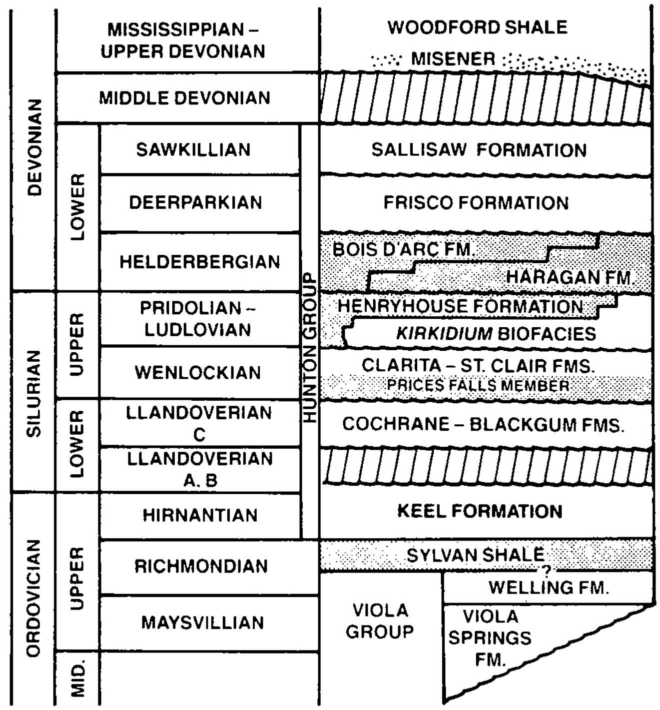


Figure 1. Stratigraphic nomenclature of the Hunton Group in Oklahoma. Shaded boxes indicate formations that contain moderate to large amounts of fine silt- and clay-sized terrigenous detritus (Amsden, 1989).

Middle Ordovician through earliest Mississippian (Kinderhookian) sediments consist of shallow-marine carbonates that are interbedded with sandstones and shales; they are moderately thick in the southern Oklahoma aulacogen (Fig. 2). The Ordovician-age Simpson Group sandstones and carbonates are overlain successively by the Viola Group limestones and the Sylvan Shale.

The Late Ordovician–Devonian-age Hunton Group is a series of carbonates that was deposited in the epicontinental sea that covered the southern Midcontinent region. In contrast to some of the lower Paleozoic carbonates, the Hunton represents a time of relative slow subsidence. As a result of local epeirogeny and sea-level fluctuation, several disconformities developed within the Hunton Group. Two major epeirogenies interrupted deposition during the Devonian (Amsden, 1975). The Early Devonian (pre-Frisco) and pre-Woodford uplifts generated unconformities that affected the entire region. Erosion associated with this later episode truncated the sequence and removed it from much of the northeast Oklahoma platform. During this hiatus,

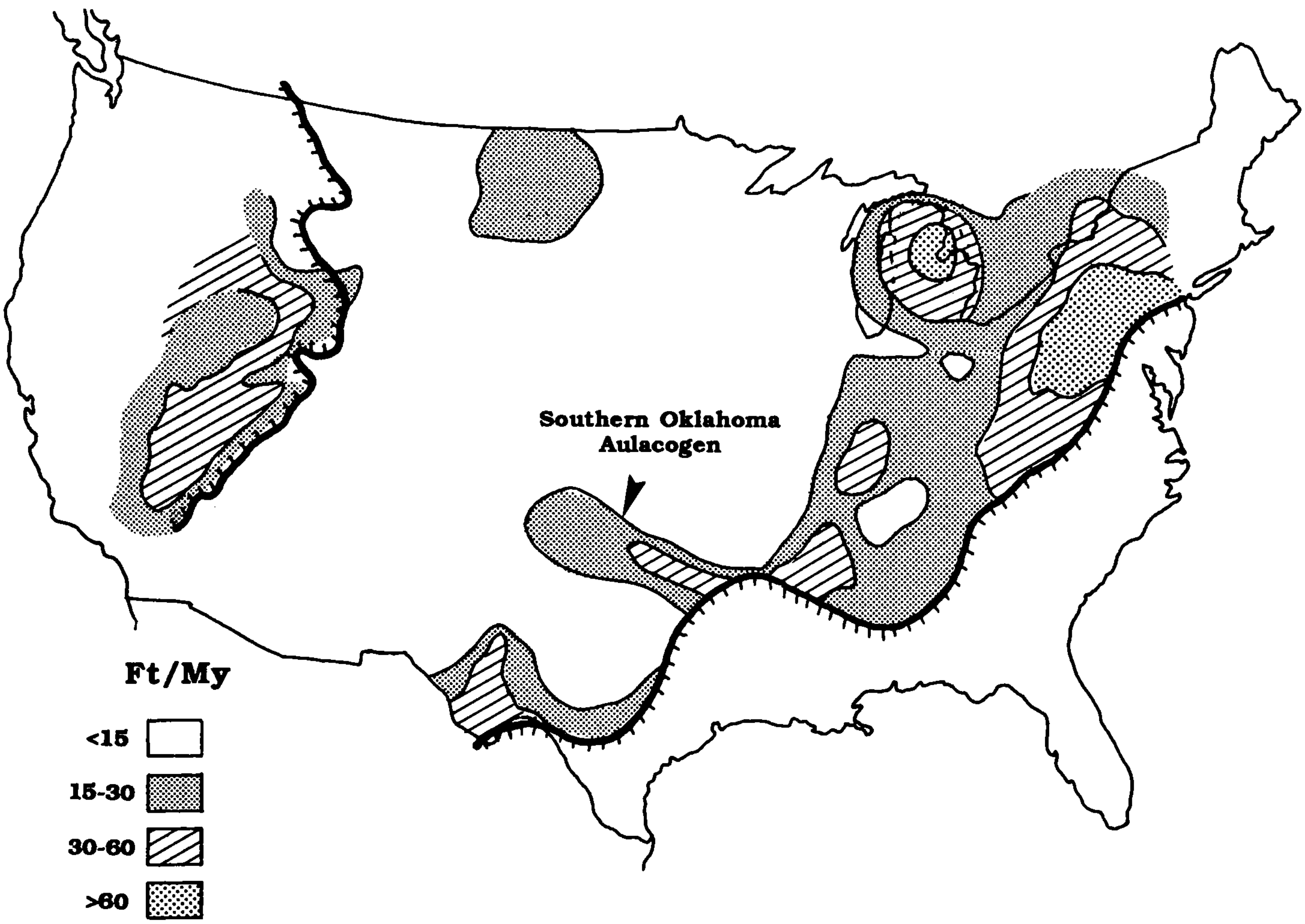


Figure 2. Subsidence rates during Middle Ordovician to Early Devonian (including the Chimneyhill Subgroup and the Henryhouse Formation of the Hunton Group). Maximum Silurian sediment accumulation in the Midcontinent region occurred during the subsidence stage of the southern Oklahoma aulacogen (after Sloss, 1988).

stream erosion superimposed a westerly flowing dendritic drainage pattern on the Hunton topography (Amsden, 1975; Beardall, 1983). Karst development greatly enhanced porosity in the Hunton rocks (Matthews and Al-Shaieb, in press). This erosional surface was subsequently buried during transgressive deposition of the Late Devonian–Mississippian Misener sandstone and Woodford Shale.

Mississippian sediments that succeeded the Woodford Shale consist mainly of shallow-marine limestones and cherty limestones, and relatively minor amounts of sandstones and shales. Shale and sandstone compose a large part of the Mississippian (Chesterian) strata in the southern part of the present Anadarko basin.

During the Pennsylvanian, major changes affected the region and the present basin configuration was developed. Prior to the Early Pennsylvanian Wichita orogeny, the entire region subsided and was uplifted as a unit. With initiation of the Pennsylvanian orogenic episodes, the region was subdivided into the tectonic provinces recognized today.

The Anadarko basin subsided rapidly throughout the Pennsylvanian orogeny. As much as 18,000 ft of Pennsylvanian clastics and carbonates were deposited in the basin. During this period of rapid subsidence, the Hunton Group was buried to depths exceeding 20,000 ft near the basin axis (Fig. 3). Woodford Shale source rocks adjacent to the Hunton matured and hydrocarbons migrated into the Hunton porosity network. Most petroleum accumulated on anticlinal structures or in stratigraphic traps influenced by the distribution of porous facies. Even with extreme burial and heating, the fluids in these reservoirs generally were not overpressured. It is believed that the extensive porosity network within the Hunton allowed pressure dissipation or venting of pressure to the outcrop.

Following the Pennsylvanian orogeny, Hunton rocks in the Anadarko basin underwent gradual burial until the Cretaceous Laramide orogeny (Fig. 3). Since Cretaceous time, the entire Midcontinent region has been raised through epeirogenic uplift (Schmoker, 1986).

DEPOSITIONAL FACIES AND ENVIRONMENTS

Introduction

During Silurian time, the developing Anadarko basin (Fig. 3) subsided at an extremely slow rate (Adler, 1971; Feinstein, 1981). Under these stable tectonic conditions, the Henryhouse Formation was deposited over much of Okla-

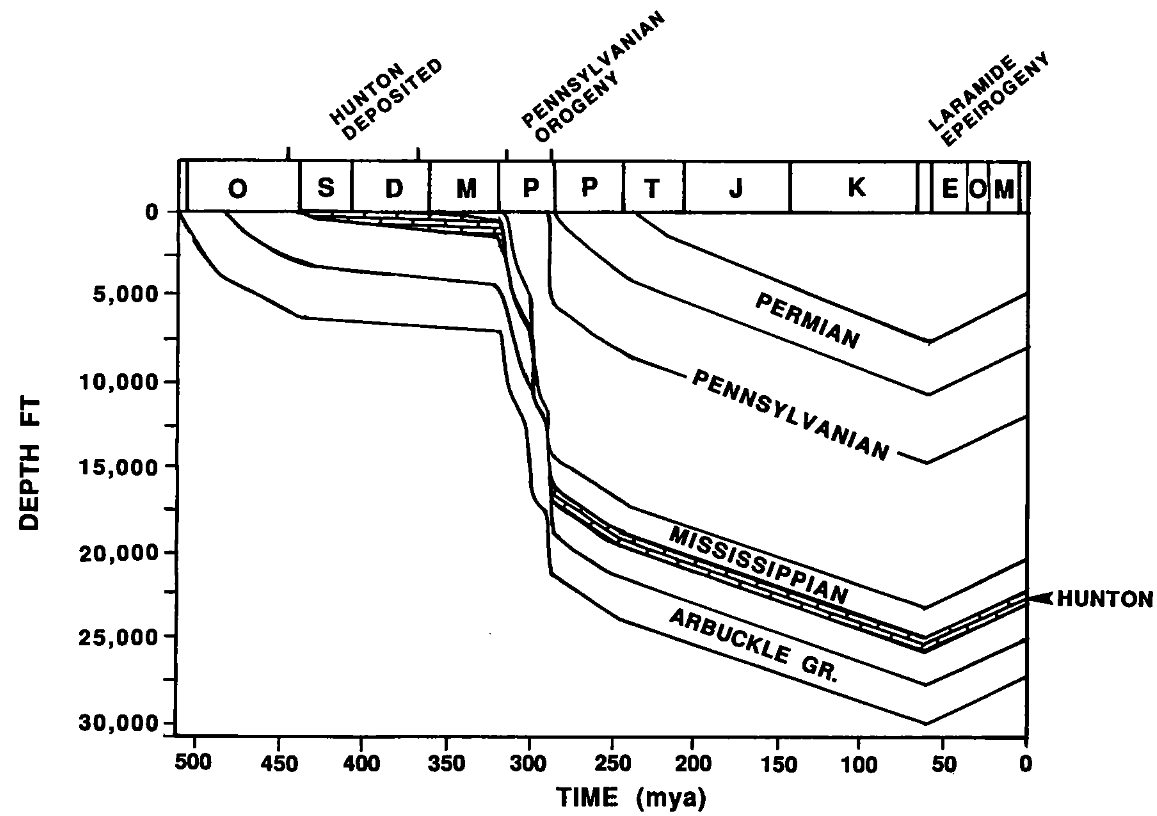
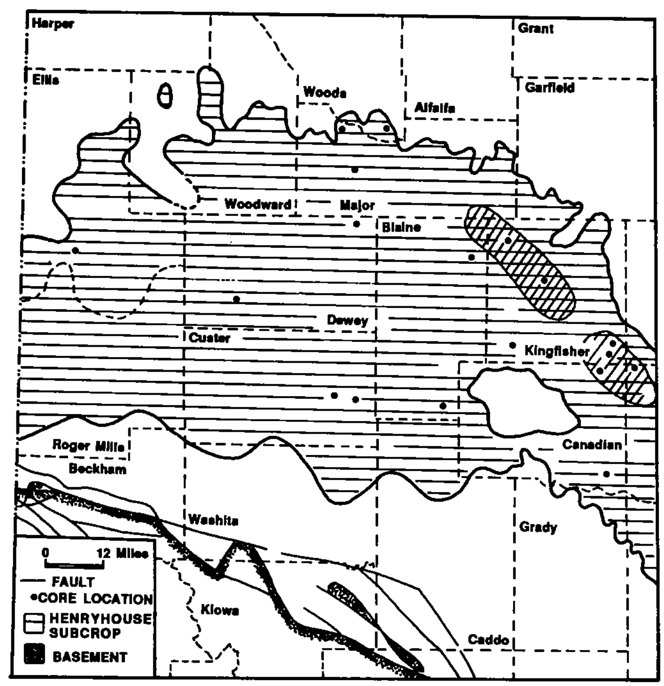


Figure 3. Burial-history curve for the Reydon field area, Roger Mills County, Anadarko basin. Note the slow subsidence during Hunton deposition, rapid subsidence and deep burial of the Hunton during the Pennsylvanian orogeny, continued slow burial during the Permian and Mesozoic, and uplift to the present depth due to the Laramide epeirogeny.

homa with no major unconformities disturbing deposition (Shannon, 1962) (Fig. 4). In this shallow and low-energy sea, several facies belts developed subparallel to bathymetric contours (Fig. 5). Transitions between facies commonly were gradational and in some places quite subtle, but where appropriate data are available, facies can be distinguished and subdivided. Facies classification is based on observable vertical successions of facies or sequences of aggradation. Transgression and regression seemingly caused extensive migration of facies, thereby producing similarity of log signatures across large areas of the basin. Diagenetic influence (compaction, dolomitization, and dissolution) has somewhat obscured the original sedimentary structures and fossil content of the Henryhouse Formation. Despite this influence, sufficient evidence still remains to recognize three distinct depositional facies (Fig. 6). Core-slab, thin-section, and X-ray-diffraction data are used to characterize each facies. Environmental interpretations of the following facies were made by using the criteria shown in Figure 7 (Flugel, 1982).

Facies I

This facies is found in the supratidal tidal flat, deposited very near, or above, mean high tide. Cryptal algal fabrics, fenestral fabrics (Fig. 8), irregular laminations, lack of fossils, and a lack of burrow mottling are the salient features of this facies. Silica nodules, silt-sized quartz, intraclasts, peloids, and low-amplitude stylolites may all be present in



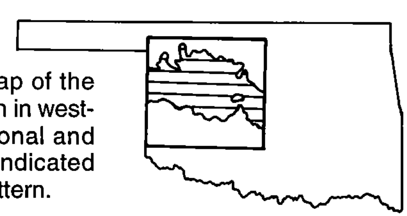


Figure 4. Subcrop map of the Henryhouse Formation in western Oklahoma. Lagoonal and oolitic subfacies are indicated by cross-hachured pattern.

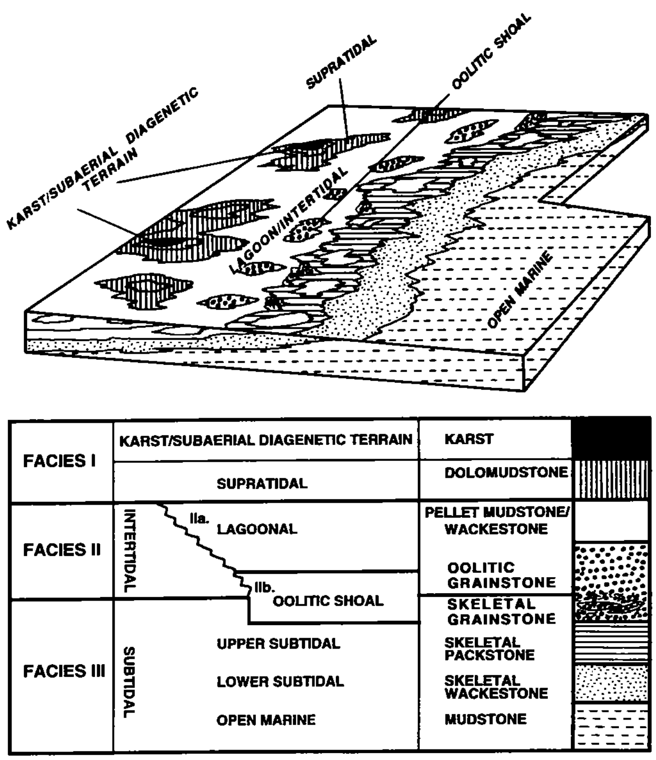


Figure 5. Schematic diagram showing a generalized model of the depositional environments and facies of the Henryhouse Formation.

what is typically a light-colored dolo-mudstone with no porosity. A shallow, restricted shoreline environment is indicated particularly by the lack of fossils and nonburrowed irregular laminations.

Facies II

This facies is encountered in a shallow, restricted subtidal to upper intertidal environment, deposited within or just below normal high tide and normal low tide. Burrow mottling makes this facies visually distinct (Fig. 9). All samples of this facies are fully dolomitized with good to excellent porosity and are predominantly dolo-wackestone. Fossil percent and type were usually estimated, due to dissolution of the original grains. Molds and occasionally remaining grains indicate that crinoids were the most common fossil. Intraclasts, silica and carbonate nodules, and silt-sized quartz are likely to be present; stylolites are rare and generally high amplitude. The highly burrowed appearance and accumulation of rather sorted crinoid fragments suggests shallow, but more seaward, conditions with increased water agitation in or adjacent to the zone of wave action.

A zone of oolitic grainstone was observed within the Henryhouse Formation in central Oklahoma (Blaine and Oklahoma Counties, T. 19 N., R. 10 W., and T. 13 N., R. 4 W., respectively; Fig. 4). This facies can be further subdivided into two subfacies: an intertidal lagoonal (IIa) and an oolitic (IIb) subfacies (Fig. 10).

Subfacies a (Lagoonal)

The lagoonal subfacies may have been deposited under restricted conditions since fossils are rare, but burrowing is present. The rock is a massive peloidal dolo-mudstone (Figs. 11,12). In the northwest and the southeast, the texture of the lagoonal subfacies deposits was severely altered by dolomitization. The lagoonal subfacies was formed in quiet water, landward of the oolitic build up.

Subfacies b (Oolitic)

This subfacies contains abundant ooliths and, locally, fossil fragments and peloids. Crossbedding and horizontal laminations are common in the calcitic oolites (Fig. 13). Dolomitization obliterates the textures in the dolomitic oolites (Fig. 14). The rock is a grainstone. The shoal probably was developed in the lower intertidal/upper subtidal environment.

Morgan (1985) mapped the extent of the calcitic-oolite subfacies in Oklahoma County, where it thins and disappears to the northwest. He interpreted this pinch-out as being due to a tidal channel, post-Henryhouse erosion, or the absence of syndepositional facies.

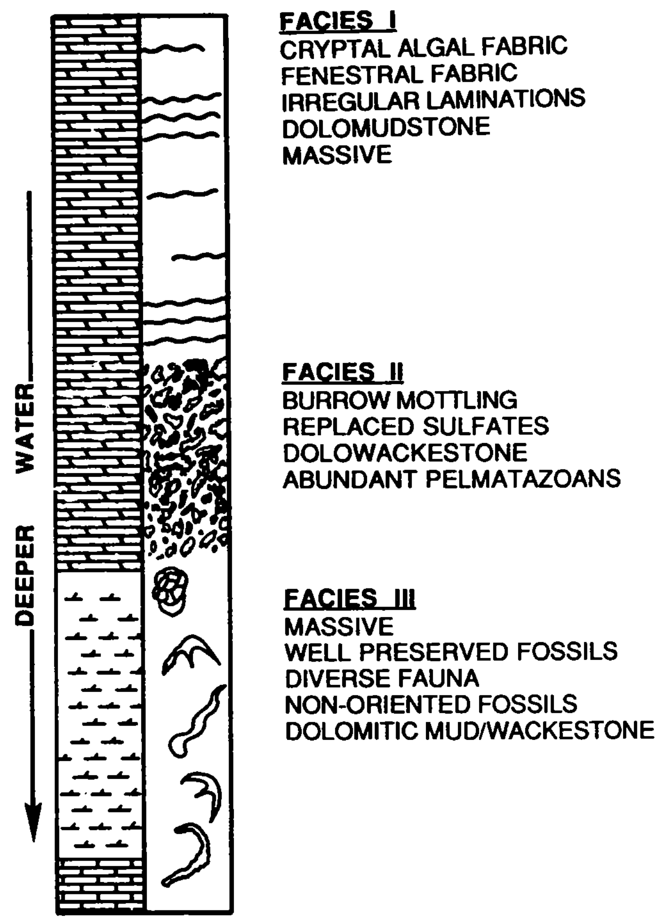


Figure 6. Typical vertical facies sequence showing sedimentary structures, along with faunal and mineralogic relationships observed in cores and thin sections from the Henryhouse Formation.

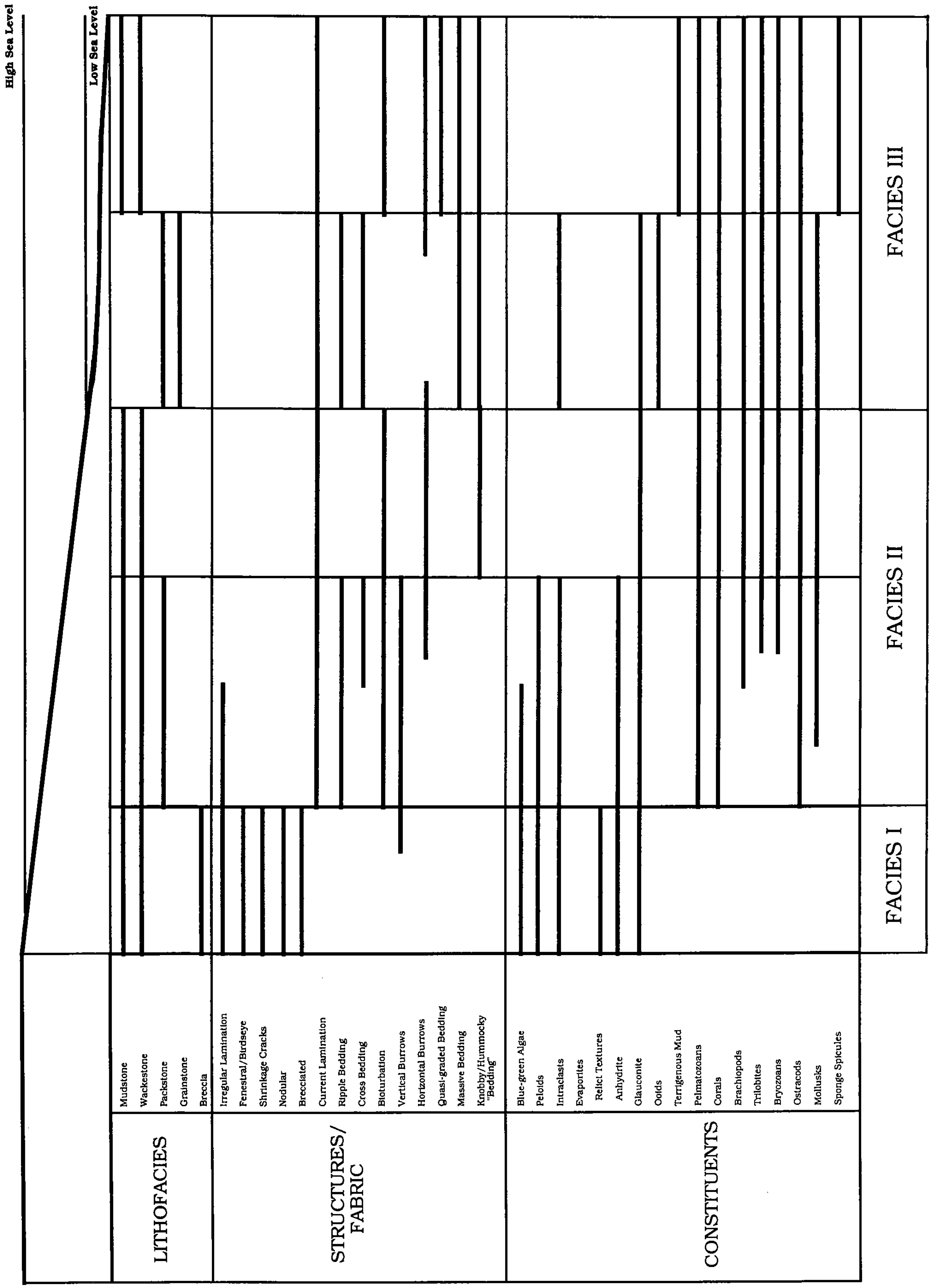


Figure 7. Criteria for recognition and classification of Hunton depositional environments and facies.

Facies III

This facies is strictly found in an open-shelf subtidal environment and, when continually submerged, it is typically a medium-dark, silty, dolomitic mudstone. Fossils include brachiopods, with less-common trilobites, ostracods, bryozoans, and echinoderms (Fig. 15). This facies often is featureless, but it may exhibit burrow mottling, nodular bedding, or storm-type deposits. The more-diverse fauna and delicate, unsorted, well-preserved fossils indicate a lower-energy, shallow open-shelf environment below normal wave base.

DOLOMITIZATION

Introduction

Reservoir rocks of the Hunton Group, and specifically of the Henryhouse Formation, are mostly dolomitized. Amsden (1975) interpreted dolomite of the Silurian Hunton rocks as having been of the penecontemporaneous replacement type, formed in marine environments seaward of the tidal zone. However, Beardall and Al-Shaieb (1984) reported three distinct types of dolomite, which are related to three different processes of dolomitization (Fig. 16).

Hypersaline Dolomite

During deposition of the Henryhouse, supratidal conditions existed in a restricted inner-platform shelf area. Anhydrite in the Henryhouse suggests that evaporation from the supratidal area was of sufficient intensity and duration to permit formation of hypersaline brines (Fig. 17). Seaward of the supratidal area, normal-marine conditions were conducive to organisms that thoroughly burrowed the sediments of the intertidal/shallow subtidal area. The anhydrite (Fig. 18) indicates that dolomitizing brines moved downward and seaward from the supratidal zone. A consequence of the precipitation of $CaSO_4$ is an increased Mg^{2+}/Ca^{2+} ratio, which is of major significance for the formation of dolomite. Fluids with high Mg^{2+}/Ca^{2+} ratios probably enhanced dolomitization of the intertidal/shallow subtidal facies. General impermeability of the deeper subtidal facies explains its incomplete dolomitization.

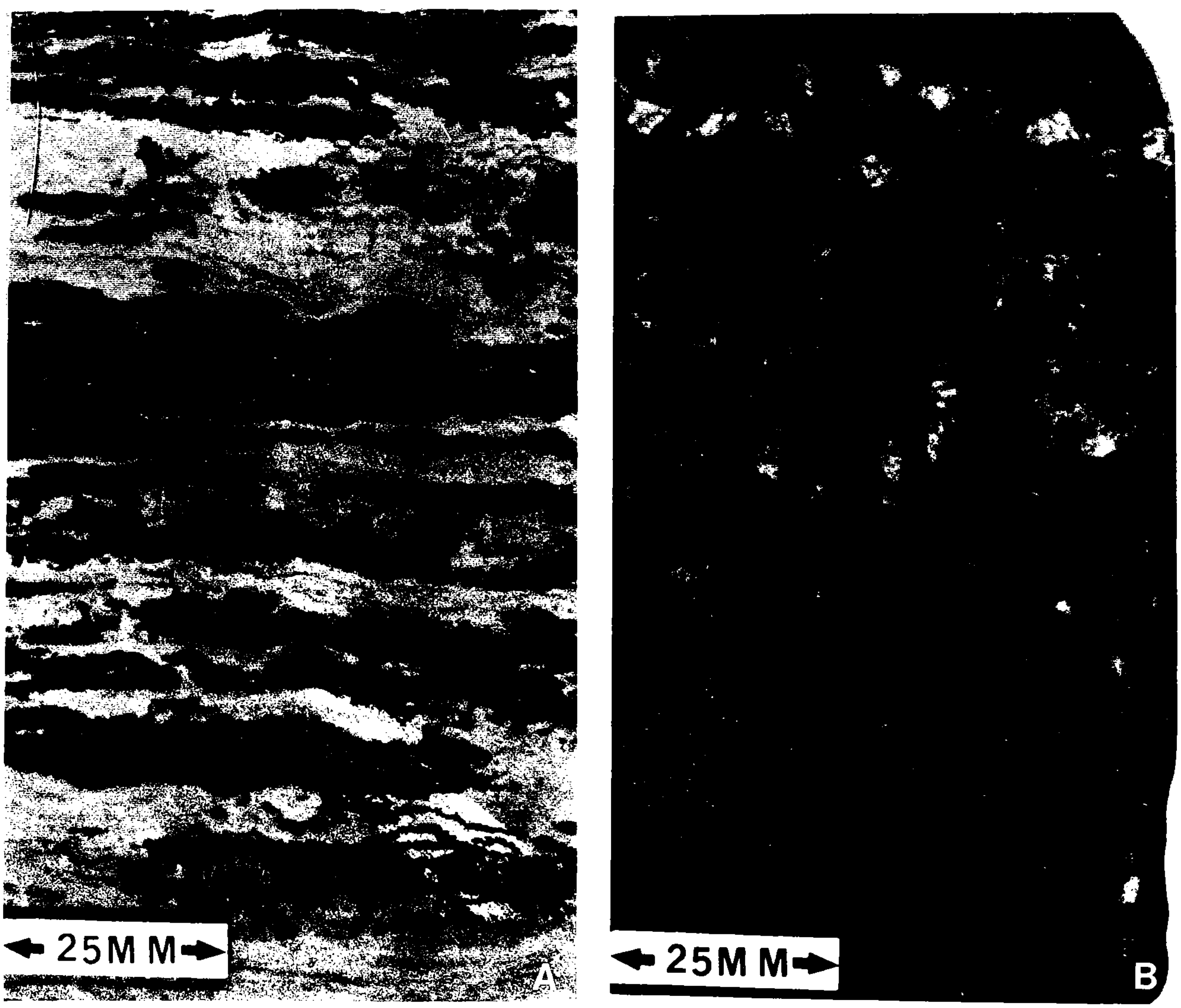


Figure 8. Supratidal facies (I). *A*—Dolo-mudstone with cryptalgal fabric. *B*—Fenestral fabric in peloidal wackestone.

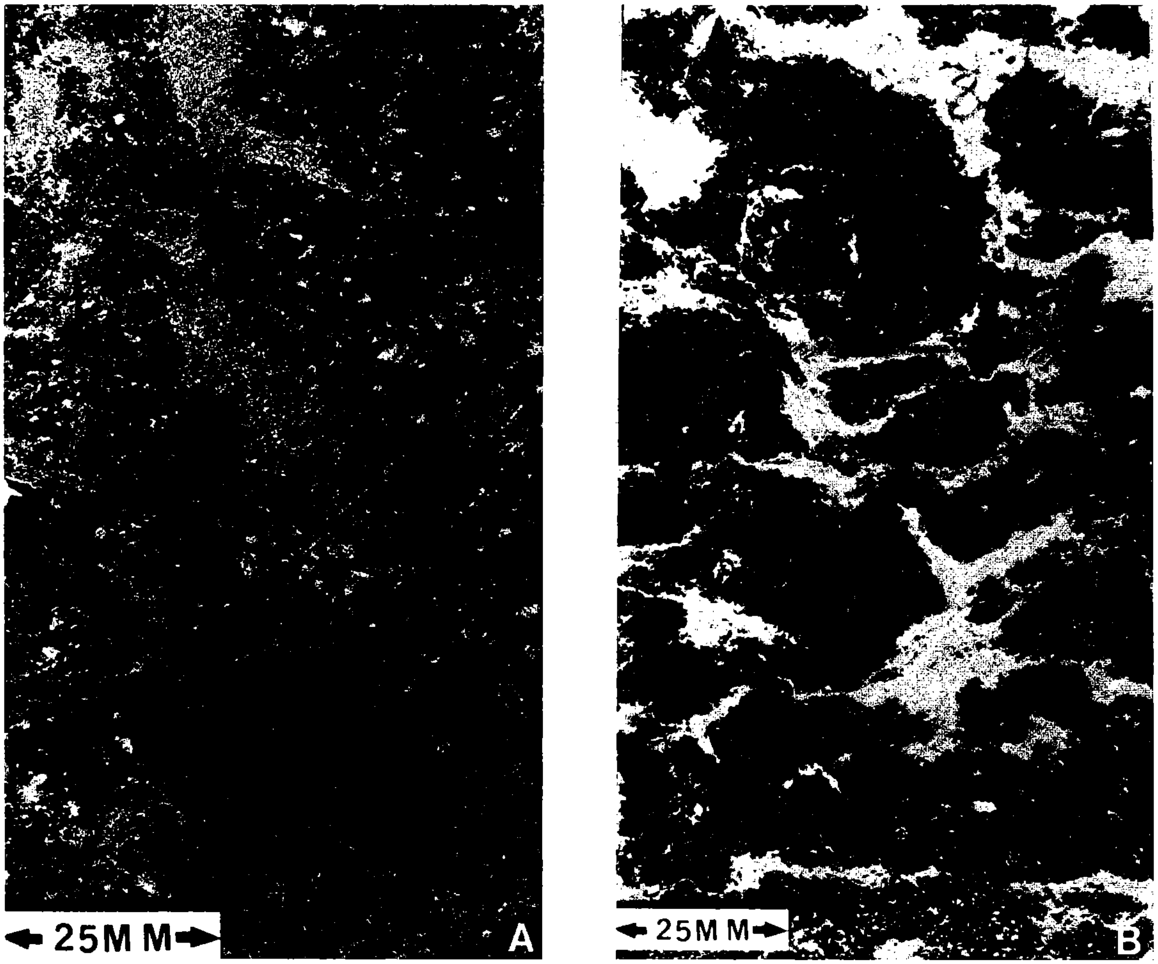


Figure 9. Intertidal facies (II). *A*—Burrowed, bioturbated, and oil-stained dolo-wackestone with excellent porosity. *B*—Algal laminae disturbed by burrowing.

Mixed-Water Dolomite

During deposition of the Henryhouse, regressive, progradational deposition of carbonate sediment eventually should have allowed meteoric water to migrate basinward through the sediments. As the shoreline advanced across the wide shelf, a migrating, brackish-water zone of mixing should have developed in the transition zone between fresh ground water and ocean-derived brines. Such waters could have been saturated with respect to dolomite (Handshaw and others, 1971), and could have allowed precipitation of younger dolomite, either on existent crystals of dolomite or as discrete rhombohedra in pores.

A mixed-water mechanism is supported by exceedingly abundant white or clear overgrowths of dolomite and their near-uniform distribution across facies. Cathodoluminescent zonation (Fig. 19) of dolomite from the Henryhouse is regionally similar, which suggests a common diagenetic history (Choquette and Steinen, 1980). Mildly reducing conditions, indicated by the brightly luminescent outermost rim of Henryhouse dolomite, are consistent with mixed-water dolomitization.

Deposition of the Hunton was preceded and followed by a hiatus (Amsden, 1975). During periods of nondeposition, a system of recharge and hydrodynamics may have developed, sufficient for formation of dolomite by mixing of marine water and fresh water. Concentration of dolomite associated with tectonic and erosional features was reported by Isom (1973), who interpreted dolomitization as probably having been associated with the movement of high-magnesium waters along fractures and unconformities. Lenses of fresh water, developed under exposed terrains of carbonate rock, could have mixed with interstitial sea water, the source of Mg^{2+}. Mixing of fresh water and connate sea water is considered the most likely mechanism for paleotopographic dolomitization of the Henryhouse.

Deep-Burial Dolomite

The final stage of dolomitization in the Henryhouse is recorded by white, cloudy, void-filling saddle dolomite (Fig. 20). The distorted crystal lattice and curved crystal faces of saddle dolomite are associated with curved cleavage traces and sweeping extinction. Under cross-polarized light, saddle dolomite of the Henryhouse is generally black and white; some samples are pearly, but the rock invariably shows diagnostic sweeping extinction. A richly ferroan composition is indicated by very dull or dark cathodoluminescence and by conventional staining with potassium ferricyanide.

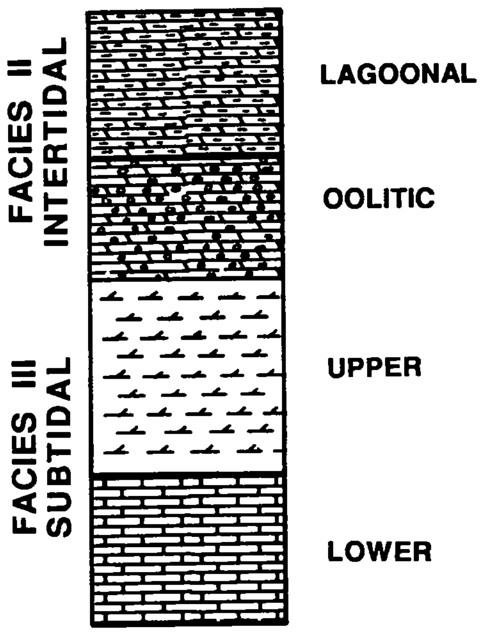


Figure 10. Typical vertical sequence of the Henryhouse Formation showing lagoonal (IIa) and oolitic (IIb) subfacies of facies II.

Saddle dolomite was recorded only in dolomitized zones of the Henryhouse, where crystal lengths are generally recorded in millimeters. These crystals predominantly fill secondary vugs and fractures, but rarely fill molds of fossils. The tendency for saddle dolomite to be the dominant or sole epigenetic carbonate mineral in dolostones was reported by Choquette and Pray (1971). This phenomenon is believed to be due to a local source of Mg^{2+} and Fe^{2+} (Choquette and Pray, 1971). Saddle dolomite generally appears to be a passively precipitated, void-filling cement. Under cathodoluminescence, saddle dolomite exhibits subtle zonation comprised of dull-red, bright-red, yellow and nonluminescent laminae, attesting to the slight variance in chemical composition across each layer (Lynch and Al-Shaieb, 1991). Inclusions are commonly present, but not in a patterned arrangement that would suggest replacement of a precursor. The fracture-filling saddle dolomite is a relatively deep-burial, late-diagenetic mineral. Radke and Mathis (1980) postulate that it forms at temperatures >80°C, and thus would imply a deep-burial origin or a hydrothermal origin at shallower depths.

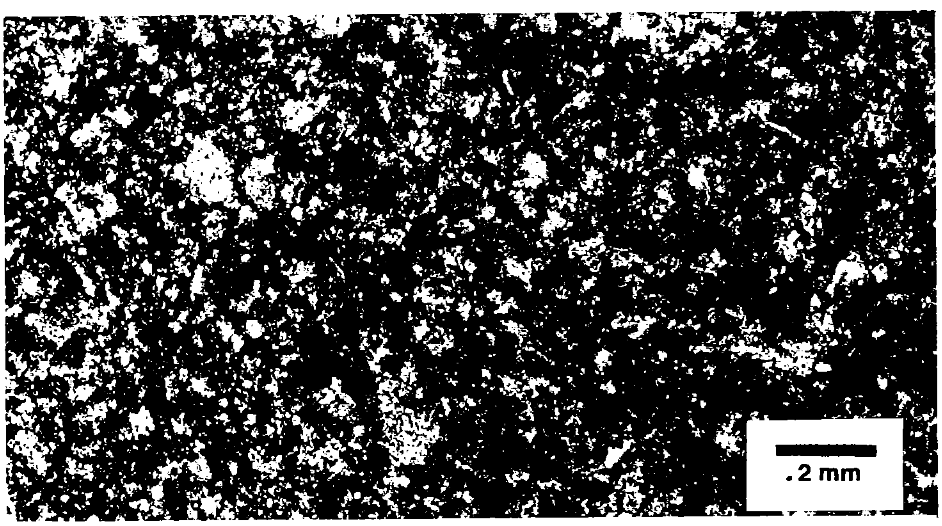


Figure 11. Lagoonal subfacies (IIa) with characteristic peloids. (Eason Van Curen well, 7,140 ft.)

Figure 12. Lagoonal subfacies (IIa) with characteristic echinoid fragment. (Eason Van Curen well, 7,143 ft.)

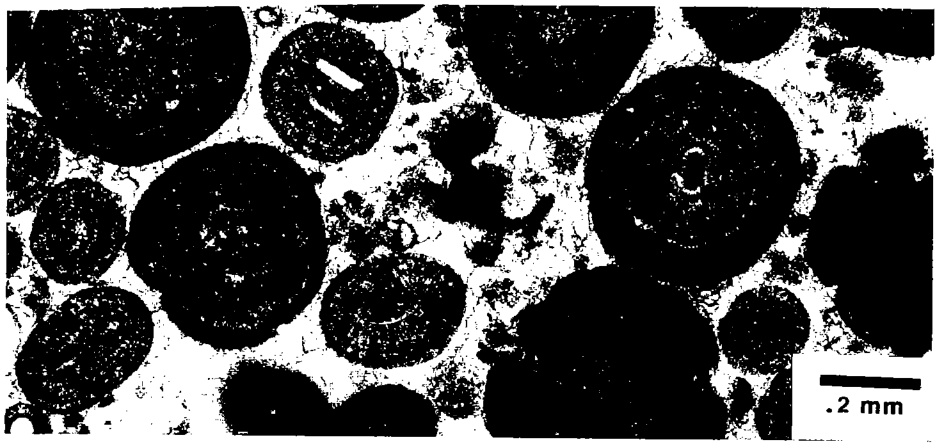


Figure 13. Oolitic subfacies (IIb) with isopachous calcite cement filling interooidal space. (Kirkpatrick Cronkite well, 7,119 ft.)

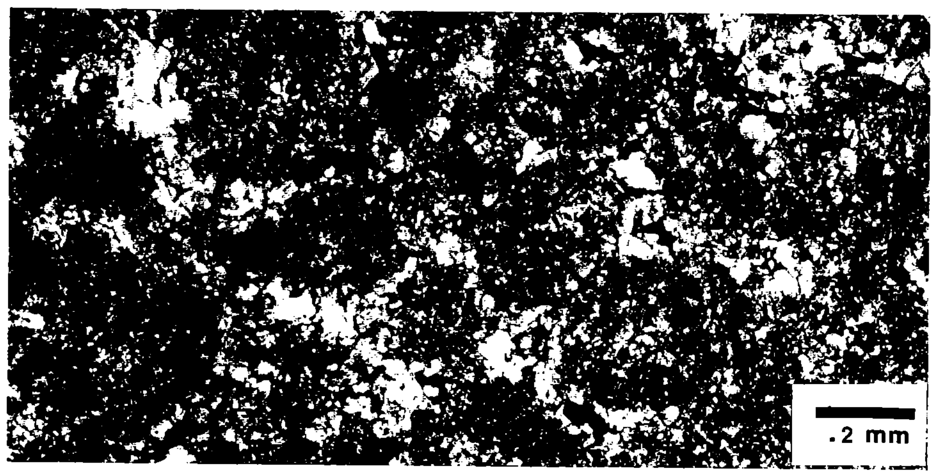


Figure 14. Photomicrograph of dolomitized oolite. Ooids appear cloudy with clear rims. (Duncan Garrett well, 8,751 ft; Medlock, 1984.)

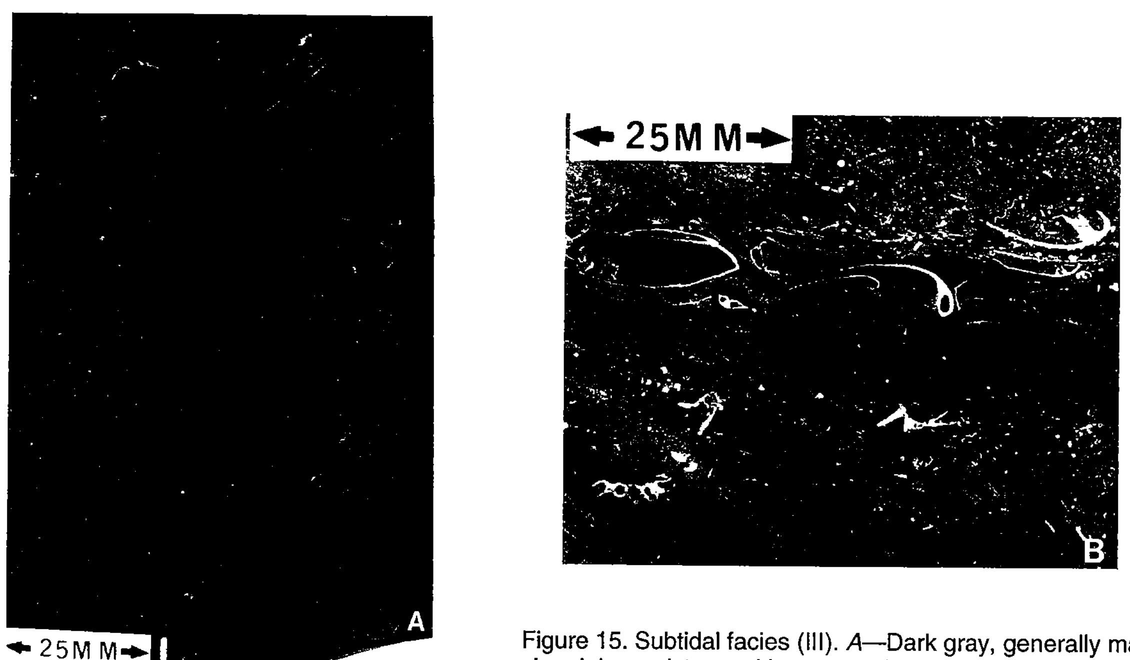


Figure 15. Subtidal facies (III). *A*—Dark gray, generally massive dolo-mudstone with scattered brachiopods. *B*—Clayey dolo-wackestone with thin-shelled fossils, including brachiopods, bryozoans, and pelmatozoans.

POROSITY

Four types of porosity are recognized in Hunton rocks. They belong to both general porosity classes (fabric selective and nonfabric selective) described by Choquette and Pray (1971).

Two fabric-selective porosity types are observed in the Henryhouse: (1) moldic, and (2) intercrystalline. These types are responsible for most high-porosity zones in the cores. Moldic porosity is caused by dissolution of fossil grains, predominantly crinoid and mollusk fragments. Occasionally, partially dissolved faunal remnants are found within the molds. Molds subsequently may be filled by late baroque dolomite or calcite. Intercrystalline pore space evolved primarily as a consequence of dissolution of nondolomitized, cryptocrystalline calcite matrix. Important solution enlargement of this porosity type is common. Intercrystalline porosity is likely to develop between larger rhombs in an idiotopic to hypidiotopic texture. Late calcite cement may occlude some intercrystalline pores.

Porosity is best developed in the intertidal facies (facies II), especially in the bioturbated and burrowed wackestone (Fig. 21). Wackestone is dolomitized completely at some localities. A positive relationship between dolomitization and porosity is clearly seen (Fig. 22). Highest and lowest values of porosity correspond to highest and lowest values of dolomitization. Further consideration reveals that this relationship is somewhat deceptive. When porosity is plot-

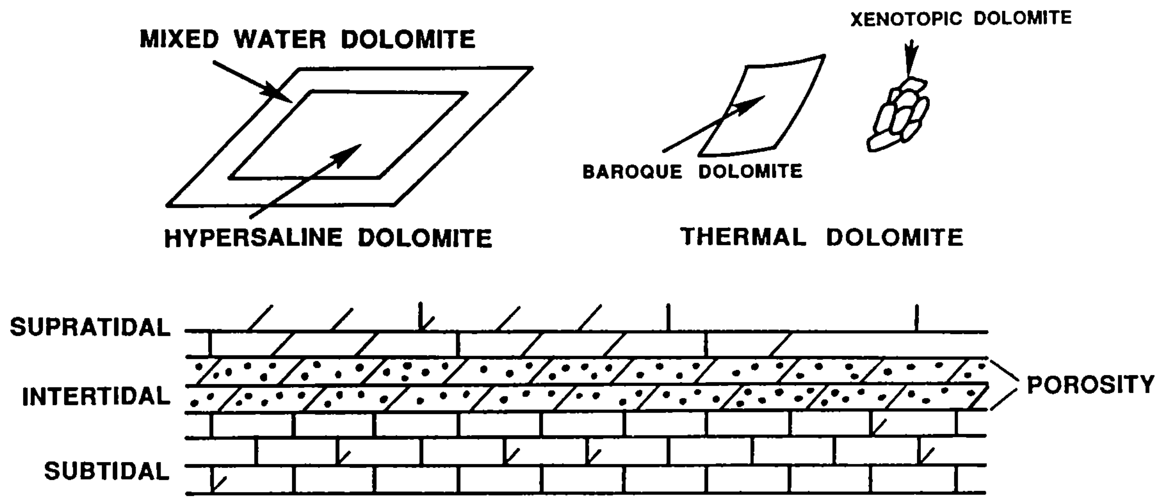


Figure 16. Schematic diagram illustrating the three distinct types of dolomite found in the Henryhouse Formation. Hypersaline dolomite formed penecontemporaneously in the intertidal/supratidal environment. Zoned hypersaline/mixed-water dolomite rhombs indicate an influx of fresh water due to epeirogeny. Thermal dolomite indicates deep burial or hot basinal-fluid migration.

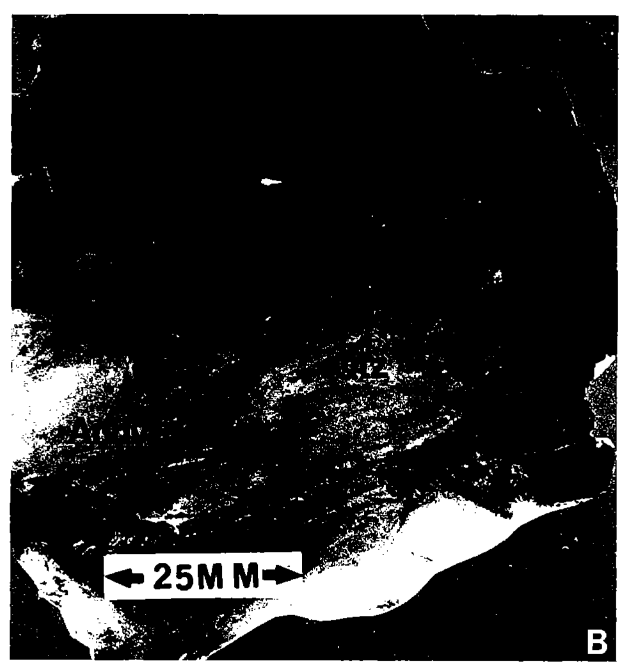


Figure 17. *A*—Hypersaline dolomite characterized by poorly formed rhombohedra, which appear cloudy and small in size. *B*—Anhydrite associated with hypersaline dolomite.

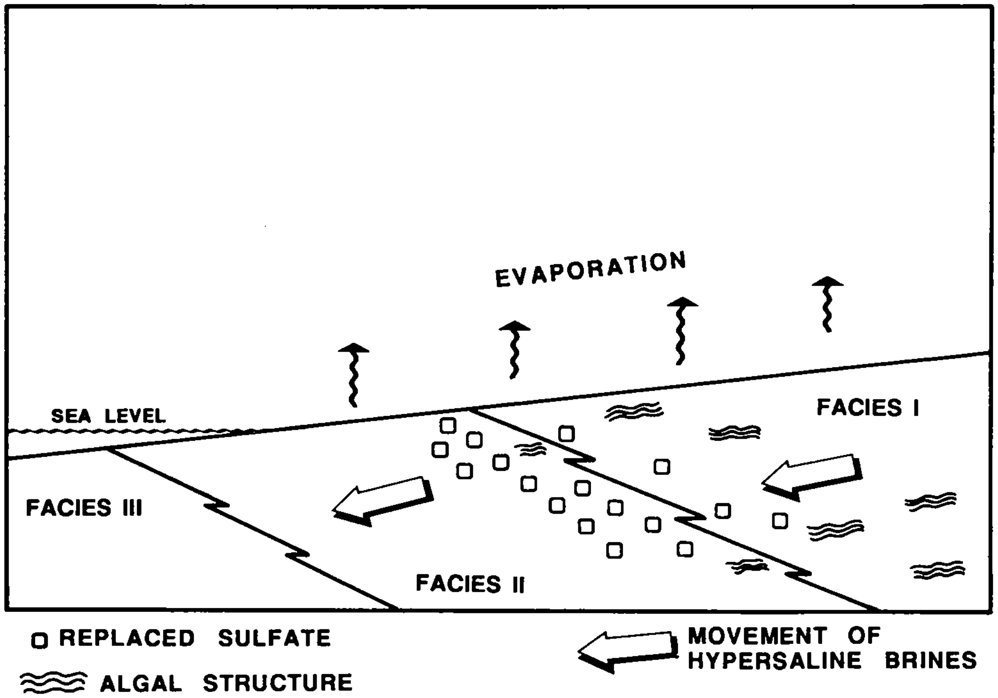


Figure 18. Schematic diagram illustrating movement of hypersaline brines and their role in dolomitization of the Hunton carbonates during regression.

ted only for zones with high values of dolomite (>90%), the dominating influence of depositional facies becomes apparent. If both facies I and facies II are completely dolomitized, only facies II will have significant porosity. Facies II was never observed to be either completely dolomitized or porous (Beardall, 1983).

The characteristics of facies II that cause it to be preferentially porous are burrowing and common pelmatozoan debris. Burrowing has enhanced permeability by redistributing the finer particles, allowing subsequent dissolution of nondolomitized matrix and grains by percolating low-pH solutions. The resultant patchy, irregular porosity distribution is very evident in cores and thin sections.

In the absence of grains (fossils), moldic porosity does not develop. Both facies II and facies III are fossiliferous, yet only the highly dolomitized facies II develops moldic porosity. Porosity in oolitic grainstones is predominantly interooid (Fig. 23). This type of porosity also shows good correlation with the degree of dolomitization. In nondolomitized oolitic rocks, porosity generally is poorly developed because of early cementation by sparry calcite.

The most important nonfabric selective porosity is dissolution porosity. Commonly, vugular pores are solution-enlarged molds where the original fossil outline has been destroyed. Coincidence of vugular porosity with fossiliferous, moldic zones supports this interpretation. Dissolution of fragmented fossils within the less densely burrowed zones was the most important mechanism in development of moldic (Fig. 24) or vuggy karstic porosity. Several cores examined for this study exhibit outstanding specimens of dissolution vugs, vugs partially or completely cemented, and vugs containing geopetal structures. Some large vugs apparently were created by the dissolution of incompletely replaced anhydrite nodules (Beardall, 1983). In rare cases, partially dissolved anhydrite is present and surrounded by replacive euhedral quartz.

Solution-enlarged fractures, joints, and channels play a significant role in karst-related porosity in the Hunton rocks. Figure 25 shows small-scale solution-enlarged fracture porosity, where the interstices have been filled geopetally with laminated cave muds, glauconite, and dolomite cement. Cavern or cave porosity is another form of karstic porosity that is evident in the Hunton rocks (Matthews and Al-Shaieb, in press).

Supratidal and distal subtidal facies tend to have low porosity. Rocks of the intertidal facies, and locally of the upper subtidal facies, have maximal probability of porosity adequate for commercial oil and gas reservoirs. Karstic processes generally enhance overall porosity in Hunton rocks, but are not required for porosity development.

CONCLUSIONS

1. The Hunton (Henryhouse) depositional model consists of three broad depositional facies: supratidal/tidal flat (facies I), intertidal (facies II), and subtidal (facies III).

2. Three main species of dolomite are observed: (1) brownish, cloudy, hypidiotopic dolomite that formed shortly after deposition from hypersaline brines; (2) white or limpid idiotopic dolomite that formed after the hypersaline dolomite due to freshwater/marine-water mixing; and (3) postcompactional mold-, vug-, and fracture-filling baroque or saddle dolomite.

3. Dolomite type is related to facies: the lagoonal subfacies (intertidal) developed more hypersaline dolomite, the subtidal facies developed more freshwater/marine-

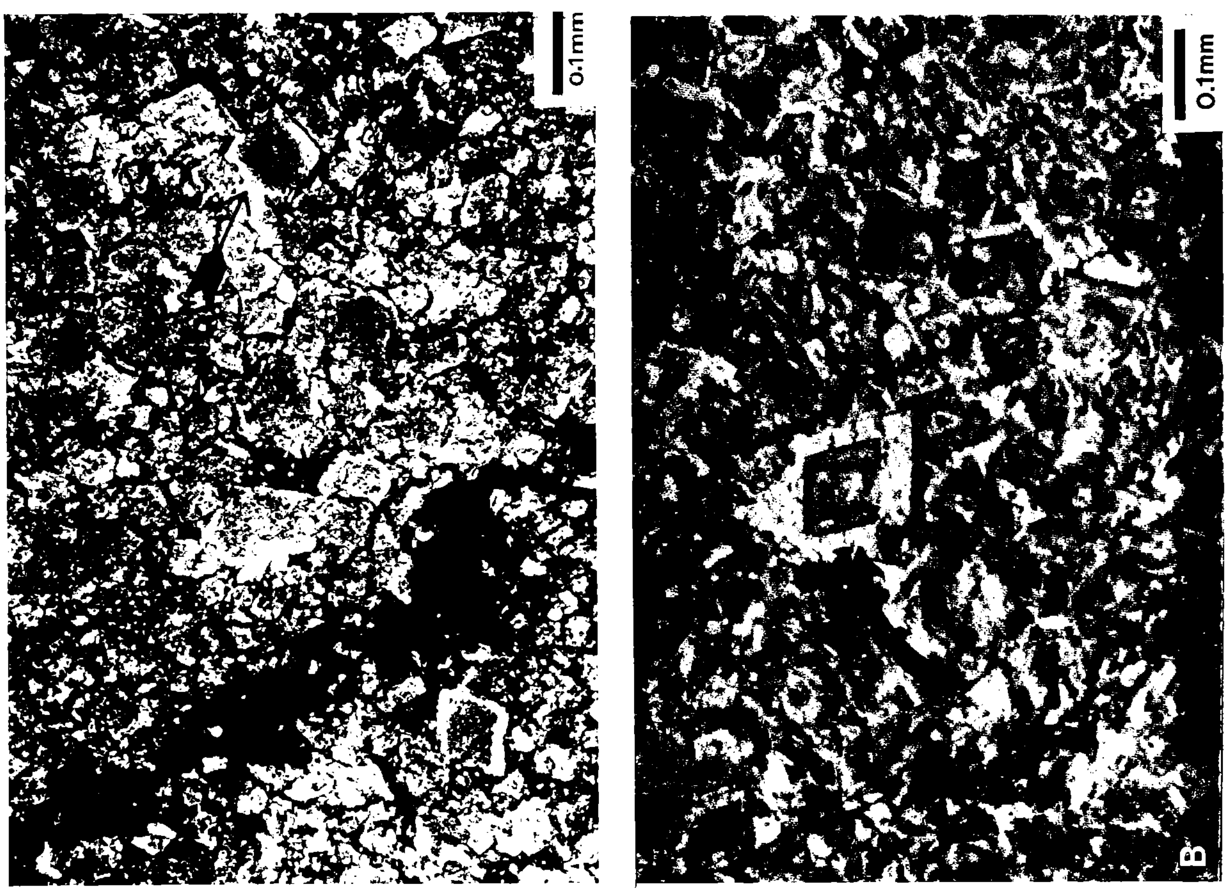


Figure 19. Dolomite of mixed fresh and marine water. *A*—Euhedral discrete and limpid rhombohedra formed around cloudy centers (arrows). *B*—Cathodoluminescence photomicrograph showing brightly luminescent zone around darker zone (arrow).

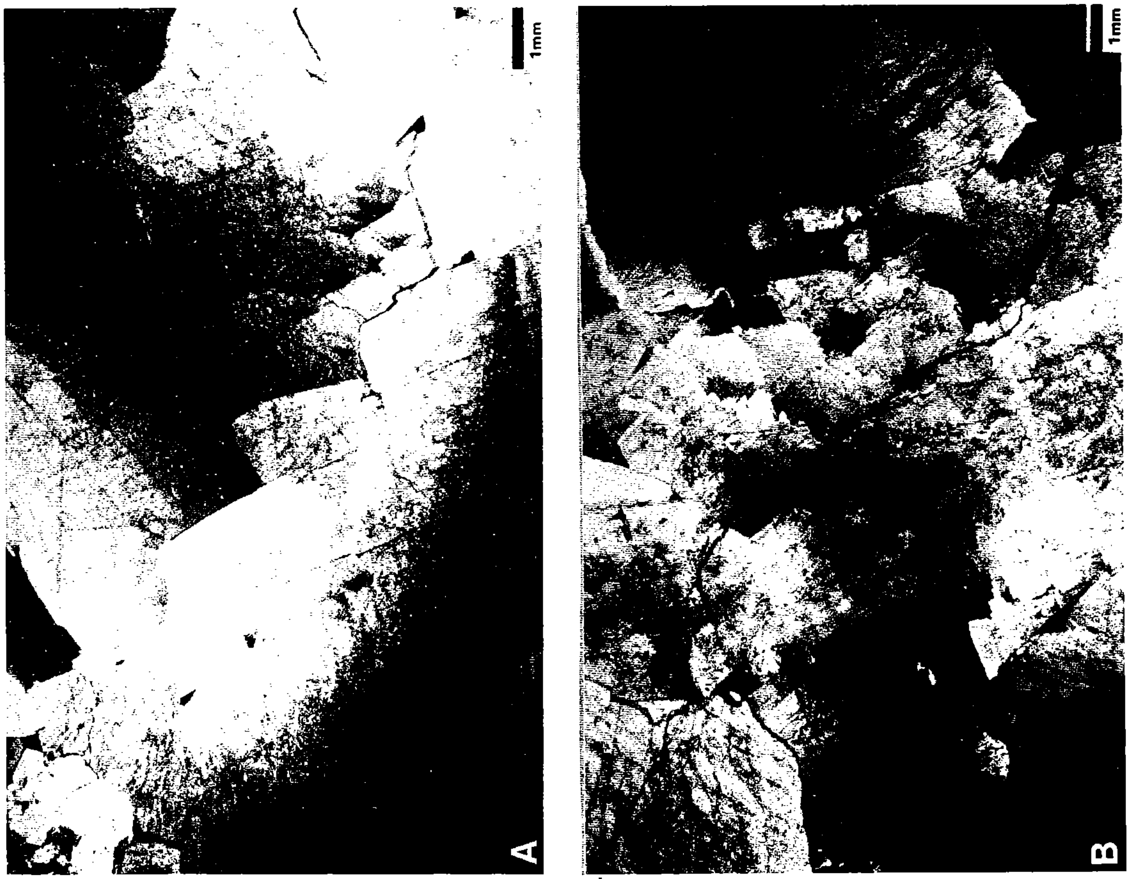


Figure 20. Saddle-type (baroque) dolomite, characterized by large size and undulose extinction. *A*—Plane-polarized light. *B*—Cross-polarized light.

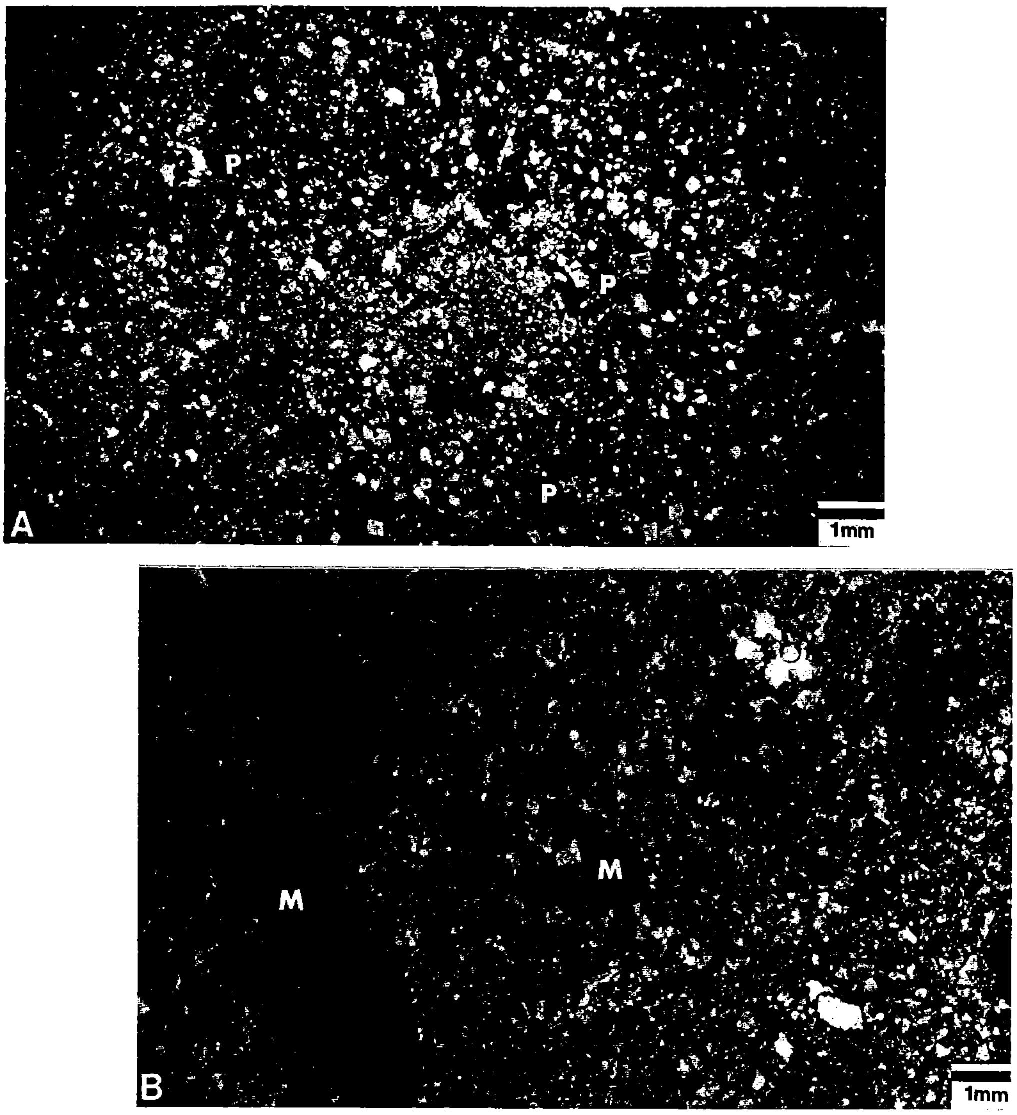


Figure 21. Porosity in intertidal facies. *A*—Secondary porosity (labeled "P") developed in loosely packed and burrowed zones. *B*—Moldic porosity (labeled "M") in dolo-wackestone. Rock is dolomitized and shows molds of pelmatozoans.

water mixed dolomite, and intertidal oolitic/fossil packstone subfacies was dolomitized as a mixture of hypersaline and mixed processes.

4. Dolomitization was most complete in the intertidal facies (facies II), due to burrowing-enhanced porosity and permeability that increased fluid conductivity and proximity to the source of dolomitizing fluids.

5. Four types of porosity developed in Hunton rocks: moldic, intercrystalline, vuggy/cavern, and fracture. Moldic and vuggy porosity are secondary, whereas intercrystalline can be either primary porosity preserved by dolomitization or late-secondary dissolution porosity. Fracture porosity enhances permeability in local areas, but is not as significant as the other types.

6. Porosity is directly related to depositional environment. Porosity was best developed in the intertidal facies (facies II). The characteristic features of this facies that caused it to preferentially develop porosity are burrowing and abundant echinoderms (pelmatozoans) and other fossils. Burrowing-enhanced porosity and permeability of the sediment allowed dolomitizing fluids to more effectively move through the carbonates. This dolomitization preserved porosity and permeability. Subsequent dissolving fluids were then able to attack the nondolomitized fossils, creating moldic porosity.

7. The supratidal facies (facies I), which often was dolomitized without burrowing and fossils, seldom develops significant porosity. Subtidal carbonate (facies III) was never highly dolomitized, due to its relative impermeability and its increased distance from the source of dolomitizing fluids.

8. Karstic processes of brecciation and dissolution generally enhance overall porosity in Hunton rocks, but are not necessary for porosity development.

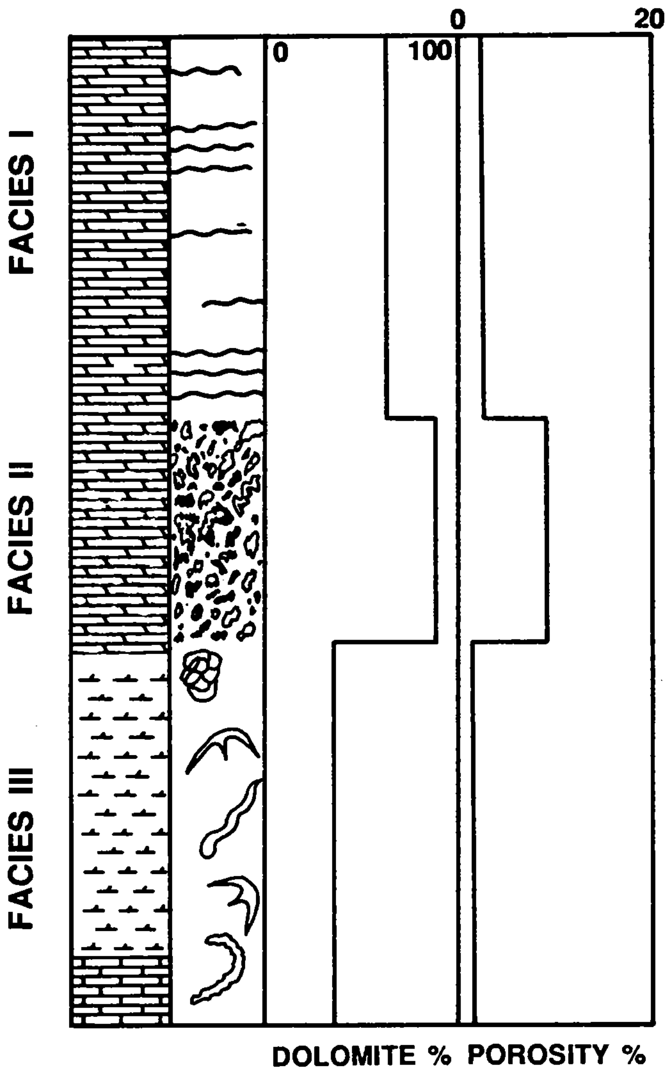


Figure 22. Schematic diagram illustrating, by percentage, the apparent relationship between porosity and dolomitization in the different facies of the Henryhouse.

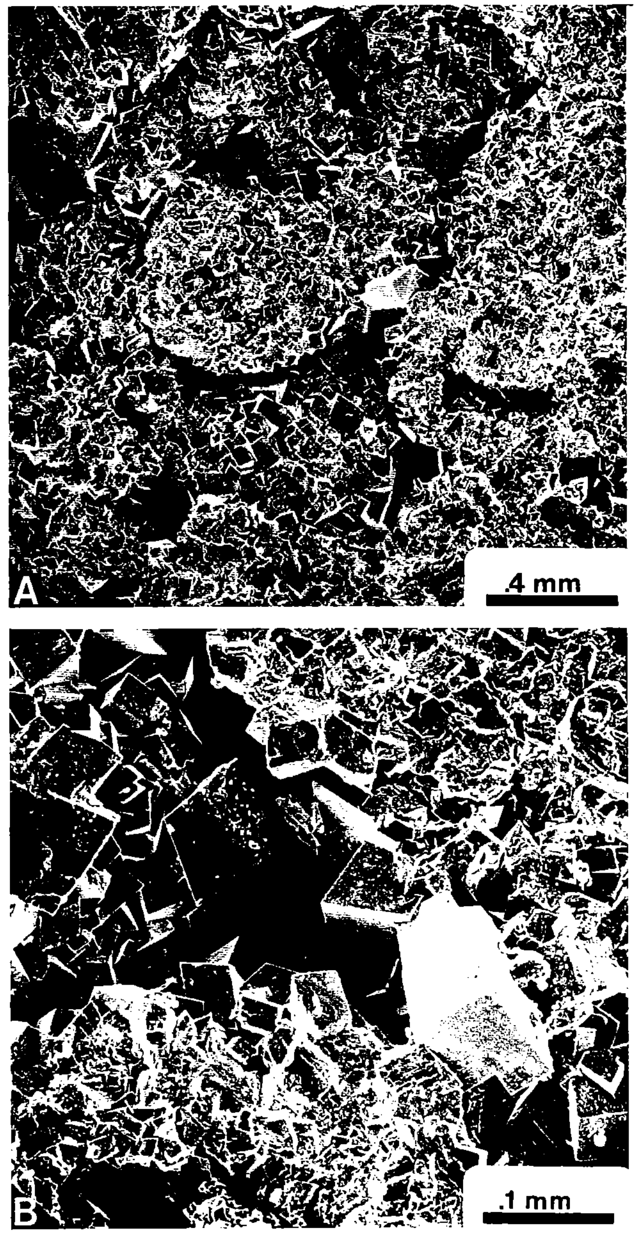


Figure 23. *A*—Interooidal and intraooidal porosity in the oolitic-shoal subfacies. Dolomitized oolitic grainstone with dominant interooid porosity and subordinated intraparticle porosity. *B*—Enlarged central part of *A*.

ACKNOWLEDGMENTS

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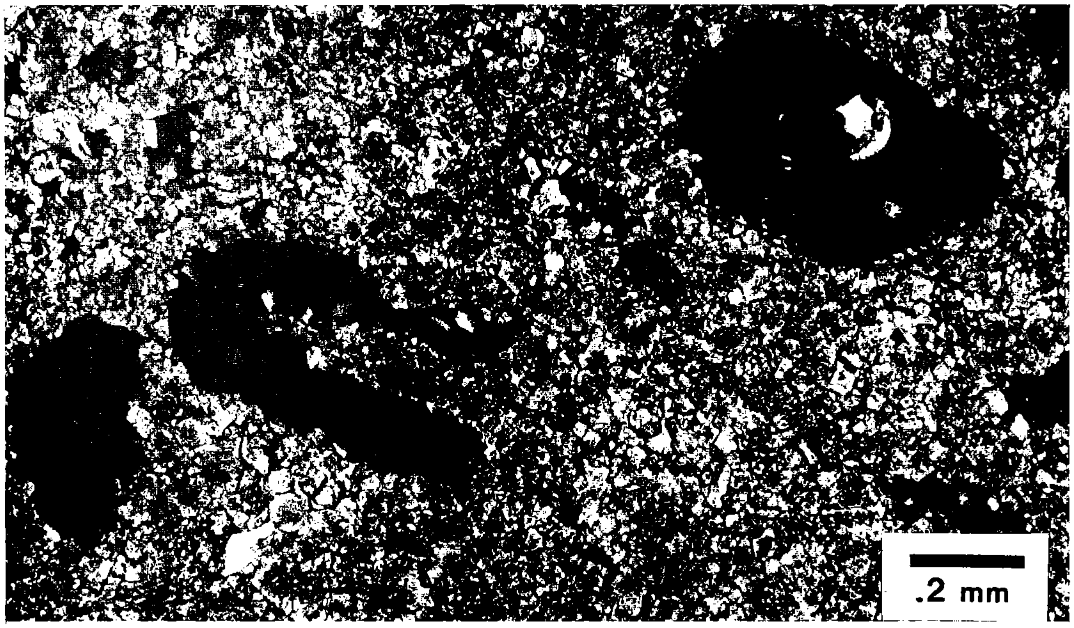

Figure 24. Crinoidal moldic porosity developed in the fossil-rich intertidal facies (II).

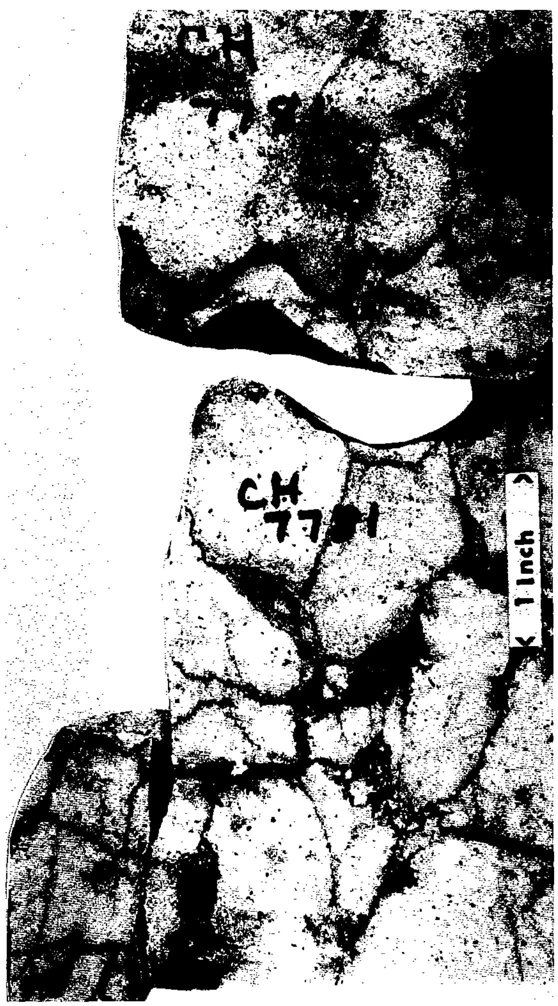

Figure 25. Small-scale solution-enlarged fracture porosity. Interstices filled with laminated cave muds, glauconite, and dolomite cement.

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APPENDIX

CORE DESCRIPTIONS

The appendix consists of core descriptions, petrolog forms, and core photographs of Hunton Group (Henryhouse Formation) carbonates from five wells drilled in the Anadarko basin.

Core: Amax Hickman No. 1-24

Location: sec. 24, T. 17 N., R. 18 W., Dewey County, OK

Cored Interval: 13,500 to 13,550 ft (depth below surface)

Stratigraphic Interval: Henryhouse Formation (Hunton Group)

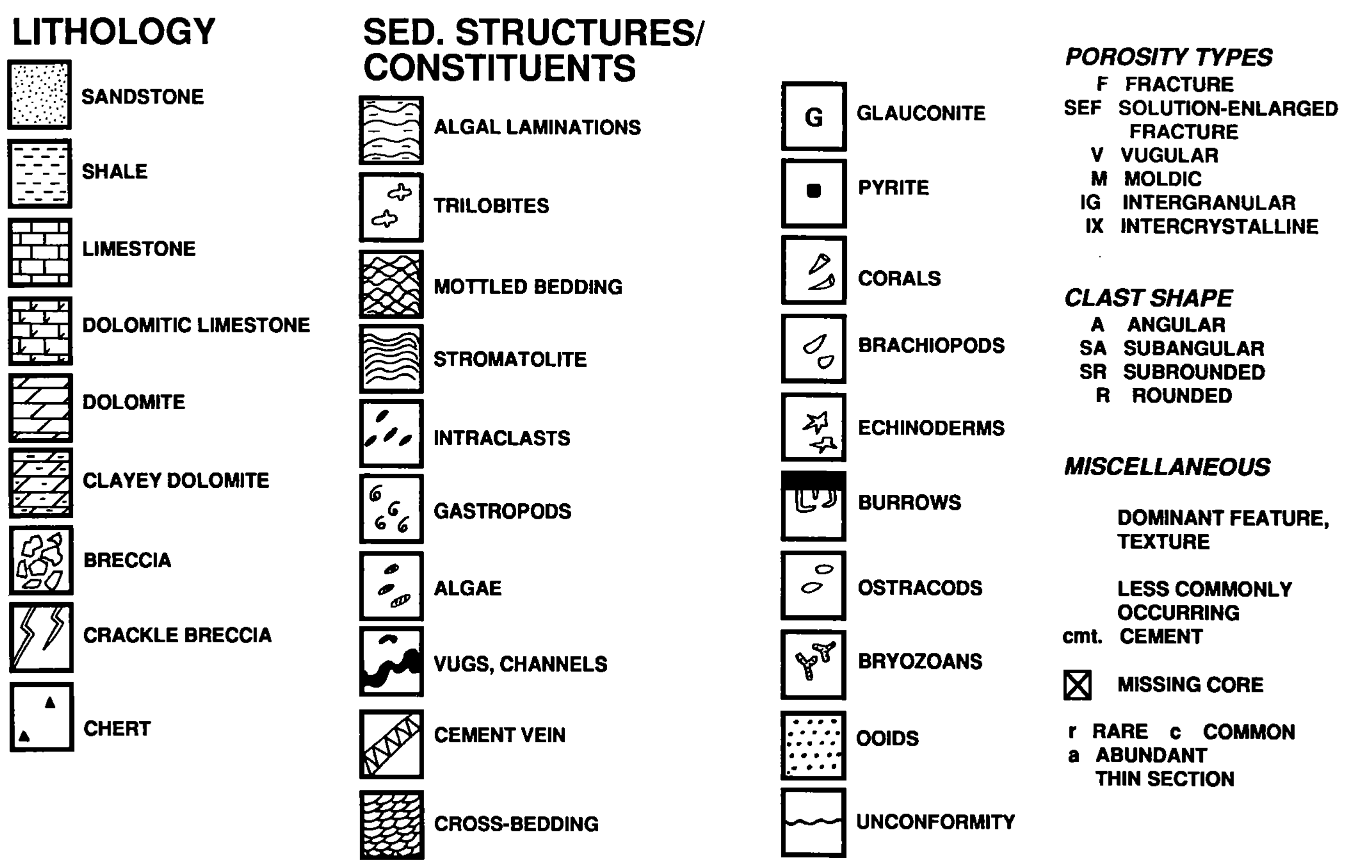
Core Description: The following description is given in reference to the core petrolog and photograph immediately following this text.

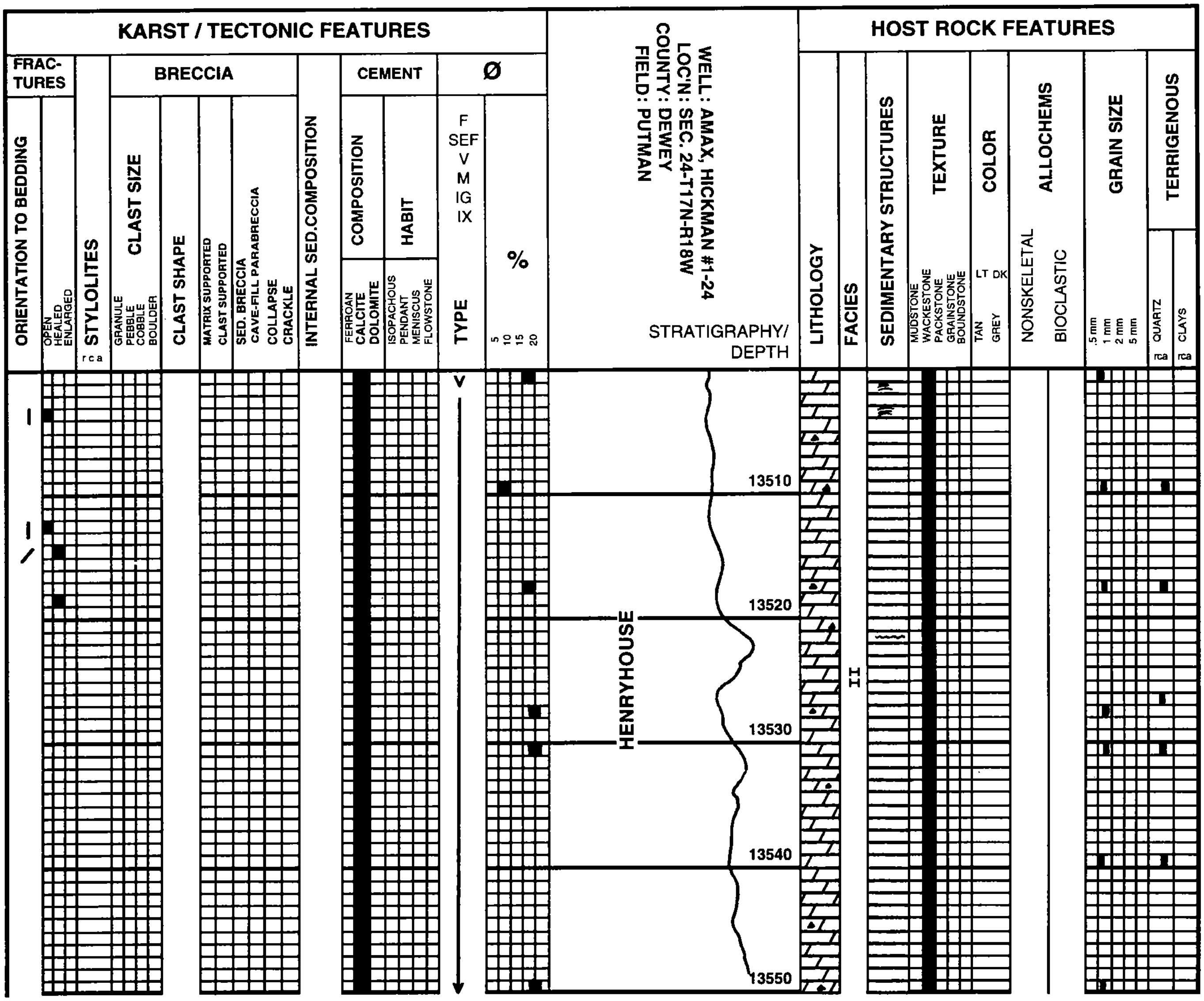
This core is medium bluish-gray dolo-wackestone with excellent moldic and vuggy porosity. The rock is oil stained and bioturbated throughout. Other sedimentary structures are faint algal laminations and sparse stylolites. In the lower portion, vugs may be filled with calcite. Measured porosity ranges from 7 to 17% in the burrowed host rock. The environment is thought to be intertidal.

PETROLOG

WELL ------------------------------------

SEC. -------- T --------R -------- ----------------- COUNTY, OK

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FRAC-TURES
BRECCIA
CEMENT
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OPEN
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CLAST SIZE
GRANULE
PEBBLE
COBBLE
BOULDER
CLAST SHAPE
MATRIX SUPPORTED
CLAST SUPPORTED
SED. BRECCIA
CAVE-FILL PARABRECCIA
COLLAPSE
CRACKLE
INTERNAL SED.COMPOSITION
COMPOSITION
FERROAN
CALCITE
DOLOMITE
HABIT
ISOPACHOUS
PENDANT
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%
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WELL: AMAX, HICKMAN #1-24
LOC'N: SEC. 24-T17N-R18W
COUNTY: DEWEY
FIELD: PUTMAN
STRATIGRAPHY/ DEPTH
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13520
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HENRYHOUSE
HOST ROCK FEATURES
LITHOLOGY
FACIES
II
SEDIMENTARY STRUCTURES
TEXTURE
MUDSTONE
WACKESTONE
PACKSTONE
GRAINSTONE
BOUNDSTONE
COLOR
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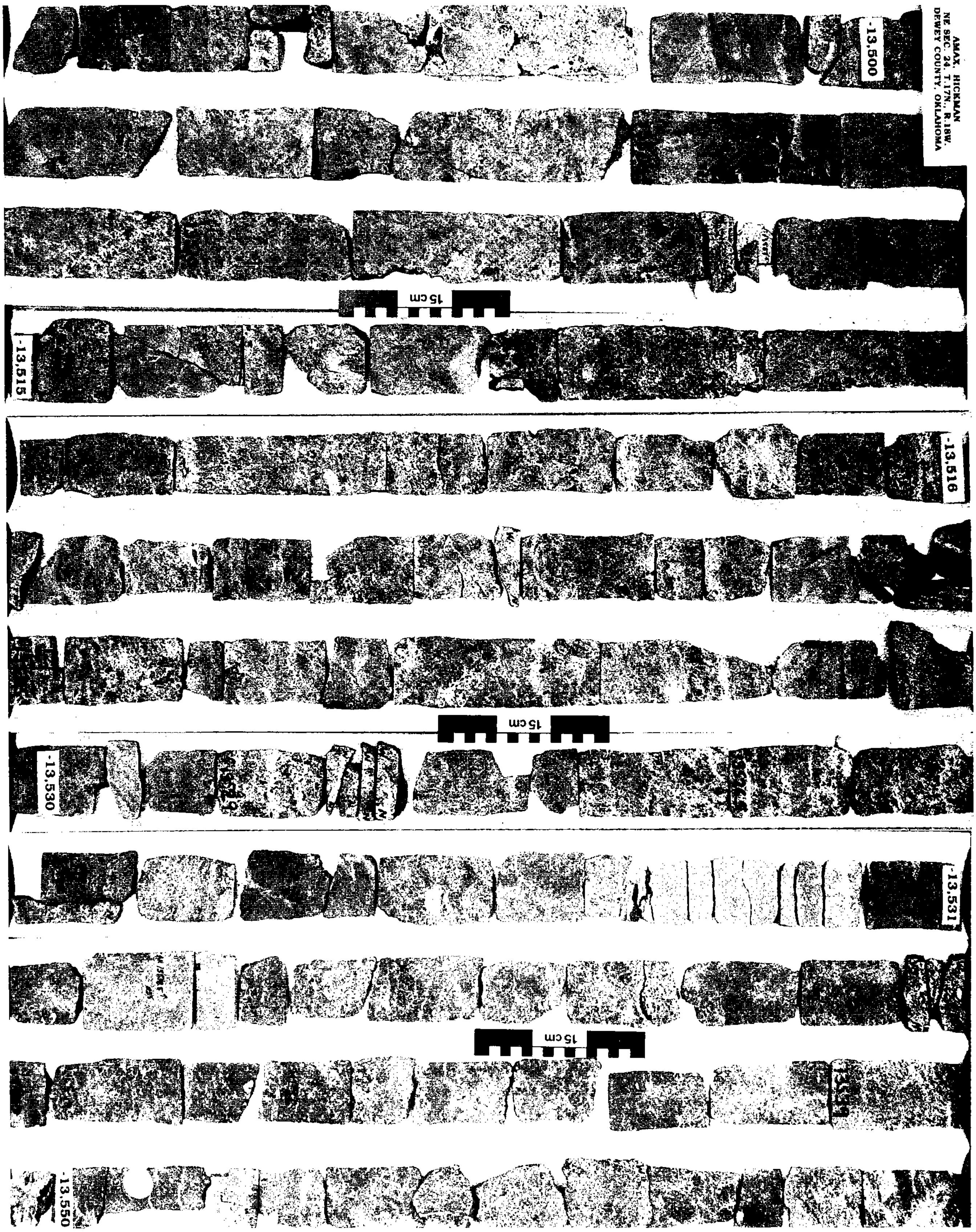
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NE SEC. 24, T.17N., R.18W.
DEWEY COUNTY, OKLAHOMA
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-13,530
-13,531
15 cm
-13,550

Core: Tenneco Jordan No. A-1

Location: sec. 3, T. 21 N., R. 14 W., Major County, OK

Cored Interval: 8,495 to 8,612 ft

Stratigraphic Interval: Henryhouse Formation (Hunton Group)

Core Description: The following description is given in reference to the core petrolog and photographs immediately following this text.

The Tenneco Jordan No. A-1 core is a teaching core from the OSU Geology Department. The noticeable "blotches" on the slabs in the photograph are caused by drops of dilute hydrochloric acid used during students' examinations of the core.

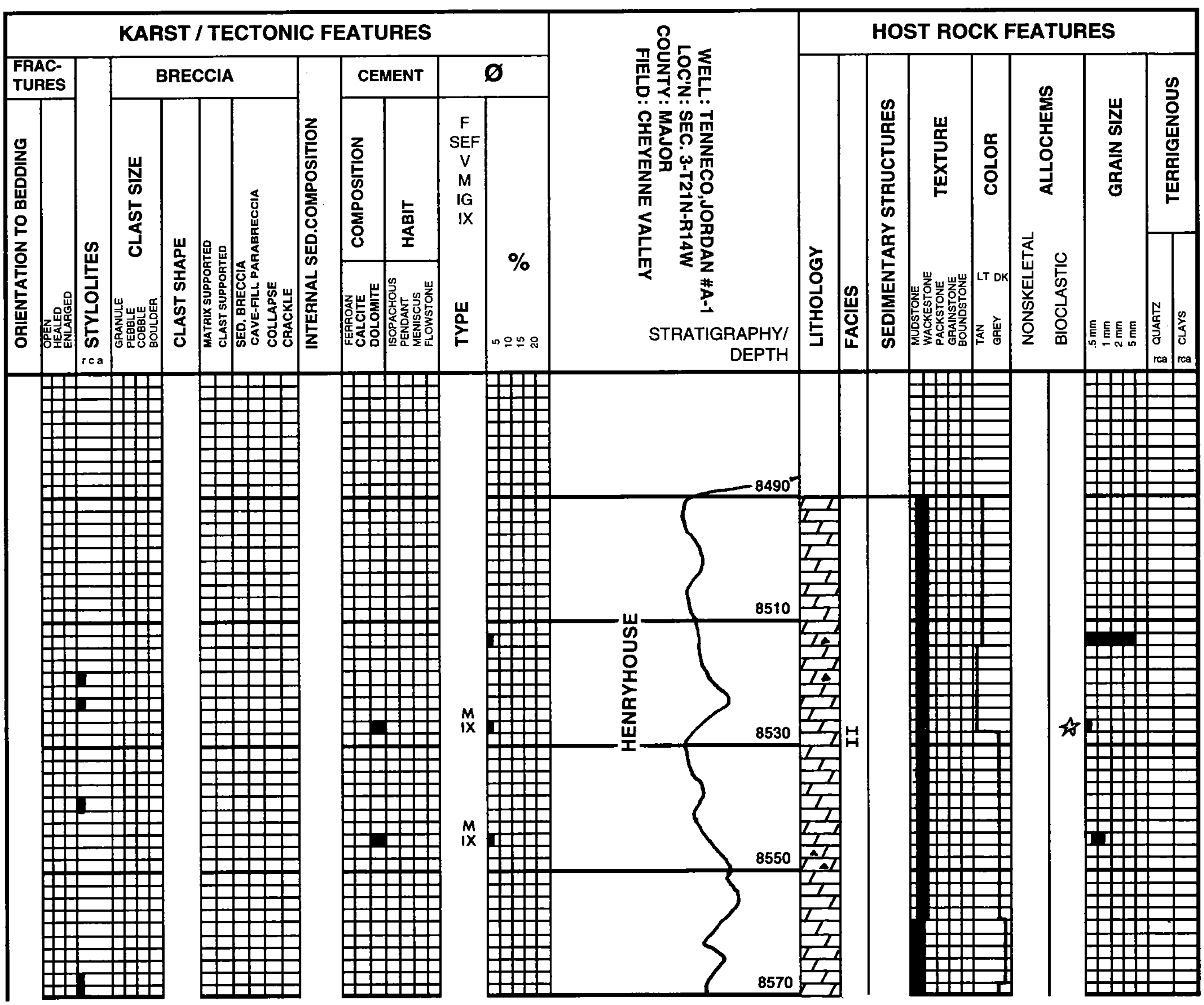
The core contains two major facies:

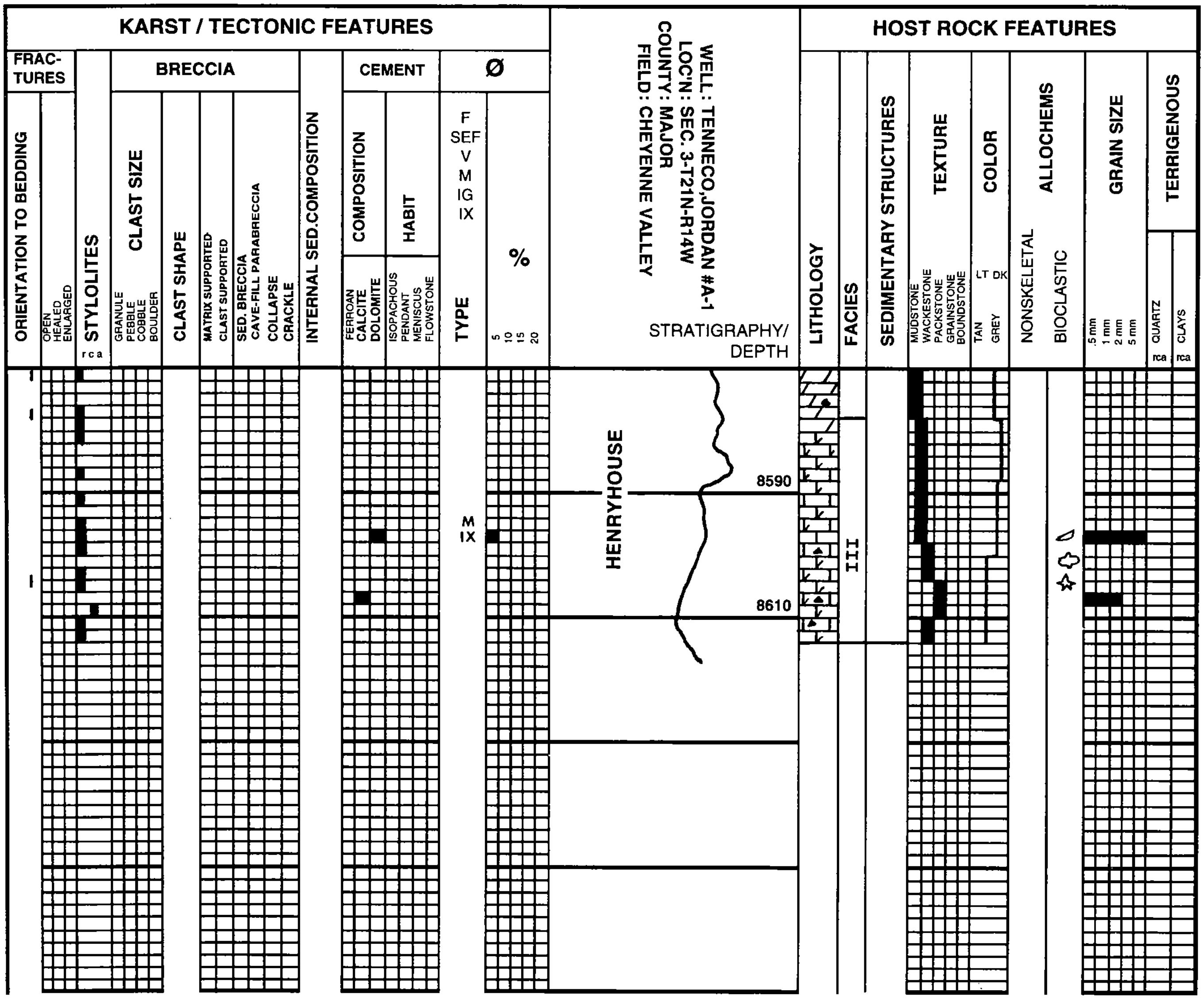
8,495 to 8,579 ft

The upper interval encompasses a dark oil-stained brown-gray, burrow/mottled, dolo-mudstone/wackestone. Fossil existence is indicated by crinoid-shaped molds. Dolomite is sucrosic matrix-replacive rhombs, providing 5–18% porosity. The porosity is also due to fractures, enlarged pores, interparticle pores, and moldic pores. At 8,514 ft, a 7-cm vug-filling anhydrite nodule is exposed. Other smaller nodules contain chert, quartz, and dolomite. Stylolites are rare. The lower part of the interval has fewer burrows. The color is lighter and grayer. This interval is believed to be an intertidal facies.

8,579 to 8,612 ft

This segment of the core is a medium-gray, dolomudstone/wackestone to packstone. The dolo-mudstone/wackestone contains some moldic porosity and horizontal and vertical burrows. Fossils include crinoids, trilobites, brachiopods, bryozoans, and corals. Molds of fossils are filled with sparite. Stylolites are common. Porosity ranges from 0 to 5%. This interval is thought to represent a subtidal environment with a subtidal shoal.




KARST / TECTONIC FEATURES
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ORIENTATION TO BEDDING
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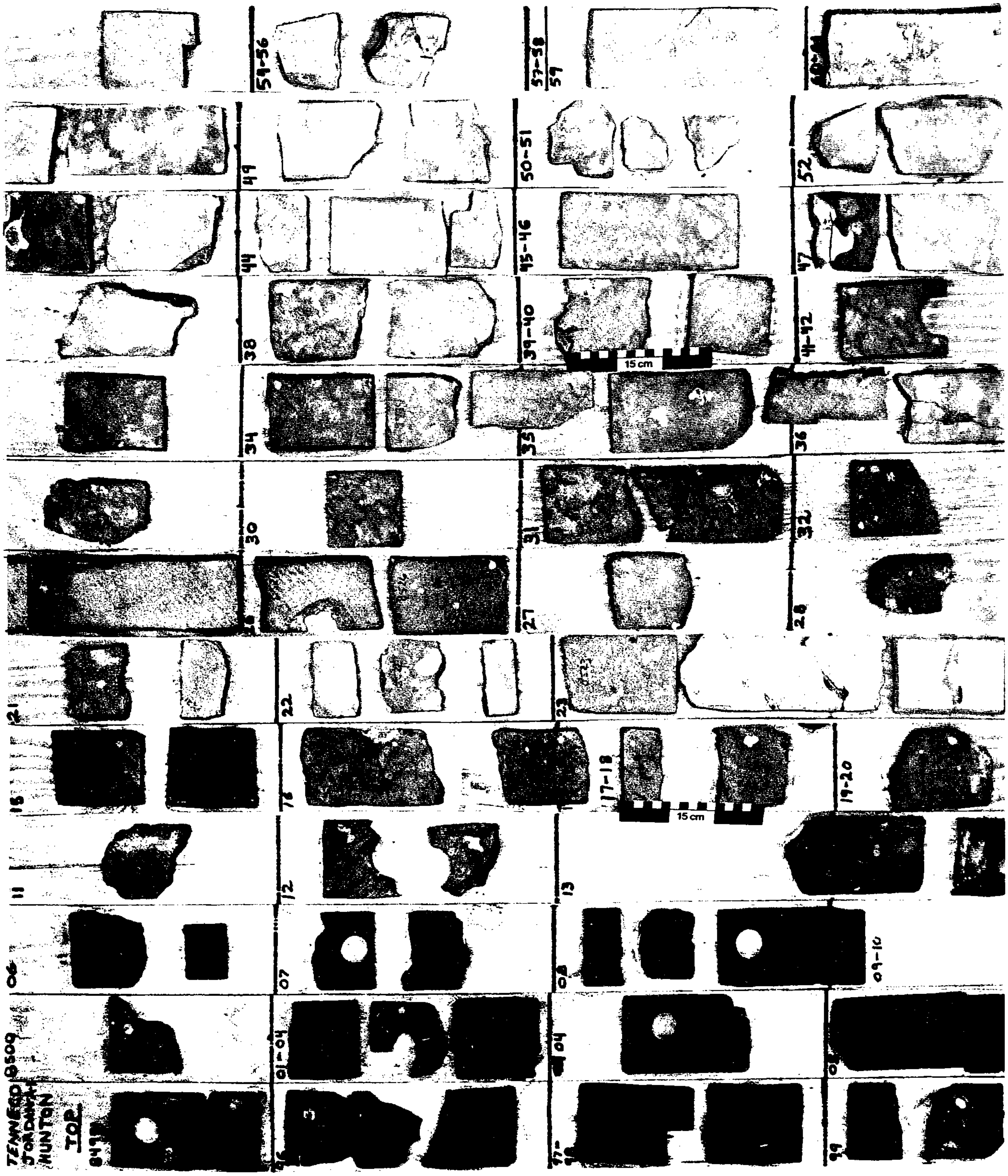
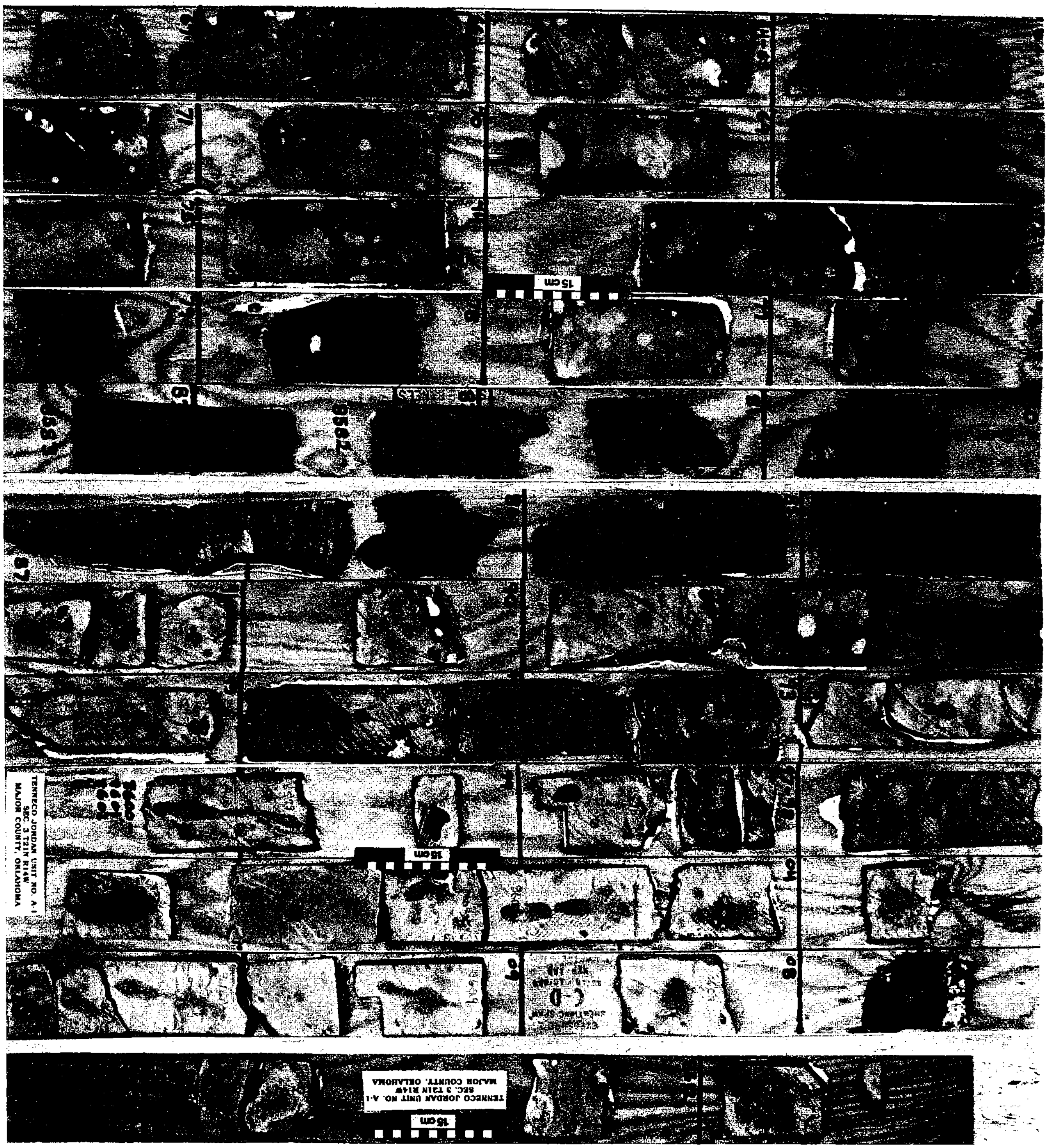
CLAST SIZE
GRANULE PEBBLE COBBLE BOULDER
CLAST SHAPE
MATRIX SUPPORTED CLAST SUPPORTED
SED. BRECCIA CAVE-FILL PARABRECCIA COLLAPSE CRACKLE
INTERNAL SED.COMPOSITION
CEMENT
COMPOSITION
FERROAN CALCITE DOLOMITE
HABIT
ISOPACHOUS PENDANT MENISCUS FLOWSTONE
Ø
F SEF V M IG IX
TYPE
%
5 10 15 20
M IX
WELL: TENNECO,JORDAN #A-1
LOC'N: SEC. 3-T21N-R14W
COUNTY: MAJOR
FIELD: CHEYENNE VALLEY
STRATIGRAPHY/ DEPTH
HENRYHOUSE
8590
8610
HOST ROCK FEATURES
LITHOLOGY
FACIES
III
SEDIMENTARY STRUCTURES
TEXTURE
MUDSTONE WACKESTONE PACKSTONE GRAINSTONE BOUNDSTONE
COLOR
LT DK
TAN GREY
ALLOCHEMS
NONSKELETAL
BIOCLASTIC
GRAIN SIZE
.5 mm 1 mm 2 mm 5 mm
TERRIGENOUS
QUARTZ
CLAYS
rca


15 cm
15 cm


TENNECO JORDAN UNIT NO. A-1
SEC. 3 T21N R14W
MAJOR COUNTY, OKLAHOMA
15 cm

Core: Eason Van Curen No. 1

Location: sec. 3, T. 15 N., R. 5 W., Kingfisher County, OK

Cored Interval: 7,136 to 7,192 ft

Stratigraphic Interval: Henryhouse Formation (Hunton Group)

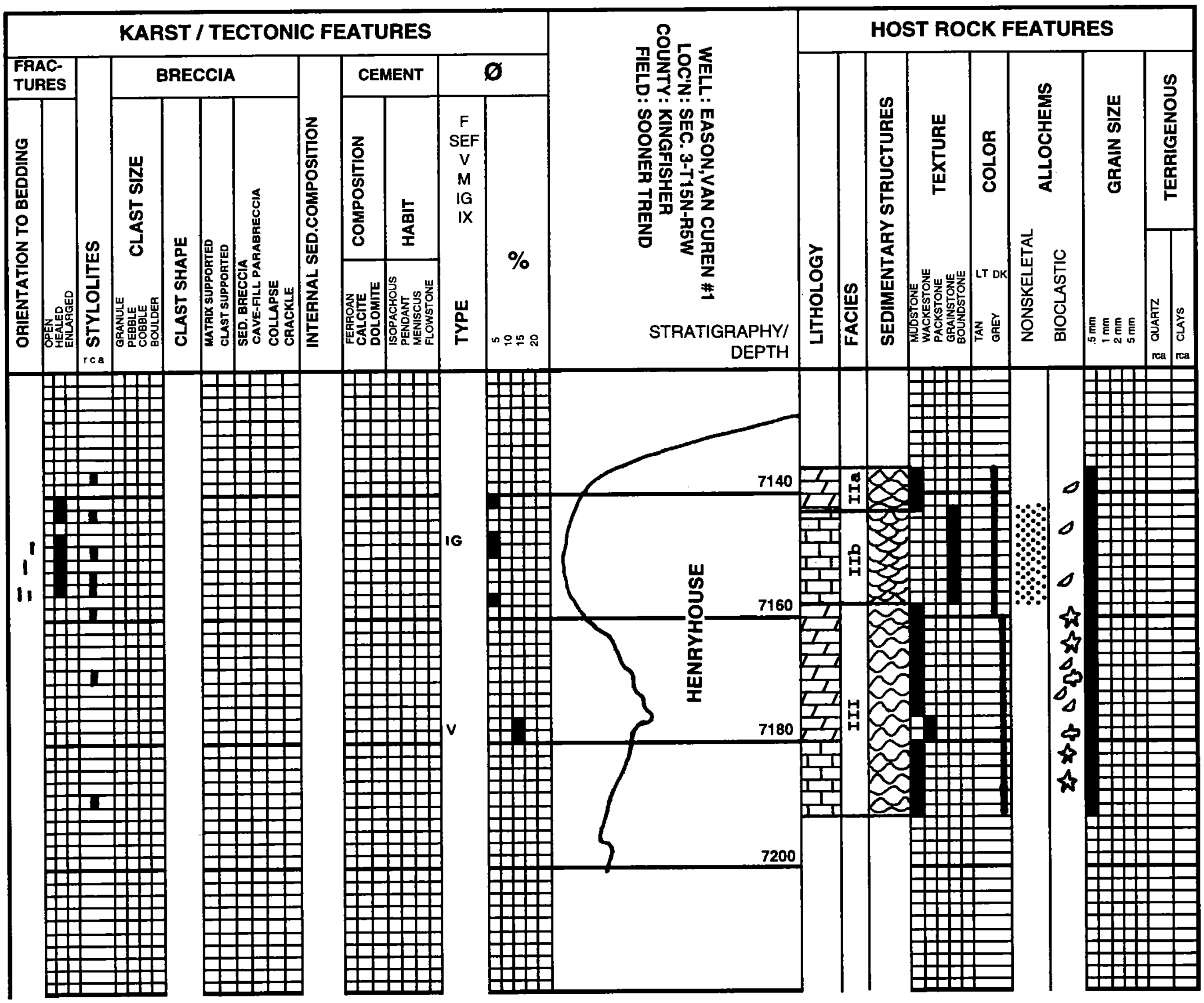
Core Description: The following description is given in reference to the core petrolog and photographs immediately following this text.

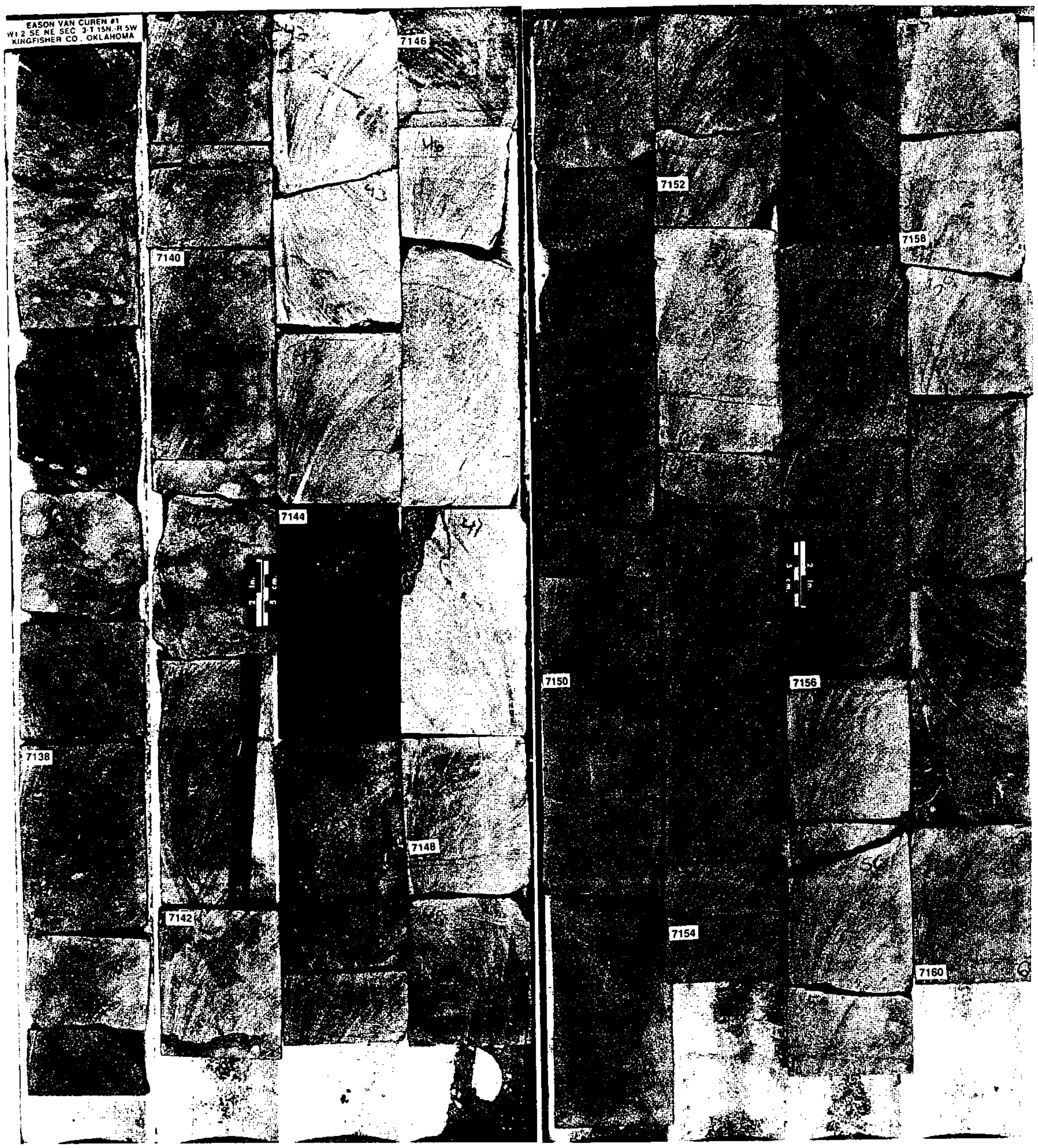
In this core, an interval of 56 ft of the Henryhouse is present. Three facies are described below: lagoonal, oolitic, and subtidal facies.

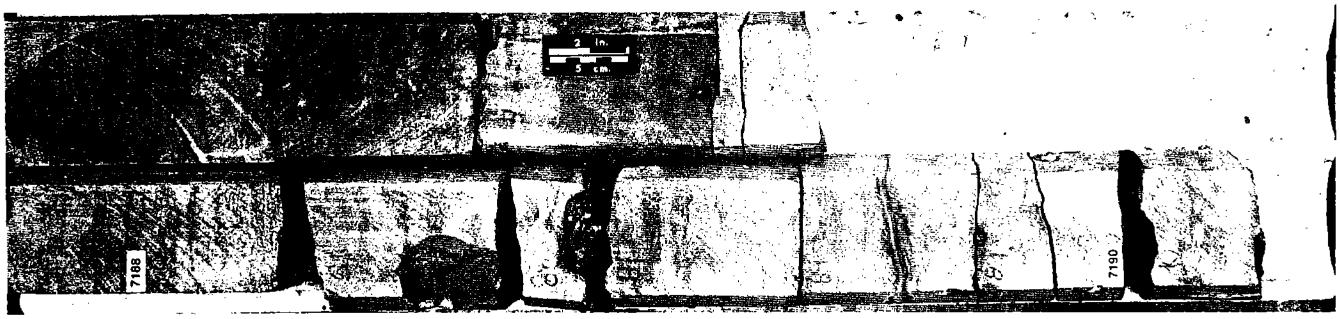
A peloid dolo-mudstone (7,136–7,143 ft) represents the lagoonal facies. The dolomite is cloudy and xenotopic, and its percentage decreases toward contact with the underlying unit. Dolomitization has obliterated many sedimentary structures, but faint burrows are present. Rare fossils include pelmatozoans and brachiopods.

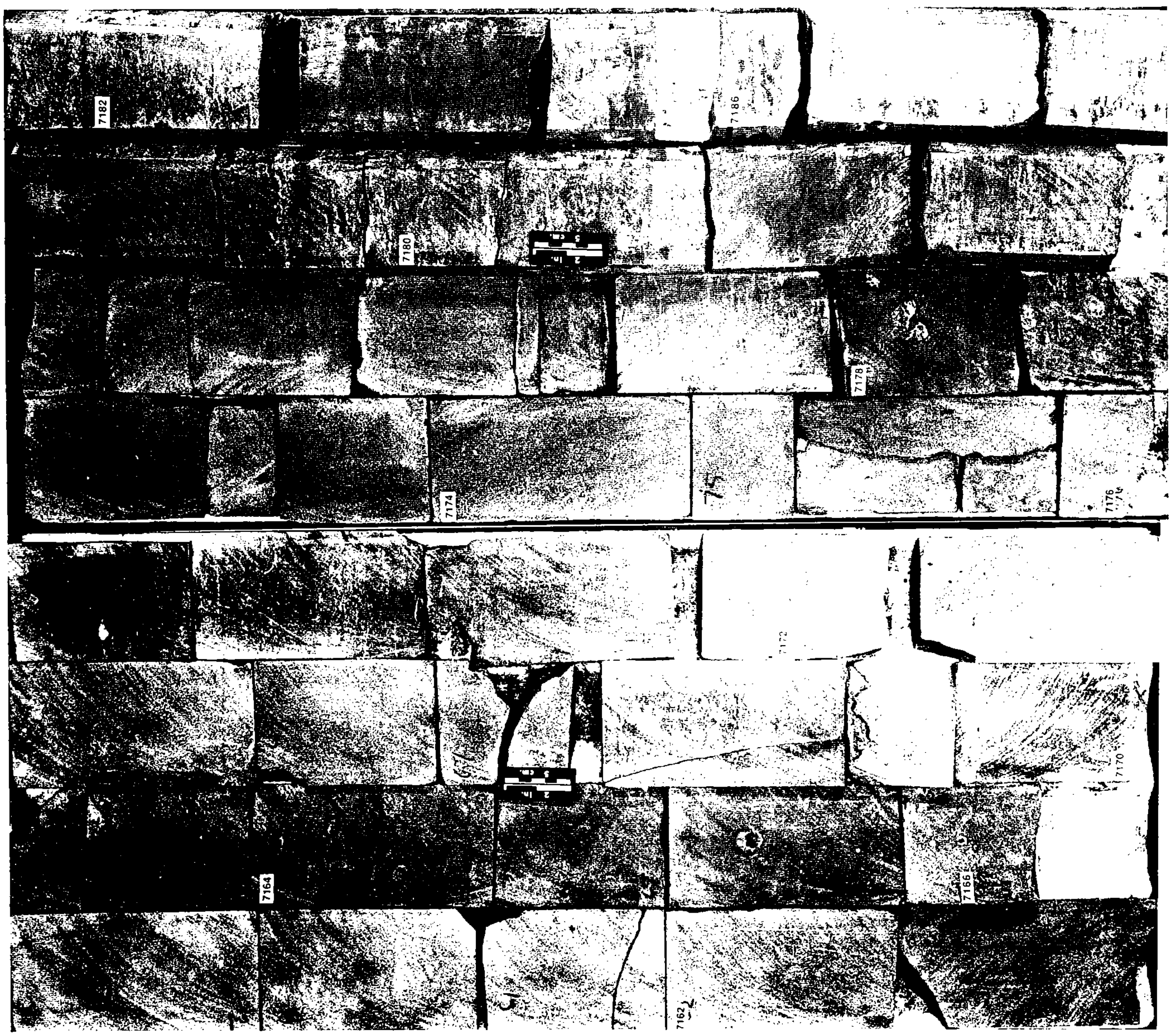
Below the lagoonal facies is an oolitic facies (7,143–7,159 ft), which is pure calcite. The rock is a grainstone with abundant ooliths and rare fossil debris. Cross and horizontal stratification is present. Drusy spar occludes most of the porosity.

The lower part of the Hunton is a subtidal facies (7,159–7,192 ft). This facies is a dolo-mudstone at the top and a lime-mudstone at the bottom. Hummocky bedding is present. Fossil fragments include crinoids, brachiopods, bryozoans, trilobites, and corals.




EASON VAN CUREN #1
W1 2 SE NE SEC 3-T 15N.-R 5W
KINGFISHER CO. OKLAHOMA
7138
7140
7142
7144
7146
7148
7150
7152
7154
7156
7158
7160


2 in.
5 cm.
7188
7190


7182
7186
7180
7178
7174
7176
7172
7170
7164
7166
7162

Core: Apexco Green No. 1

Location: sec. 31, T. 10 N., R. 26 W., Beckham County, OK

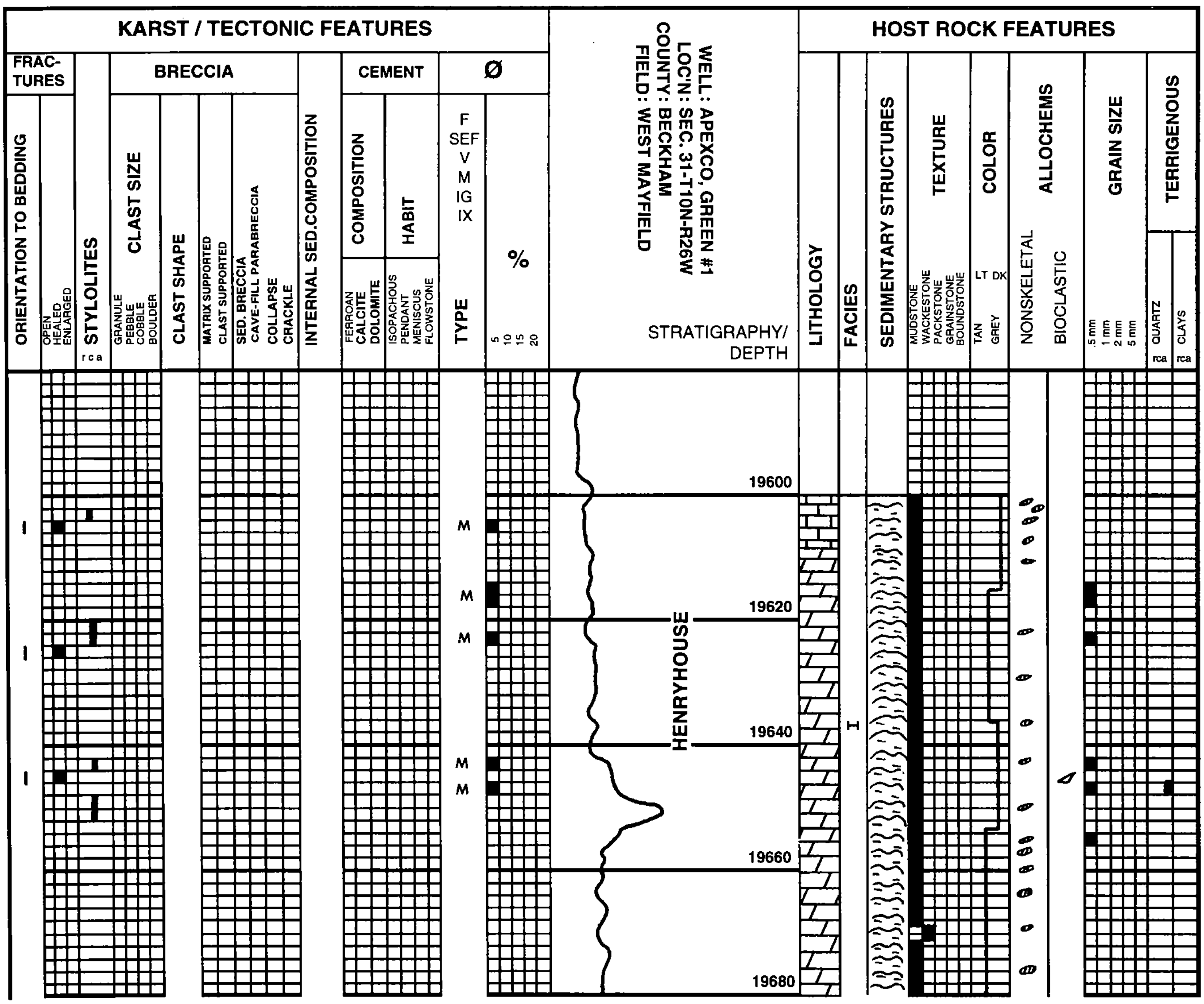
Cored Interval: 19,599 to 19,770 ft

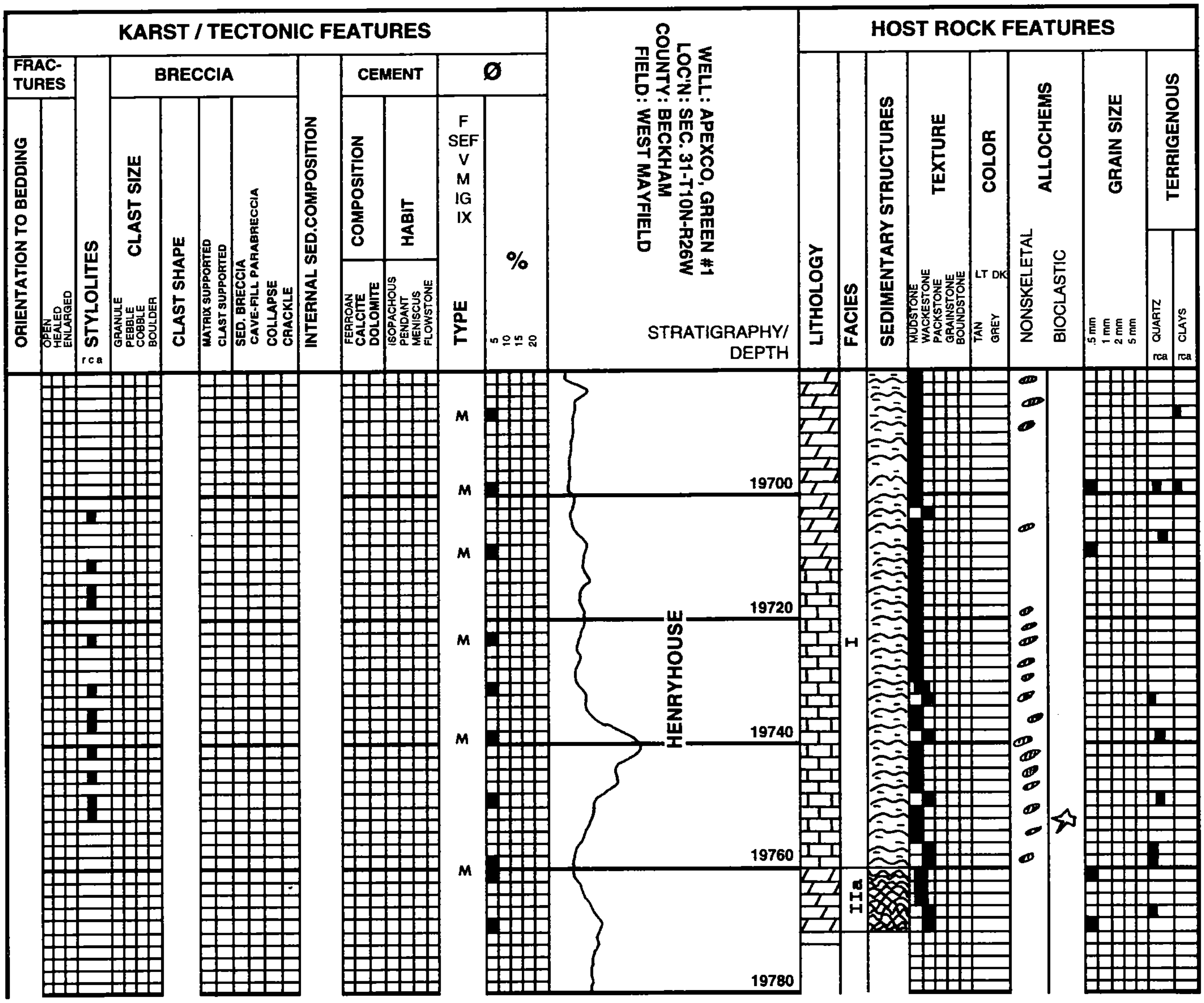
Stratigraphic Interval: Henryhouse Formation (Hunton Group)

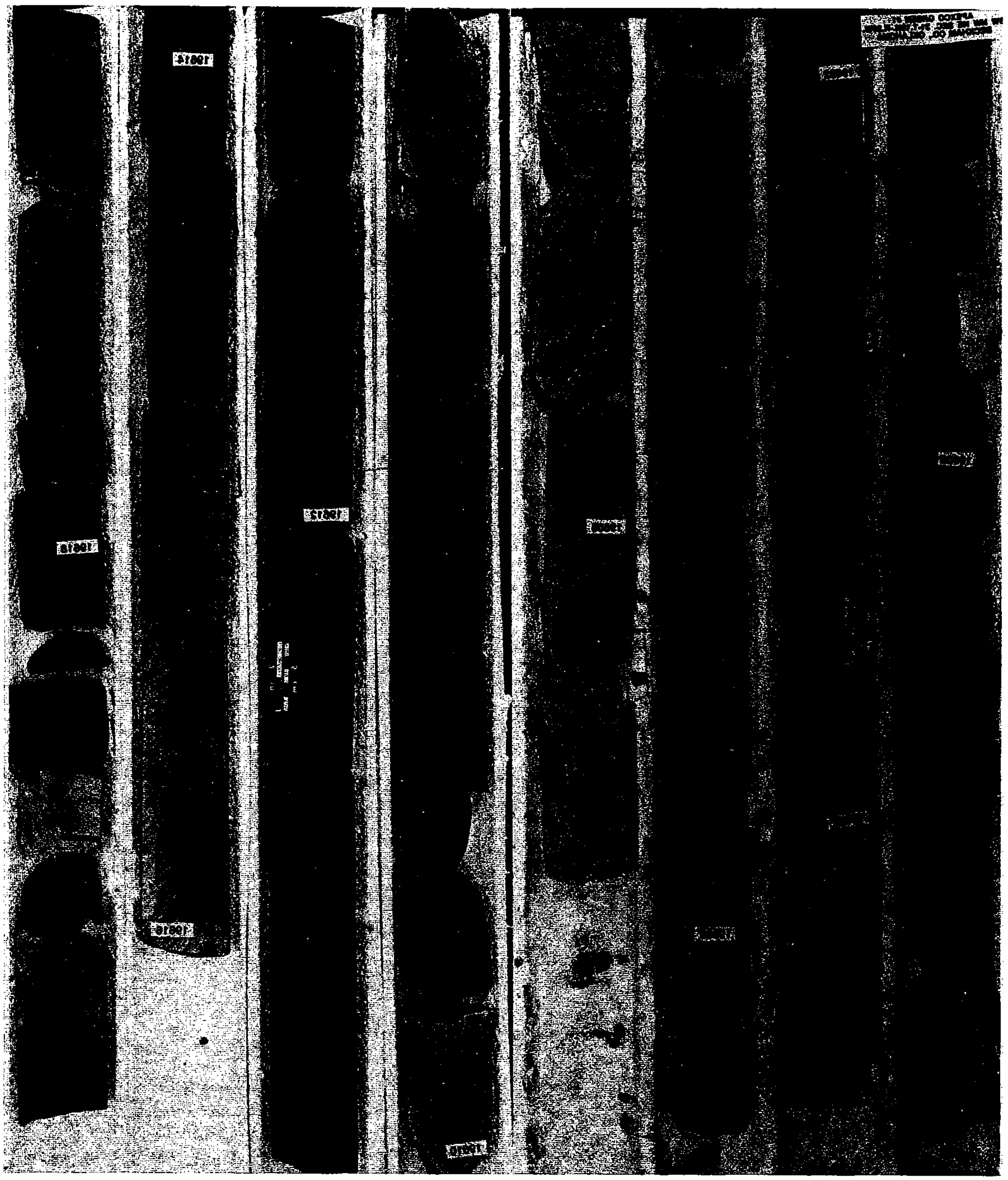
Core Description: The following description is given in reference to the core petrologs and photographs immediately following this text.

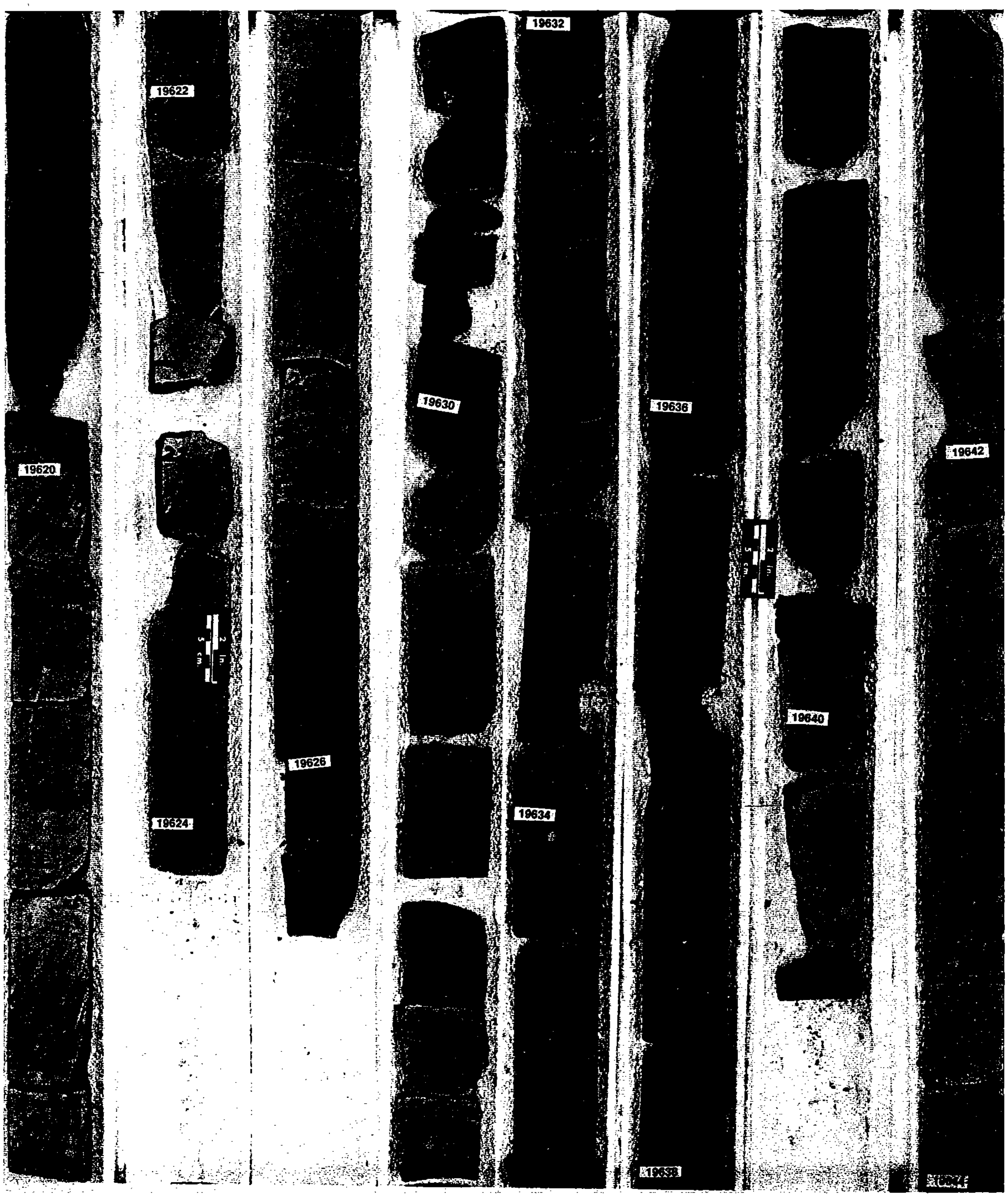
The cored interval consists largely of a single facies, with a possible second facies at the base that may indicate marine regression. The upper facies (19,599–19,770 ft) is a light- to dark-gray dolomite, dolomitic and lime mudstone, and peloidal lime-wackestone. Fossils are rare or absent. Fenestral fabrics, algal laminations, irregular carbonate laminations, and massive fabrics are all common. Intraclasts are rare and fine, whereas low-amplitude stylolites are abundant, especially in the lime-mudstone. The lack of fossils and type of fabrics is indicative of a restricted tidal-flat depositional environment (facies I).

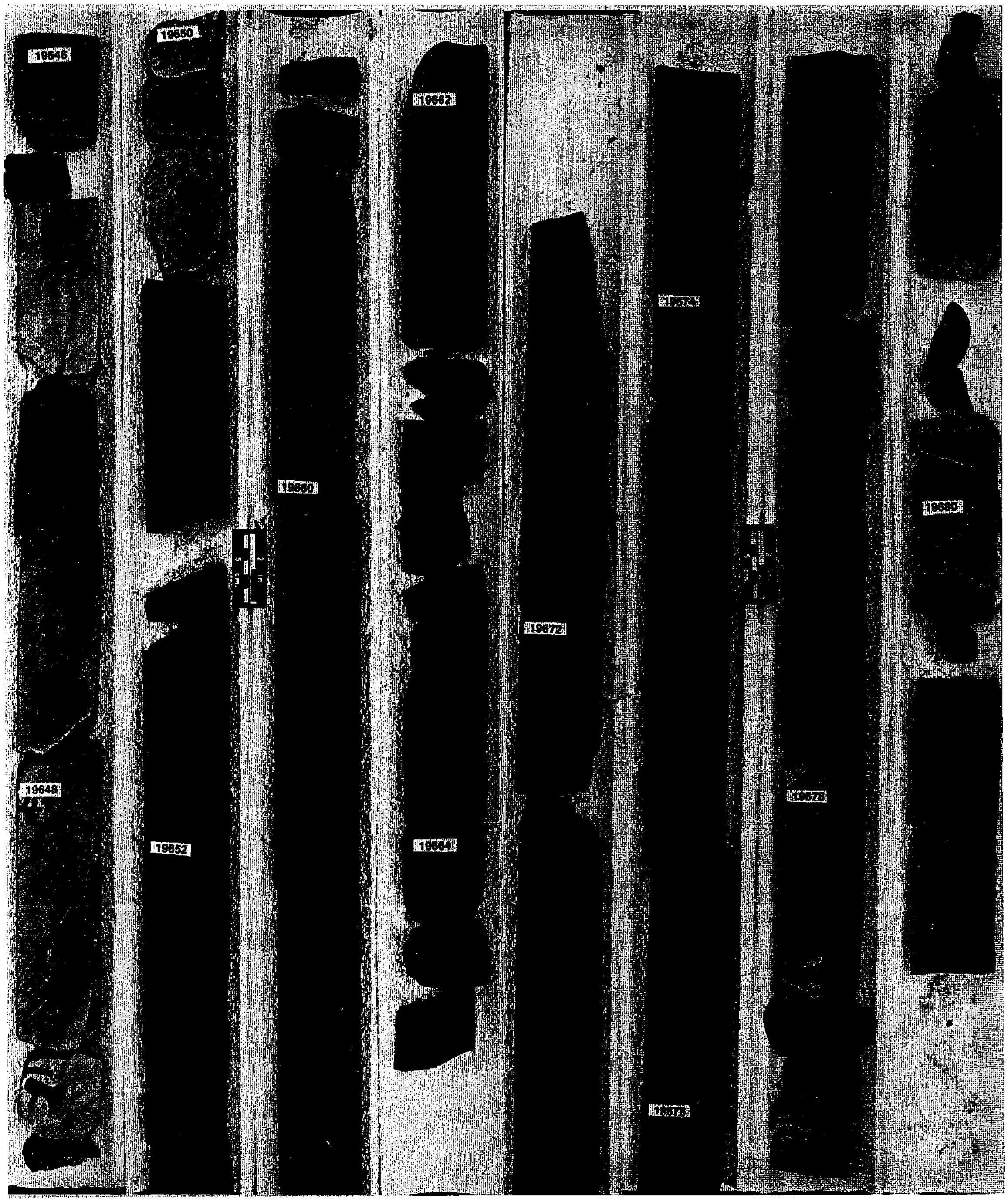
The lowermost 10 ft (19,760–19,770 ft) is represented by a small number of samples, making recognition of facies criteria difficult. Two thin sections exhibit good porosity, resulting from what are interpreted as molds and enlarged molds, along with a burrowed appearance. The increase in fossils and presence of burrows seem to indicate a transition into a zone of higher energy (facies IIa).

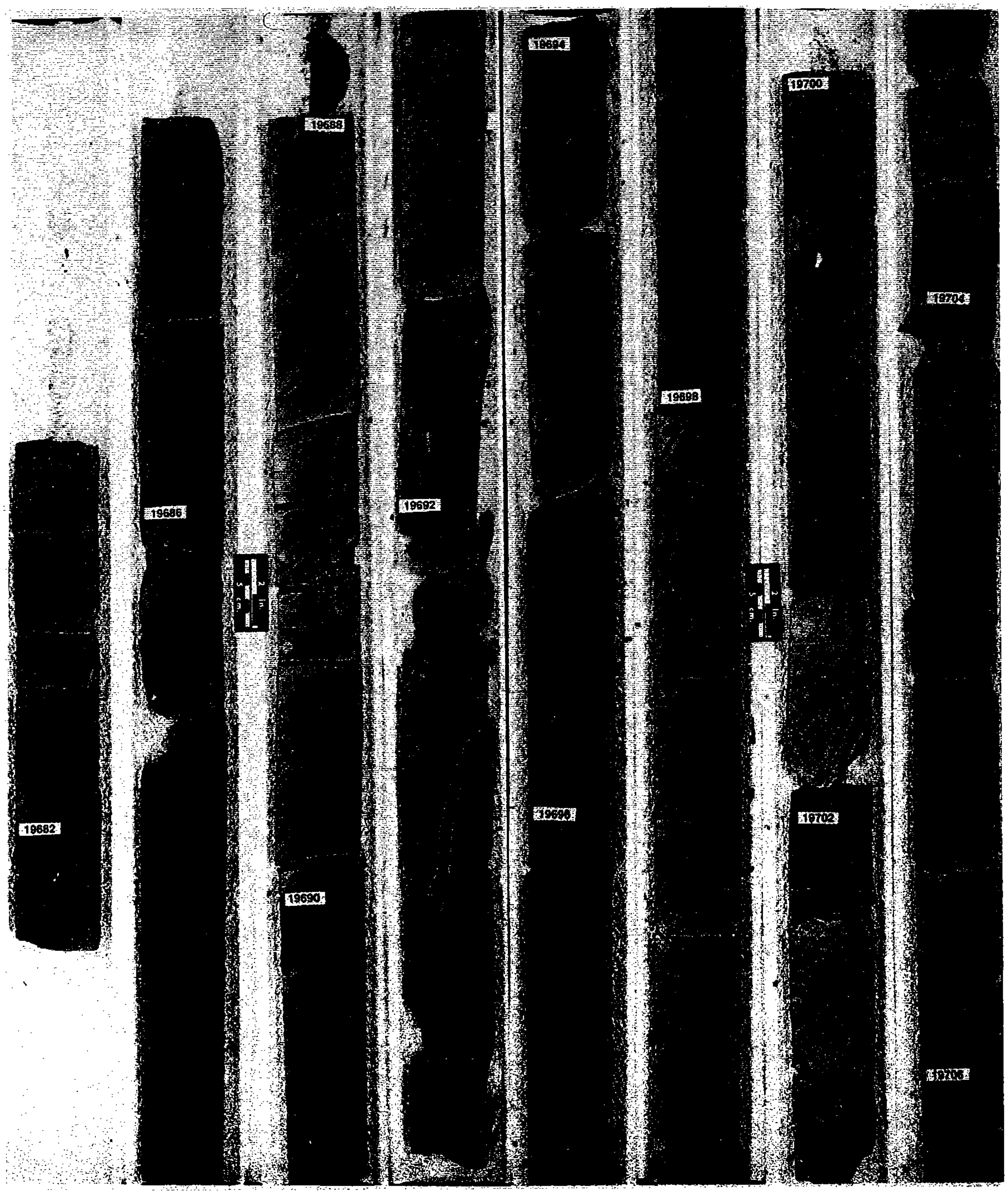


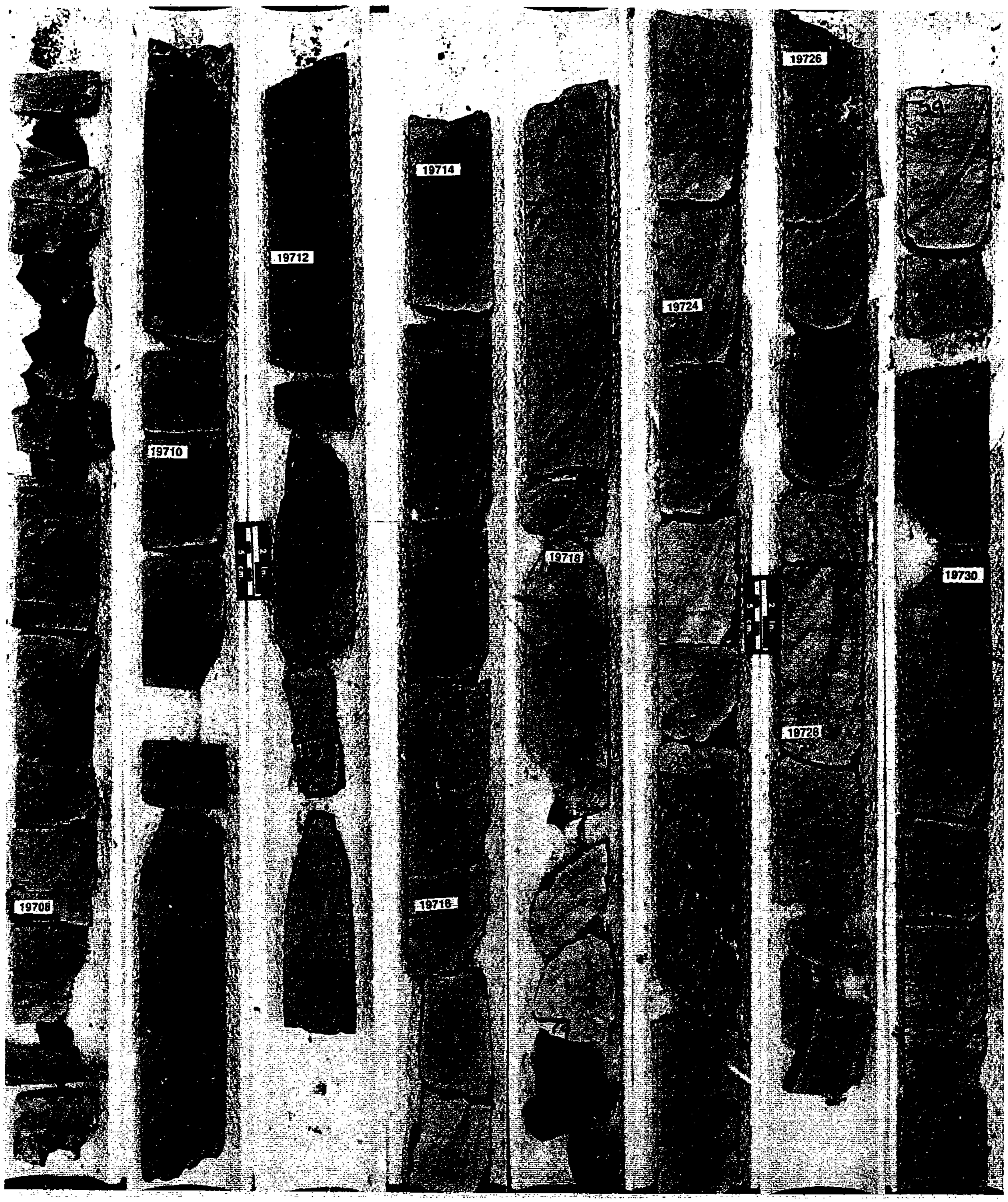

KARST / TECTONIC FEATURES
FRAC-TURES
BRECCIA
CEMENT
Ø
ORIENTATION TO BEDDING
OPEN
HEALED
ENLARGED
STYLOLITES
rca
CLAST SIZE
GRANULE
PEBBLE
COBBLE
BOULDER
CLAST SHAPE
MATRIX SUPPORTED
CLAST SUPPORTED
SED. BRECCIA
CAVE-FILL PARABRECCIA
COLLAPSE
CRACKLE
INTERNAL SED.COMPOSITION
COMPOSITION
FERROAN
CALCITE
DOLOMITE
HABIT
ISOPACHOUS
PENDANT
MENISCUS
FLOWSTONE
F
SEF
V
M
IG
IX
TYPE
%
5
10
15
20
WELL: APEXCO, GREEN #1
LOC'N: SEC. 31-T10N-R26W
COUNTY: BECKHAM
FIELD: WEST MAYFIELD
STRATIGRAPHY/
DEPTH
HOST ROCK FEATURES
LITHOLOGY
FACIES
SEDIMENTARY STRUCTURES
TEXTURE
MUDSTONE
WACKESTONE
PACKSTONE
GRAINSTONE
BOUNDSTONE
COLOR
LT DK
TAN
GREY
ALLOCHEMS
NONSKELETAL
BIOCLASTIC
GRAIN SIZE
.5 mm
1 mm
2 mm
5 mm
TERRIGENOUS
QUARTZ
CLAYS
rca
M
HENRYHOUSE
19700
19720
19740
19760
19780
I
IIa



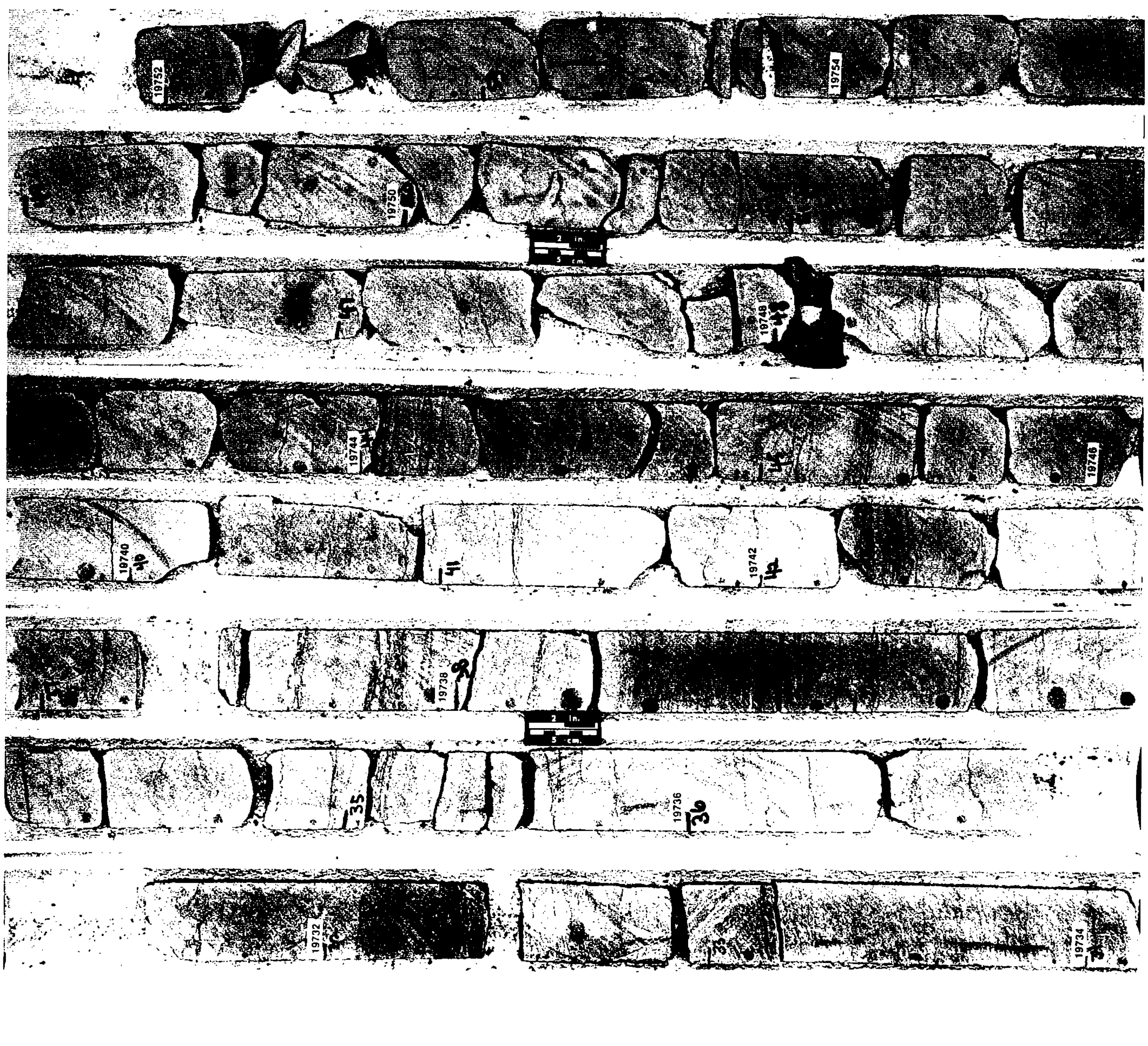

19620
19622
19624
19626
19630
19632
19634
19636
19638
19640
19642
2 in
5 cm




19682
19686
19688
19690
19692
19694
19696
19698
19700
19702
19704


19706
19710
19712
19714
19716
19718
19724
19726
19728
19730


19768
19770
19766
2 in.
5 cm.
19760
19756
19758


19752
19754
19750
2 in.
5 cm.
19748
19744
19746
19740
19742
19738
2 in.
5 cm.
19736
19732
19734

Core: W.C. Pickens Cronkite No. 1

Location: sec. 14, T. 15 N., R. 5 W., Kingfisher County, OK

Cored Interval: 7,098 to 7,136 ft

Stratigraphic Unit: Henryhouse Formation (Hunton Group)

Core Description: The following description is given in reference to the core petrolog and photograph immediately following this text.

<u>7,098 to 7,108 ft</u>

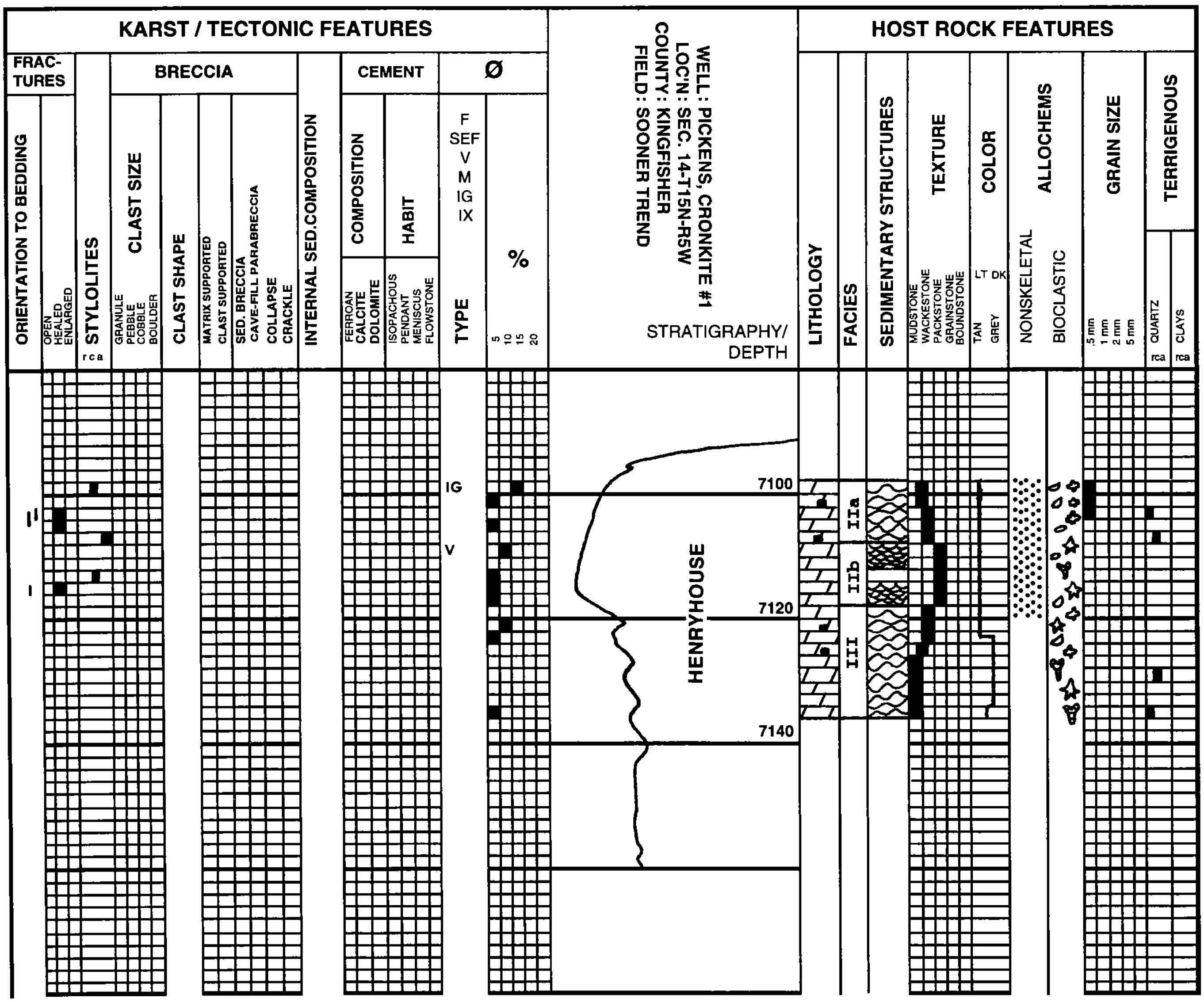
This interval contains brownish-gray, occasionally silty dolo-mudstone/wackestone. Several fossil types are present in the core, including pelmatozoans, brachiopods, trilobites, and ostracods. Whereas burrows are abundant, there are very few stylolites or fractures. This interval is believed to represent facies IIa (lagoonal subfacies).

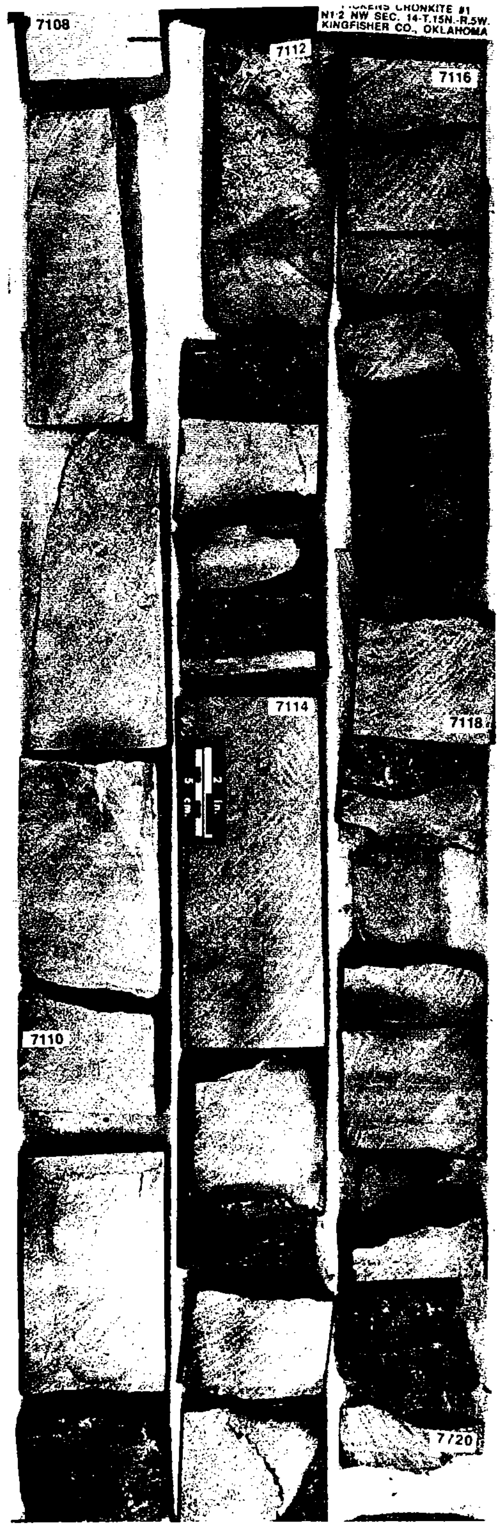
<u>7,108 to 7,118 ft</u>

Underlying the lagoonal subfacies is a sharp contact that separates a medium-gray and brownish-gray, limy to dolomitic oolitic packstone. A few fossils, namely pelmatozoans, trilobites, ostracods, and bryozoans, are observed in this interval. These rocks, which show low-angle cross-bedding structures, were most likely deposited in a proximal subtidal to a moderate-energy intertidal environment.

<u>7,118 to 7,136 ft</u>

This interval is also separated by sharp contact from the preceding one. It consists of a medium-dark-gray, slightly silty, dolomitic mud/wackestone. Fossils include pelmatozoans, brachiopods, and rare bryozoans and trilobites. The rocks are very slightly burrowed. The environment is interpreted as low-energy subtidal.




N1/2 NW SEC. 14-T.15N.-R.5W.
KINGFISHER CO., OKLAHOMA
7108
7112
7116
7114
7118
7110
7120
2 in
5 cm

Reservoir Characteristics of an Upper Hunton Gas-Producing Zone, Southwest Ringwood Area of Major County, Oklahoma

Charles E. Mear
Independent Geologist
Austin, Texas

Keith A. Hutton
Cross Timbers Oil Co.
Fort Worth, Texas

ABSTRACT.—An upper Hunton reservoir in the southwest Ringwood area is composed of two distinct porosity zones. The porosity zones are referred to as the Hunton "C" and "D"; the Hunton "C" is the deeper of the two zones. The Hunton "C"porosity zone is composed of fine- to medium-crystalline dolostone that contains silt, mica, illite, and microporosity. Log analyses indicate that the zone should be water-productive, but it produces gas with little or no water. These log calculations are affected by the low resistivity measurements caused by the irreducible water present in the micropores and in the illite clay. The upper porosity zone, Hunton "D," is a clean dolostone that contains vugular and intercrystalline porosity along with microporosity. Core, geophysical-log analyses, and pressure-transient analyses indicate that both porosity zones average ~12% porosity and 0.10 millidarcy (md).

PURPOSE

This paper presents a description of a Hunton-age gas-producing reservoir, in which much of the production is from tight microporosity that calculates to be water-bearing by geophysical-log analyses.

STRATIGRAPHY OF THE UPPER HUNTON RESERVOIR

The Southwest Ringwood area is located on the northern shelf of the Anadarko basin (Fig. 1), in parts of Ts. 20–21 N., R. 11 W., Major County, Oklahoma (Fig. 2). During Silurian and Devonian time, the area was located on a shallow-marine shelf where carbonate deposition predominated.

Although the Hunton Group rocks have been divided into several formations in Oklahoma, no attempt was made to define formations during this study, due to a lack of paleontological data. Based on its position in the section, it is probable that the upper Hunton reservoir is Devonian in age.

In the area of interest, the Hunton Group consists of 240–300 ft of limestone and dolostone beds that probably were deposited as calcite mud. Some of the calcite mud

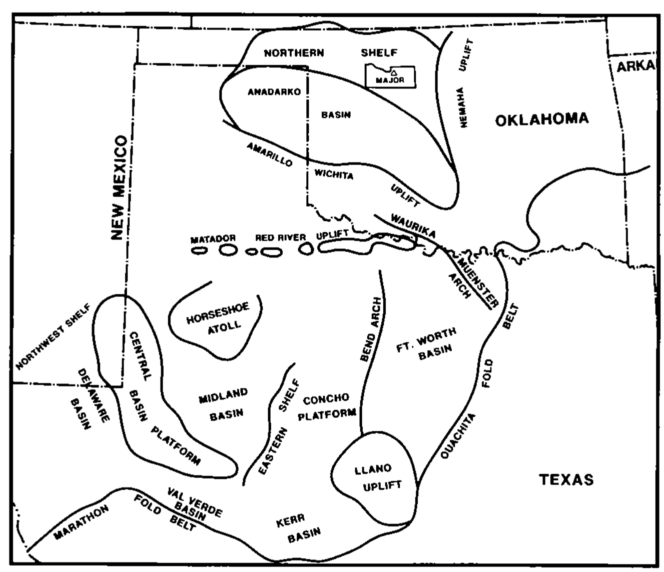


Figure 1. Regional tectonic setting of Major County, Oklahoma, in north-central part of map.

Mear, C. E.; and Hutton, K. A., 1993, Reservoir characteristics of an upper Hunton gas-producing zone, southwest Ringwood area of Major County, Oklahoma, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 41–44.

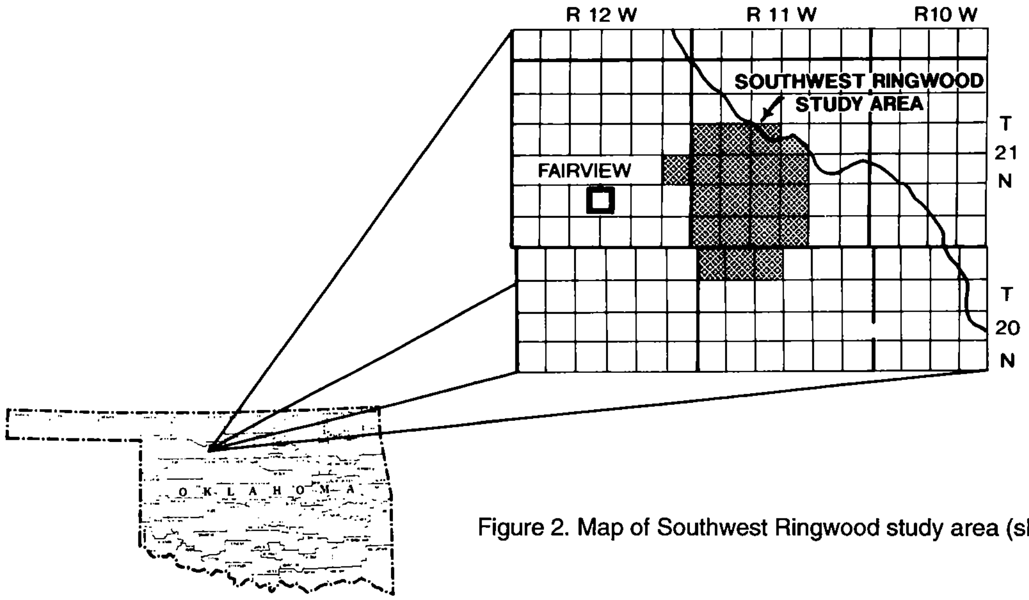


Figure 2. Map of Southwest Ringwood study area (shaded).

was altered to fine- to medium-crystalline dolostone shortly after deposition.

Following deposition of the Hunton, the northern shelf area was uplifted, and the Hunton rocks were exposed to erosion. Later, in Devonian time, ~50 ft of Woodford Shale was deposited on the truncated Hunton. Following deep burial in Mississippian through Permian time, the Woodford generated oil and gas that migrated into the porous upper Hunton.

RESERVOIR DESCRIPTION

The upper Hunton reservoir in the area of interest attains a maximum net thickness of 66 ft, which is divided into two zones of porosity contained in dolostones of different composition. In ascending order, these are called informally the Hunton "C" and Hunton "D" in this discussion.

A core of the Hunton "C" porosity zone from the Cross Timbers Oil Co. No. 4 Rankin well, SW¼SW¼ sec. 17, T. 21 N., R. 11 W., reveals that the zone is light-brown, bioturbated, finely crystalline dolostone (Fig. 3). It contains as much as 20% quartz as silt grains, 5–10% clay, and varying amounts of calcite. As shown by SEM analysis (Fig. 4), the clay is largely illite, although some muscovite is present. Solid hydrocarbon is present as a coating in the intercrystalline pore space. Much of the porosity cannot be seen at <50× magnification (it is microporosity).

Core analysis of the Hunton "C" porosity zone in the Rankin No. 4 well shows that the zone has an average porosity of 12% and an arithmetic mean permeability of 0.095 md. One foot of the core has 0.36 md, and half of the pay zone has <0.10 md permeability.

Geophysical wireline logs of the Hunton "C" porosity zone (Fig. 5) corroborate the core measurements. Gamma-ray logs show that the zone is more radioactive than the underlying dense dolomitic limestone.

Based on analyses of drill cuttings, the Hunton "D" porosity zone is described as light-brown, fine- to medium-crystalline dolostone that contains intercrystalline and vug porosity. The zone contains little silt or clay and is less radioactive than the Hunton "C" zone. The Hunton "D"

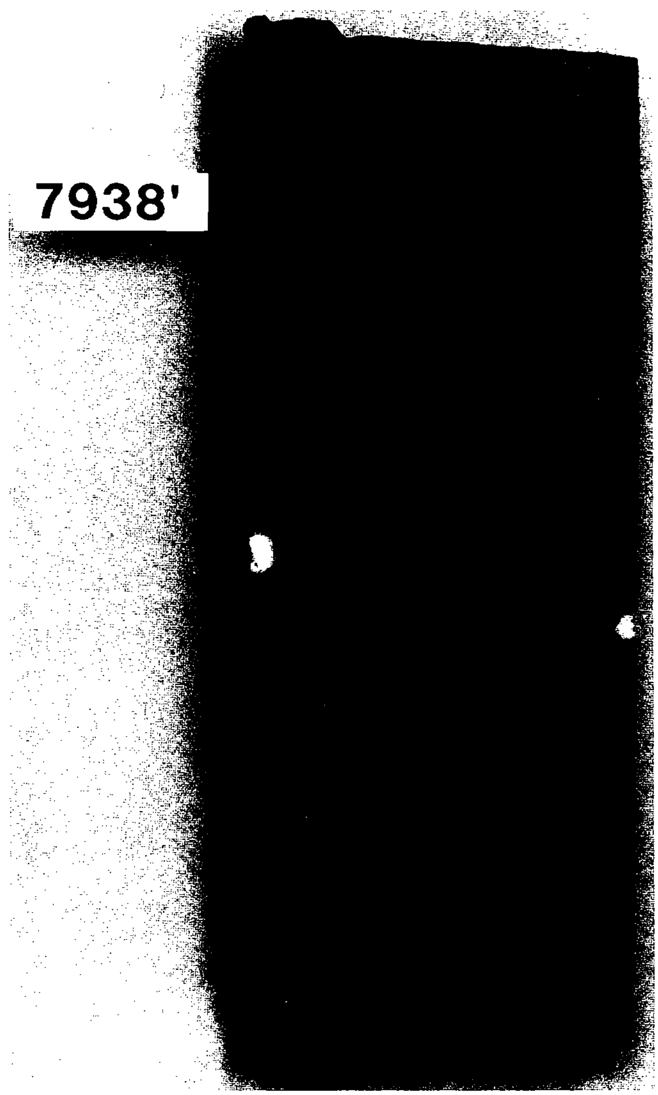


Figure 3. Photograph of core from the Hunton "C" zone, Cross Timbers No. 4 Rankin, at a core depth of 7,938 ft (log depth of 7,950 ft).

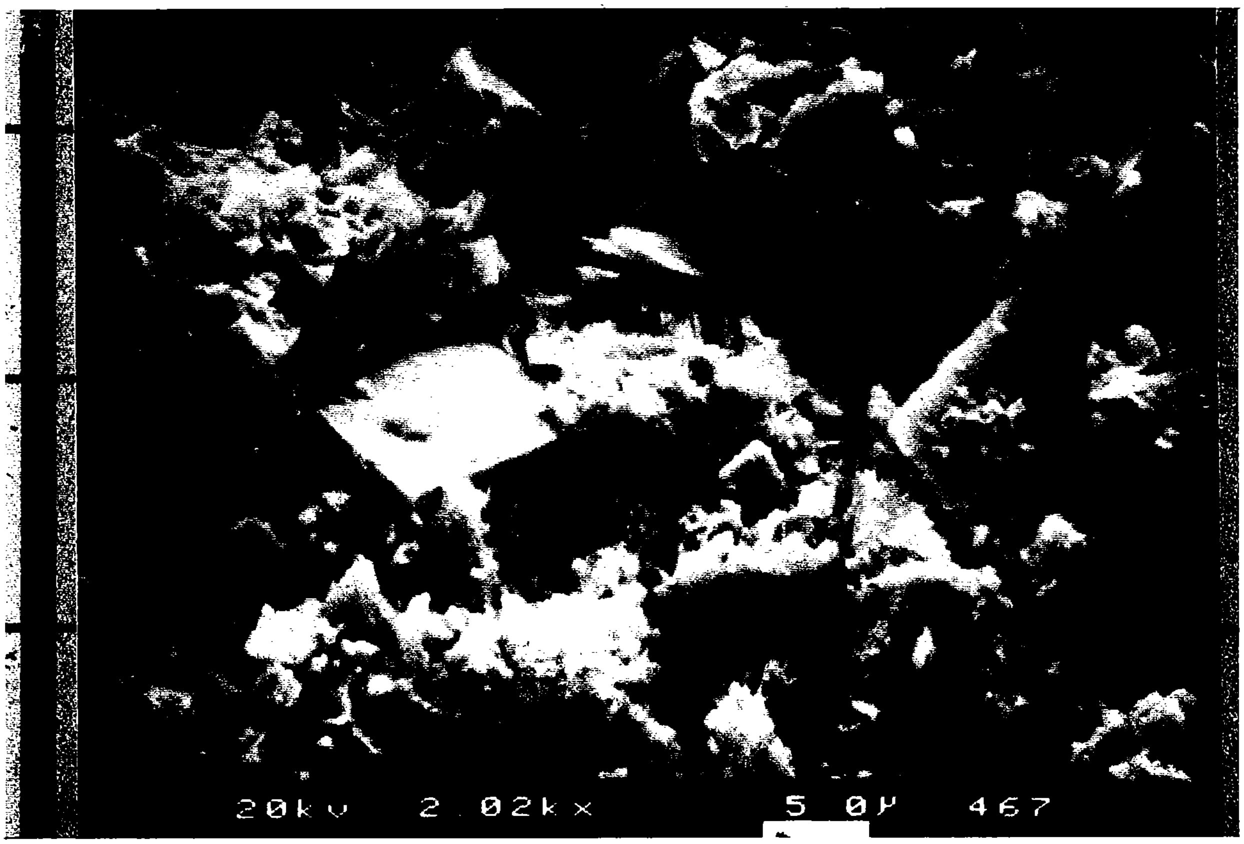


Figure 4. Scanning electron microscope (SEM) image of Hunton "C" zone dolostone in the Cross Timbers No. 4 Rankin (enlarged 2000×); core depth 7,940 ft (log depth of 7,952 ft). Dolostone consists mostly of euhedral rhombs with authigenic-illite fibers partly filling intercrystalline porosity.

zone has ~12% porosity, based on wireline analyses, and an average permeability of <0.10 md, based on pressure-transient analyses.

The Hunton "C" porosity zone has low resistivity throughout the area of interest, and log calculations indicate that the zone should be water-productive. The water detected by the geophysical logs is irreducible water present in micropores and in the illite clay. Thus, the zone produces water-free gas in the Southwest Ringwood area.

Production is from a simple stratigraphic trap, in which the reservoir is truncated in the updip northeast direction and grades laterally to dense, clean dolostone (Mear and Hutton, 1992). The trap and source of the gas is the overlying Woodford Shale.

Logs which measure rock mechanical properties were run to determine the vertical extent of induced artificial fractures in wells in the area, and they show that there is very little stress contrast between the Hunton and the overlying Woodford Shale and Mississippian limestones. This indicates that a high-rate fracture stimulation of the Hunton reservoir will extend the fractures into the overlying Mississippian limestones, which produce oil, gas, and water in the area. When communication occurs between zones, the water from the Mississippian highly degrades the productiveness of the Hunton. This degradation is caused by the imbibition of water into the micropore system, along with illite-clay movement; these two actions block the pore throats. Since the zone must be artificially fractured in order to produce at commercial rates, care must be taken to avoid fracturing into water-bearing zones. Also, the fracture fluid should contain as little water as possible.

DISCUSSION

Core of the upper Hunton in the area of interest was obtained in order to determine why the producing zone was calculated to be water-productive and which treatment of the pay zone should be used. No study of diagenesis of the dolostone was undertaken.

The upper Hunton dolostone reservoir discussed in this paper is unusual in that it contains much silt and illite and some muscovite. It is assumed that the silt and muscovite were carried into the Hunton depositional area by wind storms. Isopachous maps of the Hunton indicate that the porous Hunton "C" zone may be related to shoal areas present during deposition of calcite mud, silt, and muscovite. The calcite mud and muscovite subsequently were altered to dolostone and illite, respectively, but no clues as to how this happened were obtained during this study.

REFERENCE CITED

Mear, C. E.; and Hutton, K. A., 1992, Petroleum geology of an unconventional upper Hunton Group reservoir, Southwest Ringwood area, Major County, Oklahoma: Oklahoma Geology Notes, v. 52, p. 164–172.

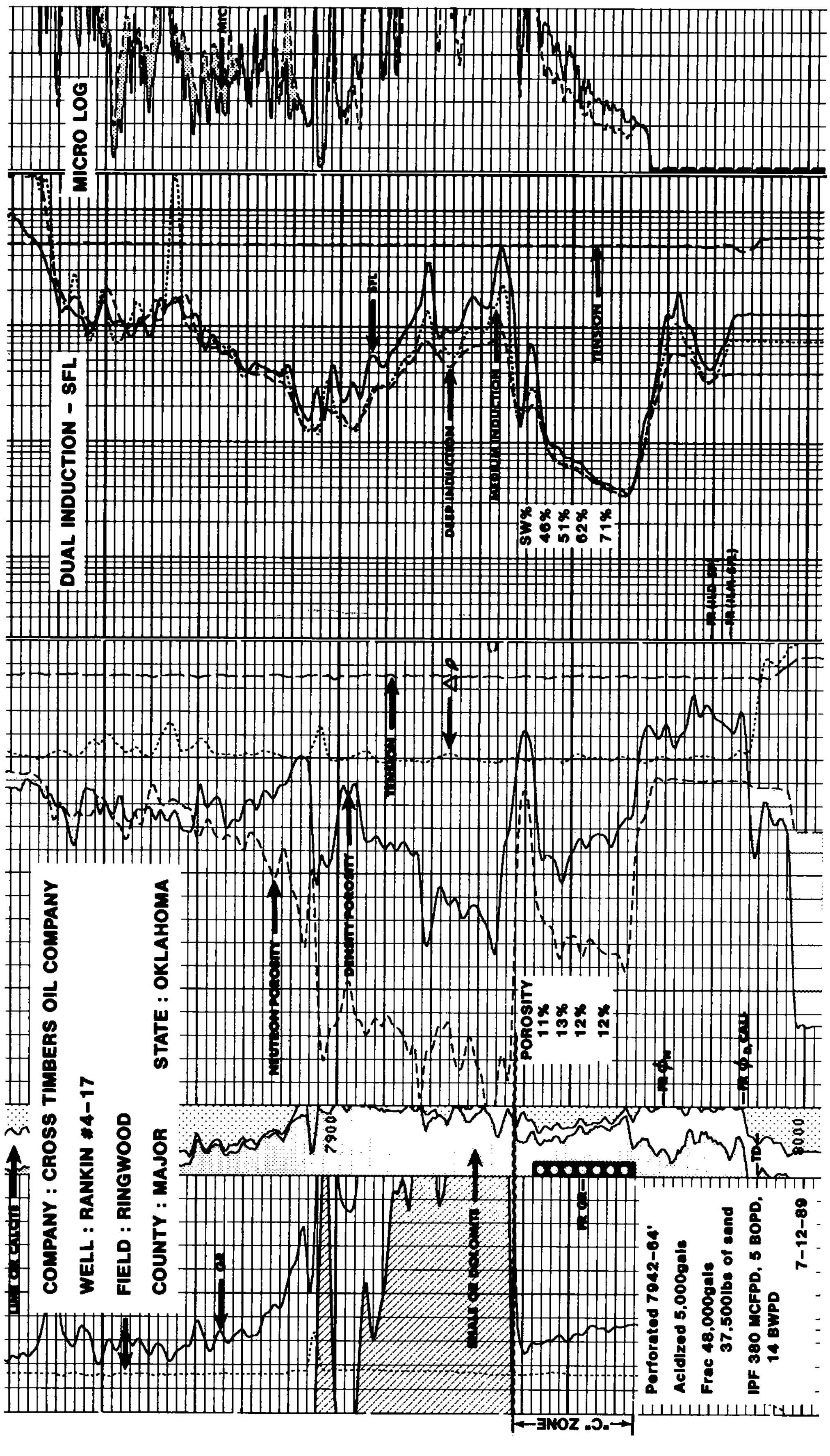


Figure 5. Wireline logs of the Hunton "C" zone in the Cross Timbers No. 4 Rankin well.

Petrophysical Characteristics (Based on One Core) of Lower Paleozoic Hunton Dolostones, Woods County, Northern Anadarko Basin, Oklahoma

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Brooklyn College and Graduate School of the City University of New York
and Northeastern Science Foundation, Rensselaer Center of Applied Geology
Troy, New York

ABSTRACT.—Petrophysical analysis of core from a well in Woods County, Oklahoma, near the interpreted paleoshoreline of the northern part of the Anadarko basin, revealed the presence of three petrofacies. Petrofacies 1, which is a vuggy, porous and permeable, medium-crystalline, hypidiotopic dolostone, includes most of the section in the studied core. It is generally characterized by concave capillary-pressure curves, large and variable median pore-throat sizes, moderate to high porosities, and low to moderate recovery efficiency. Petrofacies 2 is composed of finely crystalline, dolomitized, skeletal packstones and stromatolites; it yields a bimodal to gently sloping capillary-pressure curve and has small median pore throats, variable porosities, and high recovery-efficiency values. Petrofacies 3 is a coarsely crystalline, dolomitized oolitic grainstone (Keele Oolite) having oomoldic and vuggy porosity; it exhibits bimodal to gently sloping capillary-pressure curves. Its small median pore throats, low porosities, and low recovery-efficiency values make this petrofacies a poor reservoir. Petrofacies 1 and 2 are considered to be the best reservoirs because they have both high porosities and generally moderate to high recovery-efficiency values.

INTRODUCTION

The purpose of this paper is: (1) to review some of our research group's previous work on petrofacies in the Hunton Group of the Anadarko basin; and (2) to report on a core which serves as an example of Hunton Group petrophysics. This core has been studied previously (Amthor and others, 1988; Kopaska-Merkel and Friedman, 1989), but this report updates some of the previous findings.

GEOLOGIC SETTING

The Anadarko basin (Fig. 1) is an asymmetrical, west-northwest-trending remnant of the southern Oklahoma aulacogen (Hoffman and others, 1974; Pruatt, 1975; Sternbach and Friedman, 1986). Within the Anadarko basin, the Hunton Group is a sequence of limestones, dolostones, and lesser siliciclastics of Late Ordovician through Early Devonian age (Amsden, 1960,1975,1980; Beardall, 1983; Way, 1983; Medlock, 1984). The Hunton is as much as 1,300 ft thick.

The Hunton Group conformably overlies the Sylvan Shale (Upper Ordovician) (Fig. 2), and throughout most of the basin is overlain unconformably by either the Woodford Shale (Upper Devonian–Lower Mississippian) or lo-

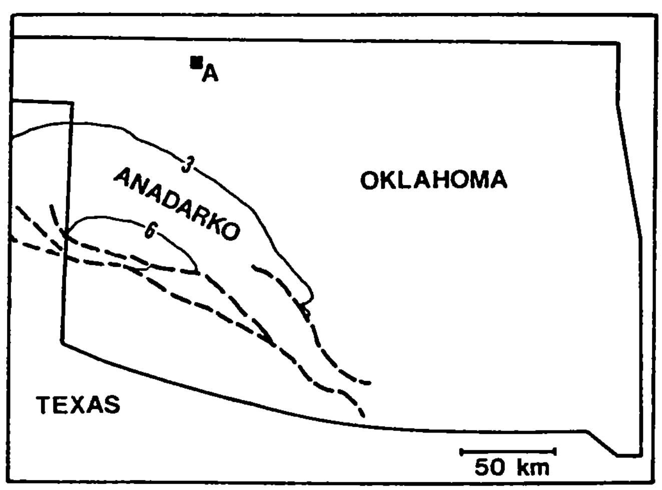

Figure 1. Location map showing Anadarko basin with major bounding down-to-the-north normal faults indicated by dashed lines, and 3- and 6-km (about 2- and 4-mi) structural contours for top Hunton Group indicated by solid lines. Note location of Calvert Mid-American Bloyd No. 2 well, Woods County, Oklahoma, designated by the letter A, whose core was examined in this study.

Friedman, G. M., 1993, Petrophysical characteristics (based on one core) of lower Paleozoic Hunton dolostones, Woods County, northern Anadarko basin, Oklahoma, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 45–51.

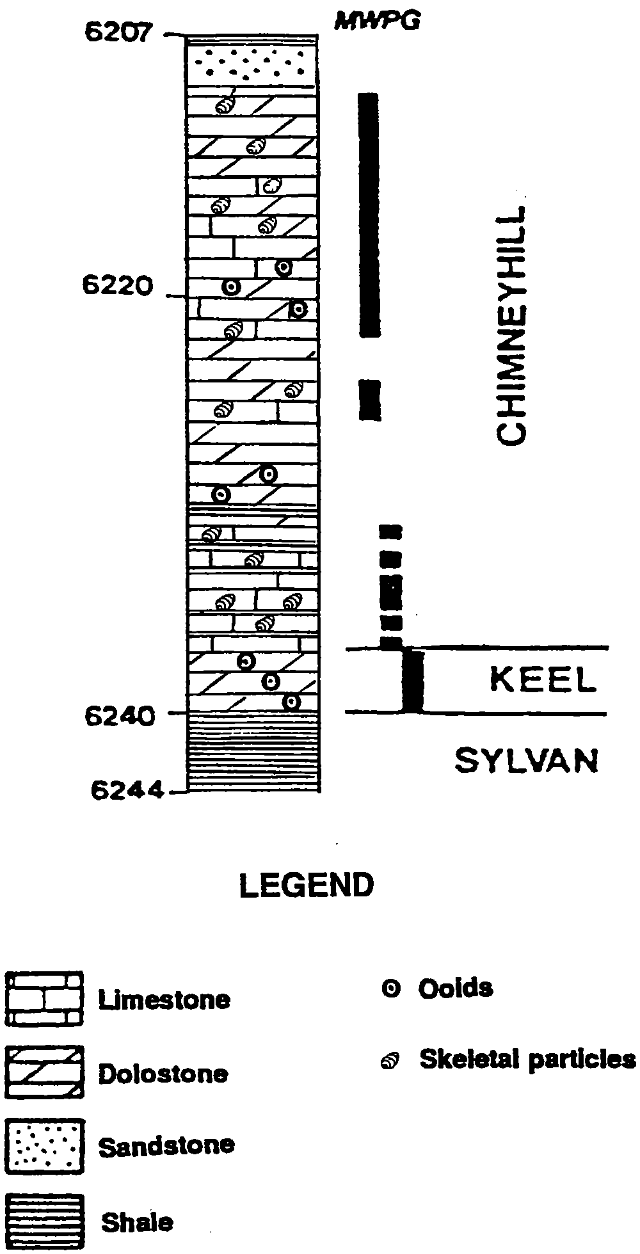

Figure 2. Core log of Calvert Mid-American Bloyd No. 2 well, examined in this study. MWPG = mudstone, wackestone, packstone, and grainstone. Depth in feet.

cally by the Misener sandstone (Middle Devonian–Upper Devonian). The depositional extent of Hunton units was severely modified by erosion at intervals during deposition of the Hunton Group, especially during the Early Devonian. Extensive erosion also occurred during the Middle Devonian, preceding deposition of the Misener sandstone and Woodford Shale.

The lower part of the Hunton, the Chimneyhill Subgroup, is commonly recognized in the subsurface, as is its basal unit, the Keel Formation (Fig. 2). The Keel is a thin (1–33 ft thick) ooid grainstone that conformably overlies the marine Sylvan Shale and retains its distinctive character throughout the Anadarko basin (Amsden, 1960,1975, 1980). Because of its lithologic homogeneity and widespread distribution within the study area, the Keel is an excellent marker bed. The Keel is commonly dolomitized at depths below 10,000 ft, where porosity is dominantly intercrystalline (Sternbach and Friedman, 1984,1986). At shallower depths, the Keel is only lightly dolomitized, and as much as 2 or 3% microporosity is typical. The upper part of the Chimneyhill Subgroup, not subdivided in subsurface sections, was deposited upon an unconformity surface that formed after deposition of the Keel (Amsden, 1960; Manni, 1985). Above the basal Keel Formation, the Chimneyhill Subgroup consists of crinoidal wackestones, packstones, and grainstones, which may be cherty and glauconitic, or argillaceous. In places, dolomitization (commonly associated with unconformities within the sequence) and formation of fossomoldic porosity in the upper Chimneyhill have produced excellent reservoir rock. Elsewhere, nodular and pervasive chertification have destroyed nearly all porosity (Amsden, 1960,1975,1980; Manni, 1985).

PETROFACIES OF HUNTON CORE FROM NORTHERN (SHALLOW) PART OF ANADARKO BASIN

Petrology

I have selected for study a core from the Calvert Mid-American Bloyd No. 2 well, because in this well the Hunton Group is encountered at shallow depth (~6,200 ft); the well is located in NW¼ sec. 21, T. 27 N., R. 15 W., Woods County, Oklahoma (Fig. 1).

This well is located near the interpreted paleoshoreline in the northern part of the Anadarko basin. Supratidal dolostones occur in this area together with algal stromatolites. These supratidal dolostones are non-ferroan. The dolomite is a distinct supratidal facies found only in the extreme north (north of the "Dolomite Front" described by Amsden, 1975).

Figure 2 shows the core log for the well studied. Overlying the Sylvan Shale is the Keel Formation, a dolomitized oolitic grainstone (Fig. 3) with oomoldic and vuggy porosity and with locally abundant skeletal fragments, mostly those of echinoderms (petrofacies 3). The rock is a coarsely crystalline, xenotopic dolostone in which both vertical and horizontal stylolites are present. In thin sections the ooids are commonly only recognized as ghosts (Fig. 3). In places, dissolution, as a result of subaerial emergence, created solution-collapse breccia in which fractures are lined with saddle dolomite and/or filled with clay minerals. In the overlying section of the undifferentiated Chimneyhill Subgroup, dolomitized skeletal packstone and dolomitized stromatolites compose the facies (Fig. 4) (petrofacies 2). This dolomite is finely crystalline; larger euhedral dolomite crystals make this fabric bimodal. In places, stylolites form the boundary between stromatolites and skeletal packstone, suggesting extensive subsurface dissolution. Porosity is mostly vuggy; locally, saddle dolomite lines the vugs. Sporadic spherical aggregates of pyrite are present. This fabric is slightly mottled in places; mottles are filled with euhedral dolomite crystals and contain small vugs. The rest of the cored section above petrofacies 2 is composed of medium-crystalline, porous and permeable, hypidiotypic dolostone with vuggy porosity in which the dep-

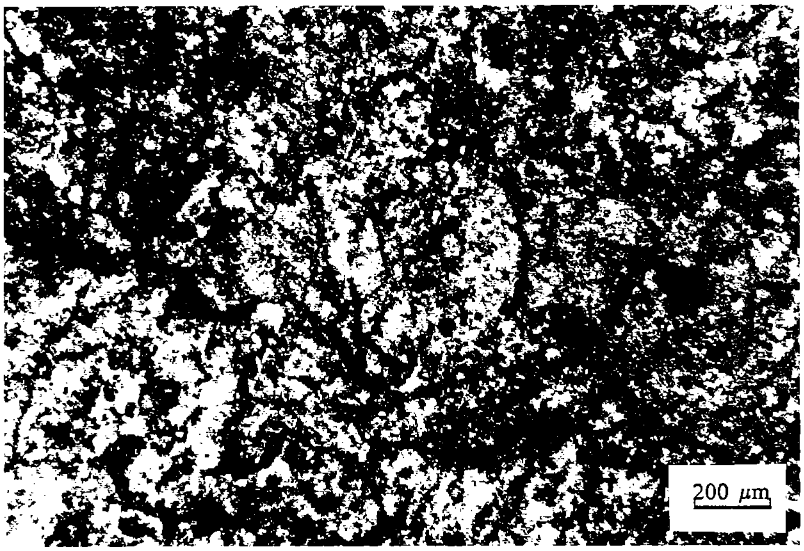

Figure 3. Photomicrograph of thin section of coarsely crystalline, dolomitized oolitic grainstone (Keele Oolite) of petrofacies 3. Note ghosts of ooids. Calvert Mid-American Bloyd No. 2 well, 6,237 ft deep.

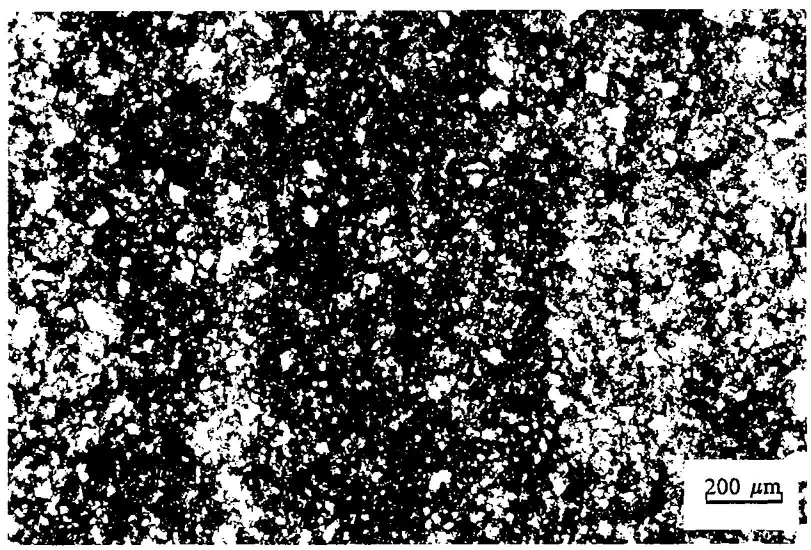

Figure 4. Photomicrograph of thin section of finely crystalline, dolomitized skeletal packstone and stromatolites of petrofacies 2. Note sporadic euhedral dolomite crystals. Calvert Mid-American Bloyd No. 2 well, 6,231 ft deep.

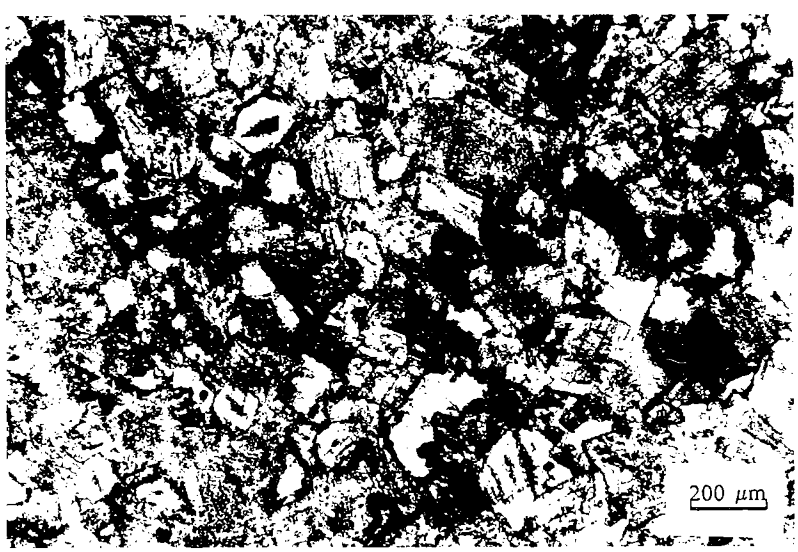

Figure 5. Photomicrograph of thin section showing porous, permeable, medium-crystalline, hypidiotopic dolostone of petrofacies 1. The depositional texture in this dolostone has been erased. This dolostone is quite vuggy, but no vugs are present in this field of view. Note concentration of organic matter or dead oil. Calvert Mid-American Bloyd No. 2 well, 6,224 ft deep.

ositional textures have been obliterated (petrofacies 1) (Fig. 5). Petrofacies 4 and 5 of Kopaska-Merkel and Friedman (1989) have not been recognized in this core.

PETROPHYSICAL CHARACTERISTICS OF PETROFACIES

The technique of petrophysical analysis has been presented elsewhere (Ghosh and Friedman, 1989; Kopaska-Merkel and Friedman, 1989; Friedman and others, 1990) and involves mercury porosimetry. This technique measures the volume distribution of pore throats. Pore throats constitute only a small fraction of the total pore-system volume. However, most mercury-porosimetry studies give information on only the pore throats (Wardlaw, 1976), because the method depends on the forcing of mercury into the small voids within the rock. The smallest voids (throats) control access to the larger ones (pores) because higher pressures are needed to force mercury or other nonwetting fluids into smaller spaces (for principles of capillarity, see Purcell, 1949). The pore throats are the bottlenecks in the pore system; their critical capillary pressures must be exceeded for mercury or other nonwetting fluids to enter the pores they surround. For a sample with a spatially homogeneous pore system, the total volume of mercury intruded at a given pressure is precisely related to the number of throats of corresponding size. Thus, a graph of cumulative mercury-intrusion volume vs. capillary pressure is equivalent to cumulative mercury-intrusion volume vs. pore-throat size. Mercury porosimetry is the only analytical method that yields accurate and quantitative information about the size distribution of pore throats.

The model used in this paper is that of Wardlaw (Wardlaw, 1976; Wardlaw and Taylor, 1976) for crystalline dolostones in which throats are assumed to be sheetlike and pores are either tetrahedra or are polyhedra that approach the tetrahedral form during progressive dolomitization (Wardlaw, 1976, fig. 12). For further details, see Kopaska-Merkel and Friedman (1989).

Three key parameters have been used in this study and need to be defined: (1) porosity, (2) median throat size, and (3) recovery efficiency.

Porosity: the term porosity is used in this paper for the porosity calculated from mercury-porosimetry data ([pore volume/bulk volume] × 100) on the assumption that 100% of pores are filled at the maximum pressure attained in the capillary-pressure analysis (20,000 psia or 138 MPa). This assumption has been used for Hunton Group carbonates (Amthor and others, 1988; Kopaska-Merkel and Friedman, 1989).

Median throat size: the median or critical throat size, related in the literature to threshold pressure (Ghosh and

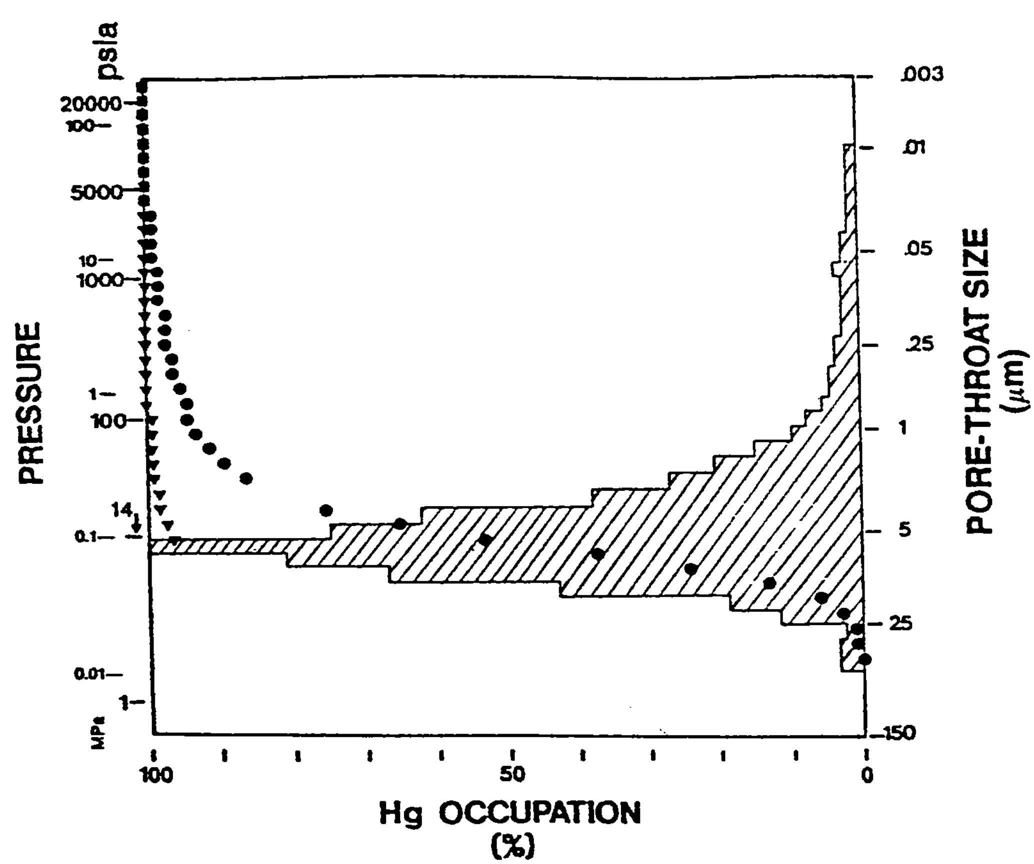

Figure 6. Capillary-pressure curve of sample of petrofacies 1 (vuggy, porous and permeable, medium-crystalline, hypidiotopic dolostone). Median pore-throat size is large; recovery is negligible. Entry pressure is well defined and extremely low, ~3 psia (0.20 MPA). Getty 1 Leutmeyer U. well, Blaine County, Oklahoma (south of Calvert Mid-American Bloyd No. 2 well), 8,838 ft deep (after Kopaska-Merkel and Friedman, 1989).

Friedman, 1989), is the capillary pressure at which maximum intrusion of mercury occurs. It is identified on capillary-pressure curves as a near-horizontal line or break in slope.

Recovery efficiency (RE): RE is the percentage of trapped hydrocarbons which may be withdrawn from a reservoir rock under a particular set of conditions. With respect to experiments in mercury porosimetry, this definition is modified slightly. Recovery efficiency in mercury porosimetry is the percentage or fraction of mercury intruded at maximum pressure which is extruded during pressure reduction to final minimum pressure (usually 14–15 psia [0.09–0.1 MPa], or atmospheric pressure).

Petrofacies 1

This petrofacies (vuggy, porous and permeable, medium-crystalline, hypidiotopic dolostone) (Fig. 5) includes most of the dolostones in the studied core, and in fact most of the Hunton Group. It is generally characterized by concave cumulative intrusion capillary-pressure curves (Fig. 6), large and variable median pore-throat sizes, moderate to high porosities, and generally low to moderate recovery efficiency (Figs. 7,8). The median throat size of this petrofacies is more variable than that of the other petrofacies, probably as a result of variability in depositional environment and diagenetic overprint.

Petrofacies 2

This petrofacies (finely crystalline, dolomitized skeletal packstones and stromatolites) is characterized by bimodal to gently sloping cumulative intrusion capillary-pressure curves, small median pore throats, variable porosities, and high recovery-efficency values (Figs. 7–9).

Petrofacies 3

This petrofacies (coarsely crystalline, dolomitized oolitic grainstone [Keele Oolite], oomoldic and vuggy porosity) is composed of samples that are petrophysically homogeneous. These samples have bimodal to gently sloping cumulative intrusion capillary-pressure curves, small median pore throats, low porosities, and low recovery-efficiency values (Figs. 7–10). This petrofacies makes a poor reservoir because of small median pore throats, low porosities, and low recovery efficiency.

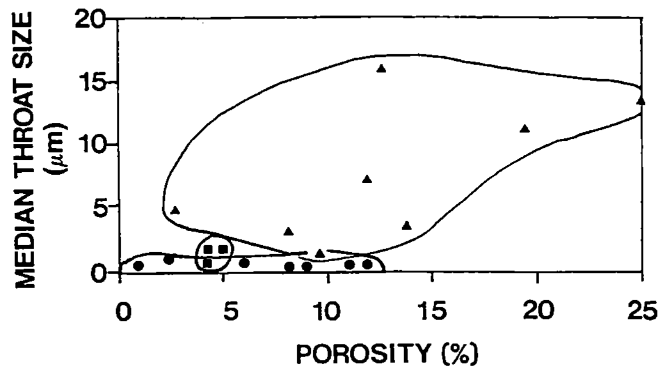

Figure 7. Bivariate plot of median pore-throat size vs. porosity for petrofacies 1, 2, and 3 in Calvert Mid-American Bloyd No. 2 well; depth of samples, 6,216–6,240 ft. ▲ = petrofacies 1 (vuggy, porous and permeable, medium-crystalline, hypidiotopic dolostone); ● = Petrofacies 2 (finely crystalline, dolomitized skeletal packstone and stromatolites); ■ = petrofacies 3 (coarsely crystalline, dolomitized oolitic grainstone [Keele Oolite], oomoldic and vuggy porosity).

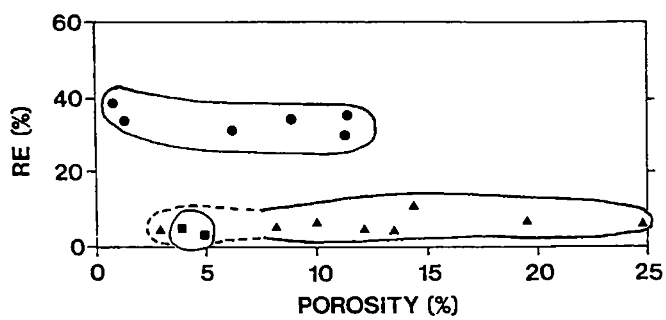

Figure 8. Bivariate plot of recovery efficiency (RE) vs. porosity for petrofacies 1, 2, and 3, Calvert Mid-American Bloyd No. 2 well; depth of samples, 6,216–6,240 ft. ▲ = petrofacies 1 (vuggy, porous and permeable, medium-crystalline, hypidiotopic dolostone); ● = petrofacies 2 (finely crystalline, dolomitized skeletal packstone and stromatolites); ■ = petrofacies 3 (coarsely crystalline, dolomitized oolitic grainstone [Keele Oolite], oomoldic and vuggy porosity).

Discussion

The best reservoir rocks analyzed in this study are those of petrofacies 1 and 2, because they have both high porosities and generally moderate to high RE values. RE in mercury porosimetry is analogous to primary recovery of petroleum from natural reservoirs because both processes involve only simple pressure reduction. Thus, rocks with high RE values produce relatively high percentages of contained nonwetting fluids (such as oil) without recourse to enhanced-recovery techniques.

Some reservoir rocks are oil-wet. For these rocks, capillary-pressure data cannot be interpreted as described in this paper.

The petrofacies, as here defined, form mappable subsurface units. Petrofacies distribution is markedly discordant to depositional lithofacies distribution and to stratigraphy, because diagenetic processes (including dolomitization, chertification, and calcite dissolution) have acted partially independently of depositional facies. The upper Chimneyhill in this well is dolomitized and is assigned to petrofacies 1, whereas this same interval in other wells is silicified and is assigned to petrofacies 4 of Kopaska-Merkel and Friedman (1989).

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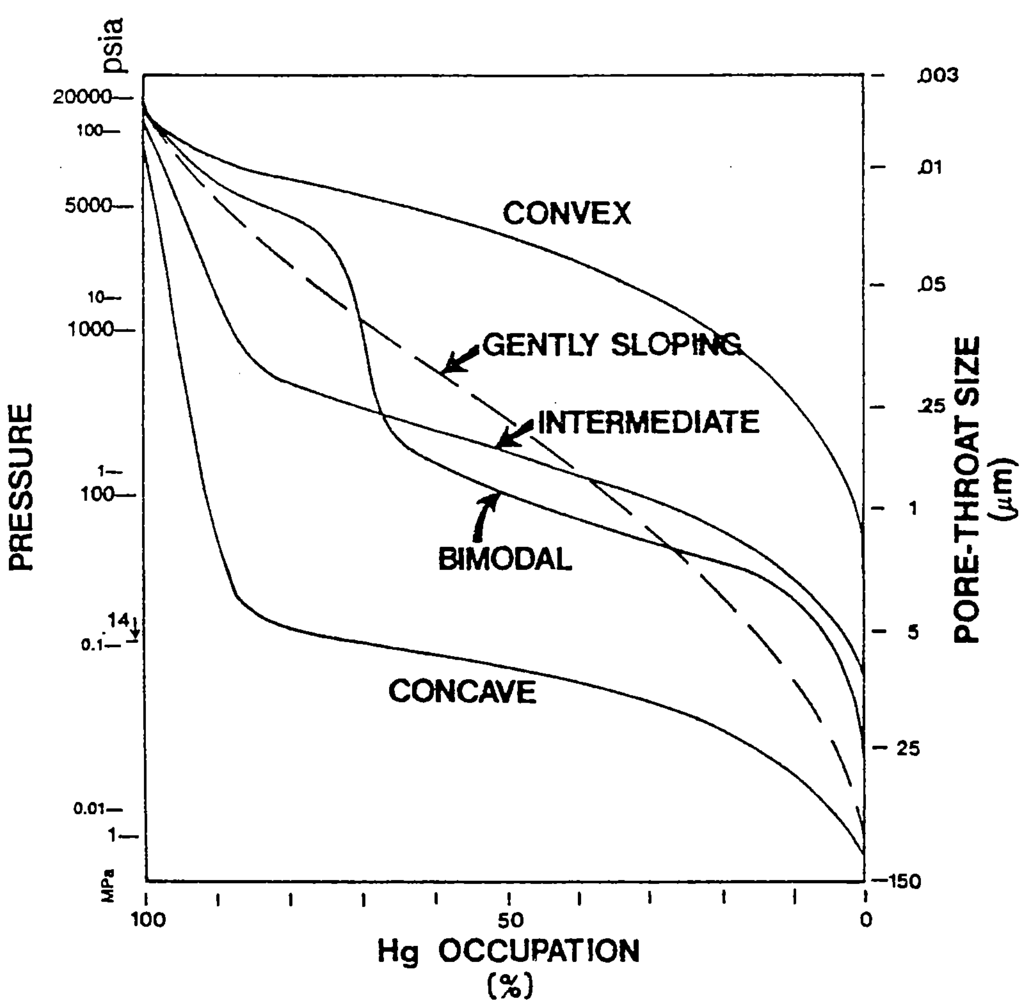

Figure 9. Schematic capillary-pressure diagram showing five different shapes of cumulative intrusion capillary-pressure curves. Petrofacies 1 is generally characterized by concave capillary-pressure curve, and petrofacies 2 and 3 by bimodal to gently sloping curves (after Kopaska-Merkel and Friedman, 1989).

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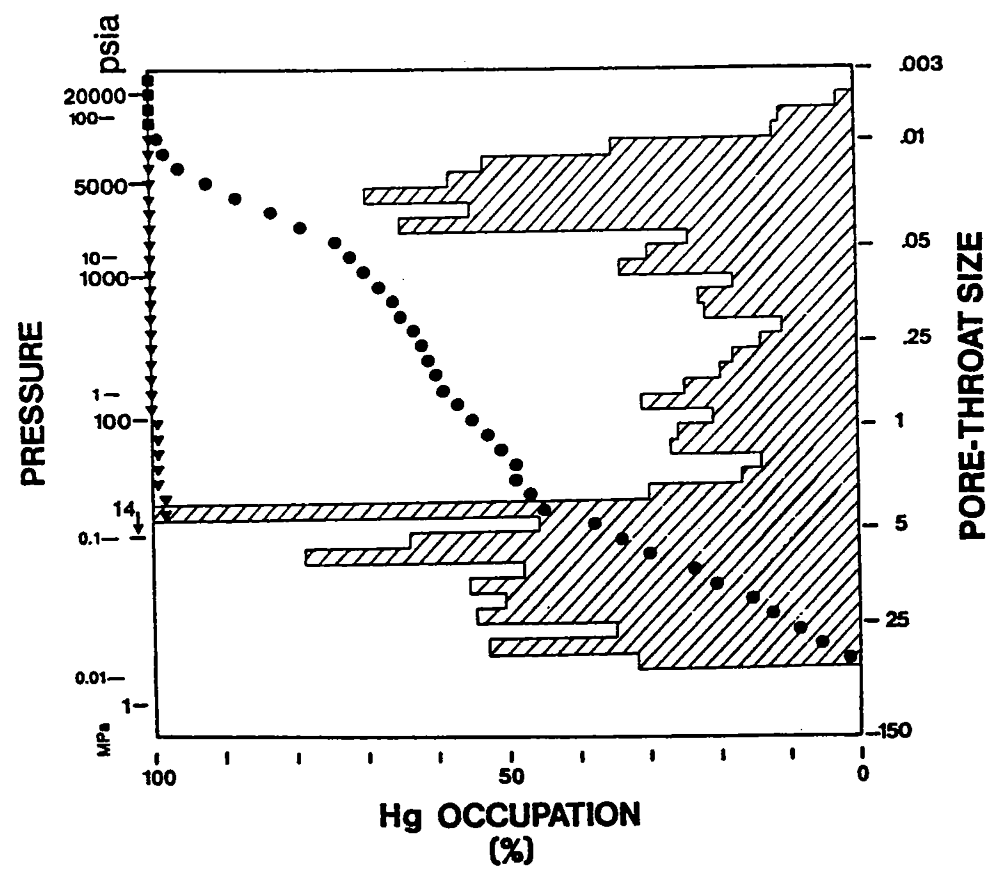

Figure 10. Capillary-pressure curve of sample of petrofacies 3 (coarsely crystalline, dolomitized oolitic grainstone [Keele Oolite], oomoldic and vuggy porosity). Recovery is negligible; entry pressure is undefined. Calvert Mid-American Bloyd No. 2 well, 6,239 ft deep (after Kopaska-Merkel and Friedman, 1989).

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Outcrop and Subsurface Evidence for Karsted Reservoirs in the Fusselman Formation (Silurian), Permian Basin, Texas

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ABSTRACT.—Outcrops of Fusselman paleokarst in the Hueco Mountains (West Texas), and analogous occurrences in the subsurface Permian basin, attest to the control of karstification on reservoir development. Many Fusselman hydrocarbon reservoirs are structural and/or combination structural/stratigraphic traps associated with paleocaves. Other reservoirs are related to the occurrence of porous carbonate strata in buried-hill (paleotopographic) traps beneath unconformities within and at the top of the Fusselman.

INTRODUCTION

Fusselman (Silurian, Alexandrian–Niagaran) hydrocarbon reservoirs in the Permian basin of West Texas–New Mexico come in many "shapes and sizes." Insofar as most Fusselman producing fields here were discovered by seismic methods, most are considered to be structural traps, and, locally, combination structural/stratigraphic traps developed along the regional Fusselman erosional truncation (L. J. Mazzullo and others, 1989; L. J. Mazzullo, 1990a,b). It was suspected as early as 1949, however, that many Fusselman reservoirs may have resulted from intense karstification during post-Fusselman subaerial exposure (Stormont, 1949; Gibson, 1965). Only in the last 15 years has the significance of karstification to reservoir development and occurrence in the Fusselman been fully appreciated (Wright, 1979; Mear and Dufurrena, 1984; Garfield and Longman, 1989; Geesaman and Scott, 1989; L. J. Mazzullo and others, 1989; S. J. Mazzullo, 1989; Mear, 1989; L. J. Mazzullo, 1990a,b; Canter and others, 1992; Mazzullo and Mazzullo, 1992; Troschinetz, 1992a,b). Although the top of the Fusselman is known to be a major sequence-bounding unconformity throughout most of the Permian basin (Hills and Hoenig, 1979; Canter and others, 1992; LeMone, 1992), specific exploration for subunconformity paleokarst reservoirs has not yet been fully applied to potential Fusselman reservoirs here. This situation is likely due, at least in part, to the fact that comprehensive models of Fusselman paleokarst reservoirs in the Permian basin are not as well-known as they are, for example, in Ellenburger (Ordovician) reservoirs (e.g., Kerans, 1988; S. J. Mazzullo, 1990).

This paper describes sedimentologic and diagenetic evidence that confirms regional karstification in the Fusselman, and it provides exploration models for karsted Fusselman reservoirs. The approach of this paper is first to examine karsted Fusselman carbonates in outcrops, and then to describe the sedimentologic, diagenetic, and mappable attributes of subsurface Fusselman strata in the Permian basin that relate to the development of karsted reservoirs.

LOCATION AND STRATIGRAPHY

The Fusselman Formation crops out in several mountain ranges in West Texas and southeastern New Mexico, in areas that immediately surround the subsurface Permian basin. Outcrops in the southern Hueco Mountains in Hudspeth County, Texas (Fig. 1), were specifically chosen for study because the more-or-less complete and continuous section of Fusselman strata here is well-exposed and readily accessible. The section consists of ~600 ft of dolomites and dolomitic limestones that unconformably overlie carbonates of the Montoya Group (Upper Ordovician), and, in turn, are overlain unconformably by Middle Devonian carbonates (Fig. 2) (McGlasson, 1968; Beard, 1983; LeMone, 1992). Fusselman lithofacies here are of shallow-marine origin, deposited on the periodically exposed Diablo platform (McGlasson, 1968). Several representative Fusselman fields and adjoining areas in the north-central Midland basin and on the central basin platform were also examined (Fig. 1). The Fusselman varies in thickness from 60 to 100 ft on the central basin platform, and it is ~300 ft thick in the north-central Midland basin away from its erosionally truncated edge (Fig. 1). It unconformably overlies a thin section of Sylvan Shale or the Montoya Group, and, in turn, is unconformably overlain by Niagaran to Cayugan carbonates of the Wristen Formation (Fig. 2). Fusselman sedimentary facies in these areas also are mostly of shallow-marine origin (Garfield and Longman, 1989;

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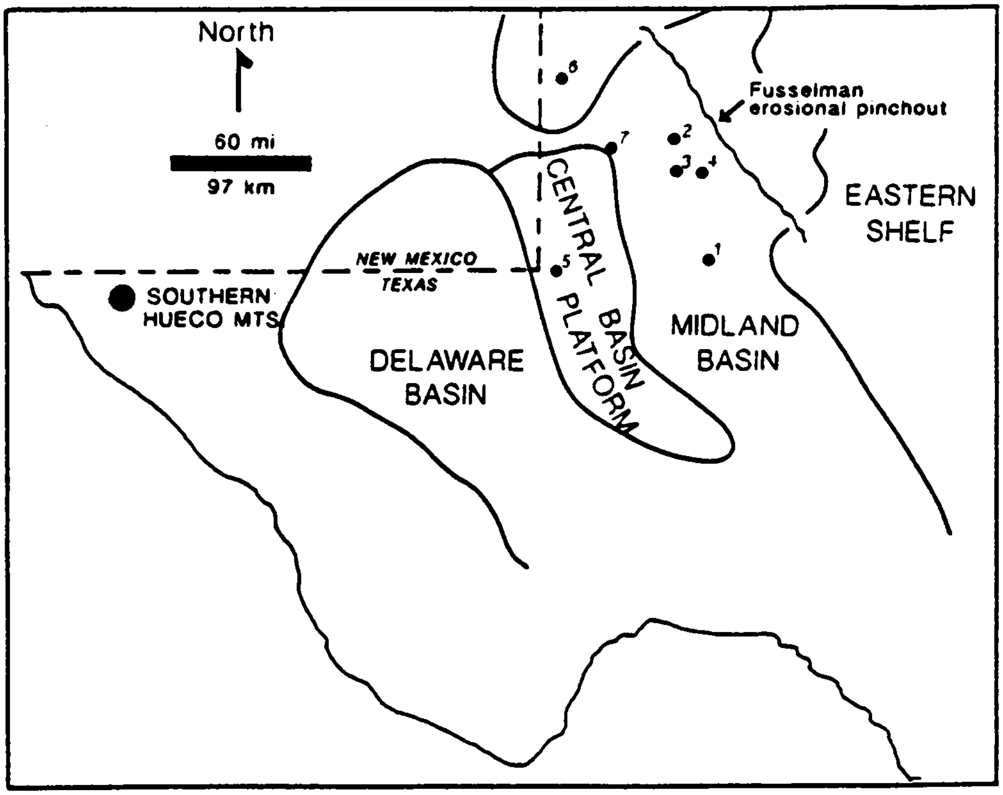

Figure 1. Location of study areas in southern Hueco Mountains and Midland basin. Fields referred to in text are: (1) Mid-Mar, Midland County; (2) Wells, Dawson County; (3) Patricia and Milagro, Dawson County; (4) Tex-Hamon, Dawson County; (5) Flying W, Winkler County; (6) Brahaney Northwest, Yoakum County; (7) Seminole, Gaines County.

Geesaman and Scott, 1989; L. J. Mazzullo and others, 1989; Mear, 1989; Canter and others, 1992).

Both on the outcrop and in the subsurface, the Fusselman is regarded to be of Lower to Middle Silurian (Alexandrian to Niagaran) age (see LeMone, 1992, for review). It is therefore equivalent stratigraphically to the lower part of the Hunton Group of Oklahoma (Fig. 2).

FUSSELMAN PALEOKARST IN SOUTHERN HUECO MOUNTAINS

Features of paleokarst origin previously have been noted in Fusselman outcrops in the Hueco Mountains (King and others, 1945; Lucia, 1971; Beard, 1983; S. J. Mazzullo, 1983) and in the nearby Franklin Mountains (Harbour, 1962; Lucia, 1971; McGlasson, 1971; LeMone, 1987). Both megascopic and microscopic features of paleokarst occur throughout the entire Fusselman section in the Hueco Mountains, and they are described below.

Megascopic Karst Features

The most conspicuous karst features in the Fusselman are paleocaves and dissolution-enlarged fractures. Filled paleocaves vary greatly in size and orientation; the largest paleocave so far examined is 320 ft long by 80 ft high, in the lower part of the section (Fig. 3A). The roof of this paleocave is fractured and disrupted by collapse brecciation (Fig. 3A,B), and the paleocave is filled completely by coarse dolomite-clast and local limestone-clast breccias and associated bedded (layered, graded, cross-stratified) sediments composed of sand- to pebble-size fragments of dolomite host rock (Fig. 3C–E). The breccias and associated sediments are interpreted to have formed during cave collapse. The vertical stratigraphy of this paleocave (Fig. 4) is, except for the absence of the low-porosity/low-permeability shaly and sandy cave-fill facies, essentially similar to that described by Kerans (1988) in Ellenburger paleocaves in the Permian basin. Smaller paleocaves in the Fusselman are either horizontally or vertically disposed and generally have irregular cross-sectional geometries; they also are filled with collapse-breccias overlain by layered sediments.

Vertical and subvertical, dissolution-enlarged fractures (joints and fissures) are intimately associated with paleocaves in the Fusselman. They are a few meters to tens of meters wide and high, and locally they extend laterally for hundreds of meters. They also are filled with collapse breccias and bedded sediments; hence they represent "breccia pipes." Galleries of small, filled paleocaves (so-called "pocket breccias") occur locally along some of these fractures.

Microscopic Karst Features

Microscopic features of precipitational origin are associated with the filled paleocaves and dissolution-enlarged fractures, and these features are typical of karsted carbonate rocks (Esteban and Klappa, 1983; Choquette and James, 1988). The features include: (1) pisolitically coated clasts (Fig. 3B,D); (2) dripstone (pendant) cements beneath grains

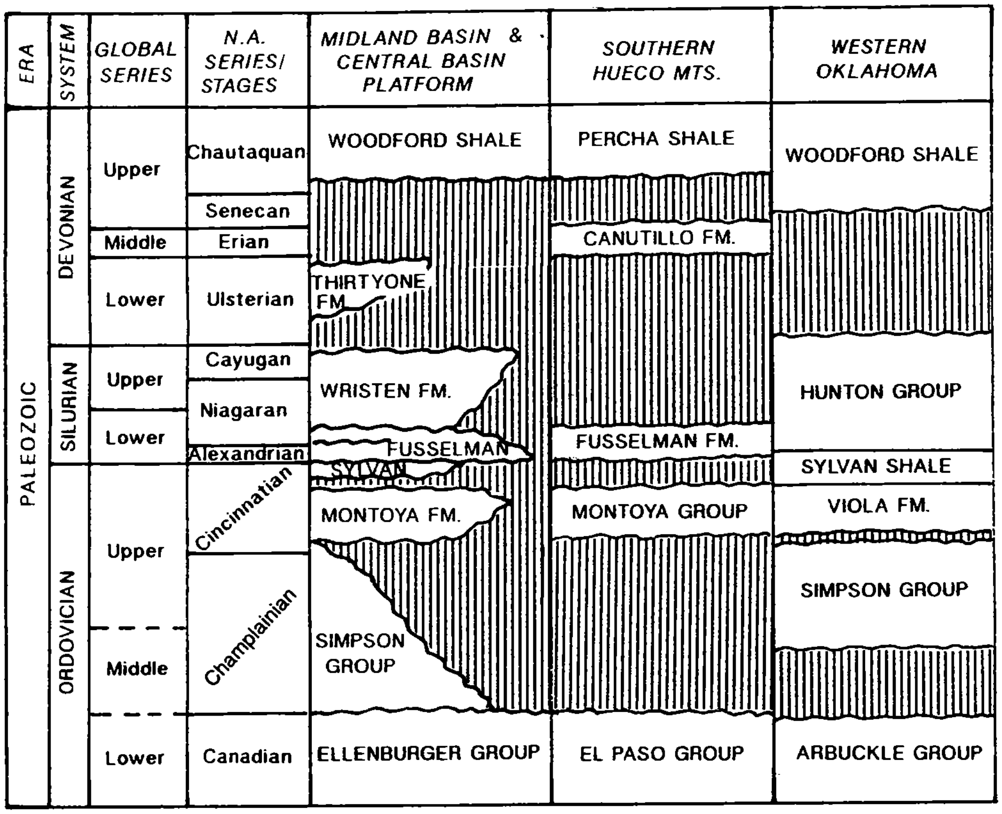

Figure 2. Stratigraphy in study areas and western Oklahoma (modified from Canter and others, 1992).

(Fig. 5A); (3) flowstones (Fig. 3B); and (4) soilstone crusts (Fig. 3C).

Porosity in Outcrop Examples

Three types of porosity occur in cave-filling and fracture-filling deposits: (1) small *vugs* within some dolomite clasts, in the matrices of the collapse breccias, and in bedded sediments (Fig. 5B); many of these vugs are partially occluded by fine-crystalline (25 microns) dolomite cement; (2) *interparticle pores* in the matrices of collapse breccias composed of dolomite-cemented grainstones, and also in dolomite-cemented, bedded sediments (Fig. 5C); and (3) *fabric-selective, particle-moldic pores* in collapse breccias and bedded sediments (Fig. 5B).

SUBSURFACE FUSSELMAN PALEOKARST

Various lines of evidence, which suggest that traps and porosity in many Fusselman reservoirs in the Permian basin are related directly to karstification, are described below.

Evidence of Paleocave-Associated Reservoirs

Stormont (1949) and Gibson (1965) were among the first workers to suggest that cavernous and vuggy porosity at and near the top of productive Fusselman sections (Fig. 4) were related to karstification and the formation of (at least partially unfilled) paleocaves of various sizes. Such extensive porosity typically is manifested by rapid bit drops during drilling and lost mud circulation. Their ideas later were confirmed by other workers (Mear and Dufurrena, 1984; Garfield and Longman, 1989; Geesaman and Scott, 1989; L. J. Mazzullo and others, 1989,1990a,b; Mear, 1989; Canter and others, 1992; Mazzullo and Mazzullo, 1992), who also suggested that unconformities exist at the top of, and within, the Fusselman in the Midland basin and on the central basin platform (Fig. 2). Hence, multiple paleocave systems related to these separate unconformities apparently are present locally in the subsurface (Fig. 4). Filled paleocaves in the subsurface Fusselman that are analogous to those in the southern Hueco Mountains are evidenced by the occurrence of porous, dolomite-clast, collapse breccias associated with dissolution enlarged fractures at the top of, and within, the Fusselman in producing fields and surrounding regions in the study area (Fig. 1). There are examples in the Mid-Mar field (Garfield and Longman, 1989), in several fields in the north-central Midland basin (L. J. Mazzullo and others, 1989), and in the Flying-W field (Mear, 1989).

These studies, and those of Mear and Dufurrena (1984), S. J. Mazzullo (1989), L. J. Mazzullo (1990a,b), and Mazzullo and Mazzullo (1992), suggested that cavernous, vuggy, interparticle and intercrystalline porosity in many Fusselman fields, associated with collapse breccias and dissolution-enlarged fractures, are all related to karstification. Hence, the development of reservoir porosity in many Fusselman fields largely is the result of karstification, notwithstanding the presence of fractures (Mear, 1989) or field occurrence in structural traps.

Fusselman Fields in Paleogeomorphic (Buried-Hill) Traps

Karsted unconformity horizons in the subsurface are paleogeomorphic landscapes that characteristically are associated with surficial dissolution features (sinkholes, dolines, dissolution valleys) and residual hills (a.k.a. "buried hills" or "karst towers") that locally overlie caves and caverns. Such surficial features commonly are recognized in and around many Fusselman fields in the Permian basin (Mazzullo and Mazzullo, 1992). In addition to reservoirs in paleocaves, Fusselman production from buried-hill traps also is known in the Permian basin.

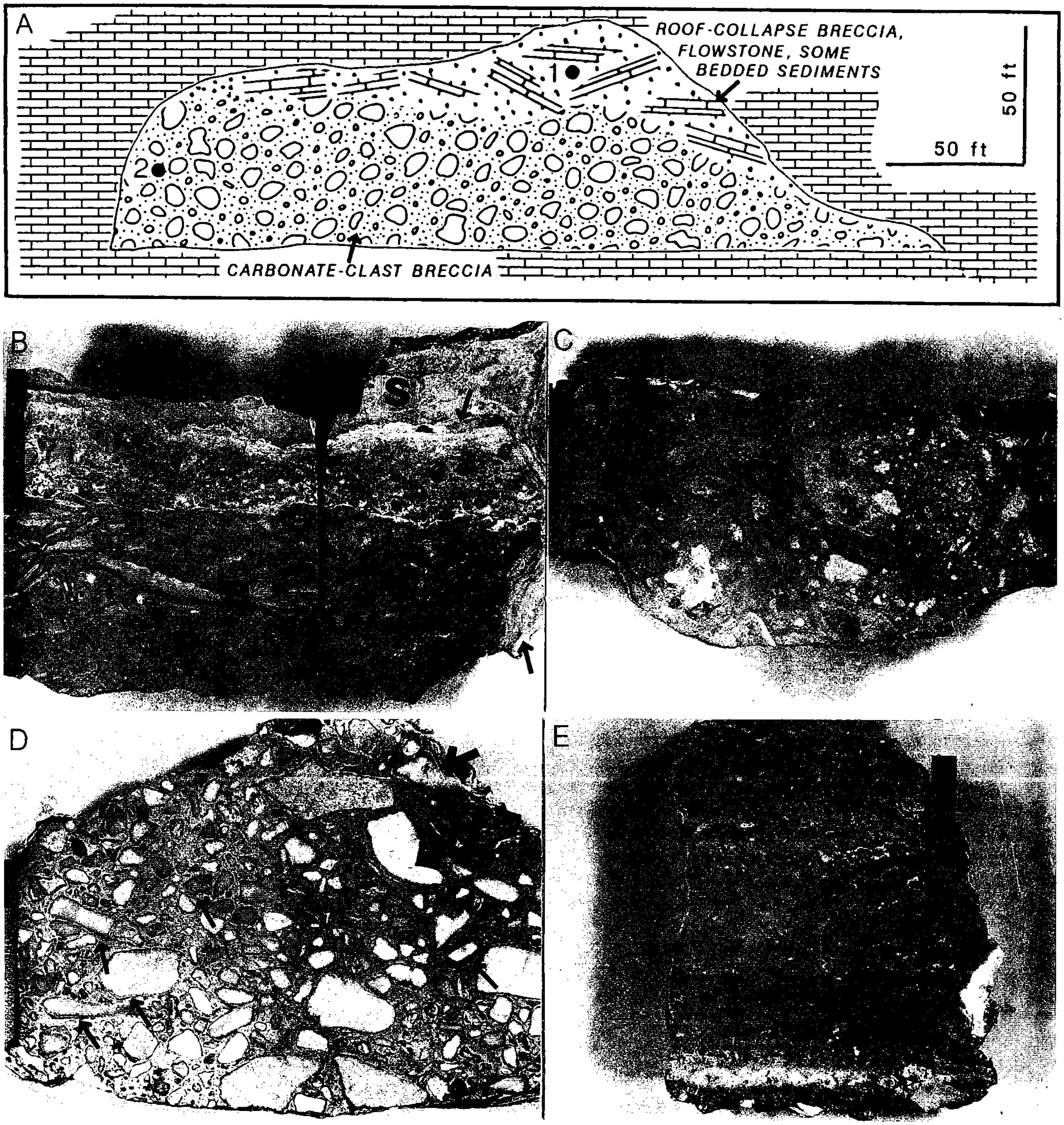

Figure 3. *A*—Sketch of large Fusselman paleocave in southern Hueco Mountains. Dots labeled 1 and 2 refer to Figure 3B,D. *B*—Roof-collapse breccia ("B") overlying pisolitically-coated clasts ("P"), overlain by sediment ("S"); note flowstone encrustations (arrows). Scale on left 5.0 cm. *C*—Soilstone crust (arrows) in surficial collapse breccia; scale is 5.0 cm. *D*—Carbonate-clast breccia with pendant and isopachous pisolitic coats on clasts (small arrows). Note vuggy porosity in dolomite clasts (large arrows); scale is 5.0 cm. *E*—Graded sediments with vuggy and grain-moldic porosity; scale is 5.0 cm.

L. J. Mazzullo (1990a,b) and Mazzullo and Mazzullo (1992) have described reservoirs in buried-hill traps from several fields in the north-central Midland basin, specifically in Patricia, Wells, Brahaney Northwest, Milagro, and Seminole fields (Fig. 1). For example, erosionally thin Wristen carbonate sections overlying the Fusselman in Patricia and Milagro fields (Fig. 6A) correspond to paleotopographic highs held up by the Fusselman (Fig. 6B), which are interpreted as buried-hill traps (Fig. 6C). Porosity in these fields is due to the presence of vugs, interparticle and intercrystalline pores and fractures, and it often is associated with terra rosa soils and soil pisolites.

CONCLUSIONS

Evidence of karstification of the Fusselman is found in outcrops throughout West Texas and southeastern New Mexico; it includes paleocaves, dissolution-enlarged fractures, and features of vadose precipitational origin (e.g., flowstones, dripstones, soilstone crusts, pisolites). Similarly, many Fusselman reservoirs in the Permian basin likewise originated during periods of subaerial exposure and karstification during, or immediately after, Fusselman deposition. Reservoirs occur in unfilled and breccia-filled paleocave systems and in buried hills in structural and combination structural/stratigraphic traps. The documented occurrence of such subtle traps breathes new life into the exploration for additional Fusselman reservoirs in this mature hydrocarbon province.

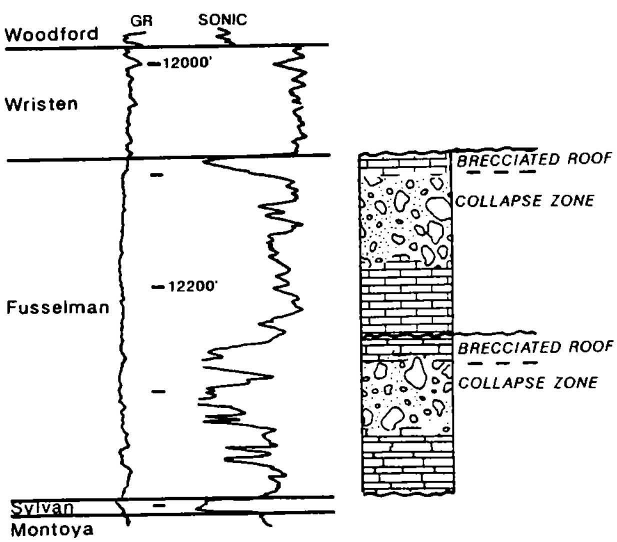

Figure 4. Type log from Wells Fusselman field (Dawson County, Texas) showing zones of cavernous and vuggy porosity and inferred stacked paleocave zones.

Figure 5 (right). *A*—Pendant calcite cement beneath dolomite clast ("D") in collapse-breccia; arrows point to dolomitized laminae in this cement. Scale is 2.5 mm. *B*—Porous, dolomitic grainstone matrix in collapse breccia, with pisolites ("P"), rock fragments (small arrows), vugs partially occluded by dolomite cement (fingers), and vugs within pisolitic coats around grains (large arrow). Scale is 1.0 mm. *C*—Dolomitized grains (arrows) in grainstone matrix of collapse-breccia. Clear dolomite cement has occluded most interparticle porosity although remnant interparticle and vuggy pores are present; cross-polars. Scale is 0.25 mm.

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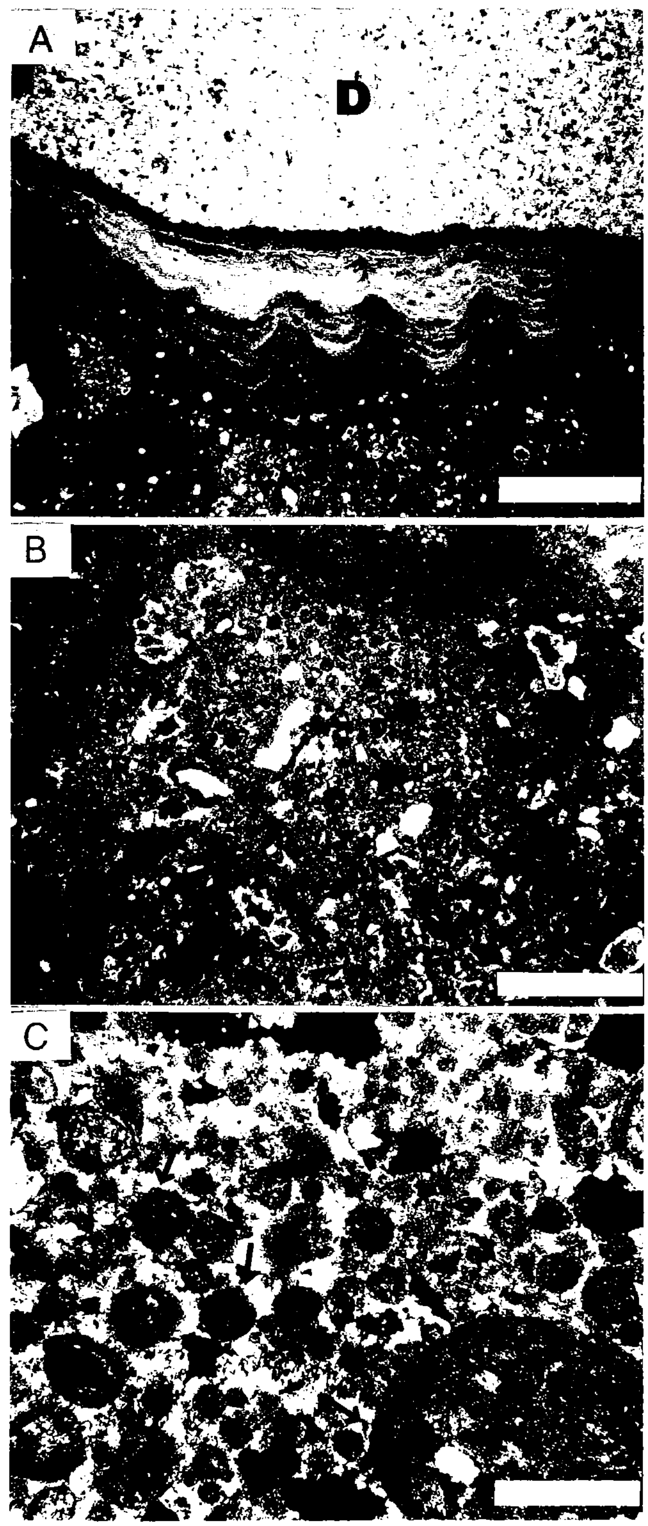

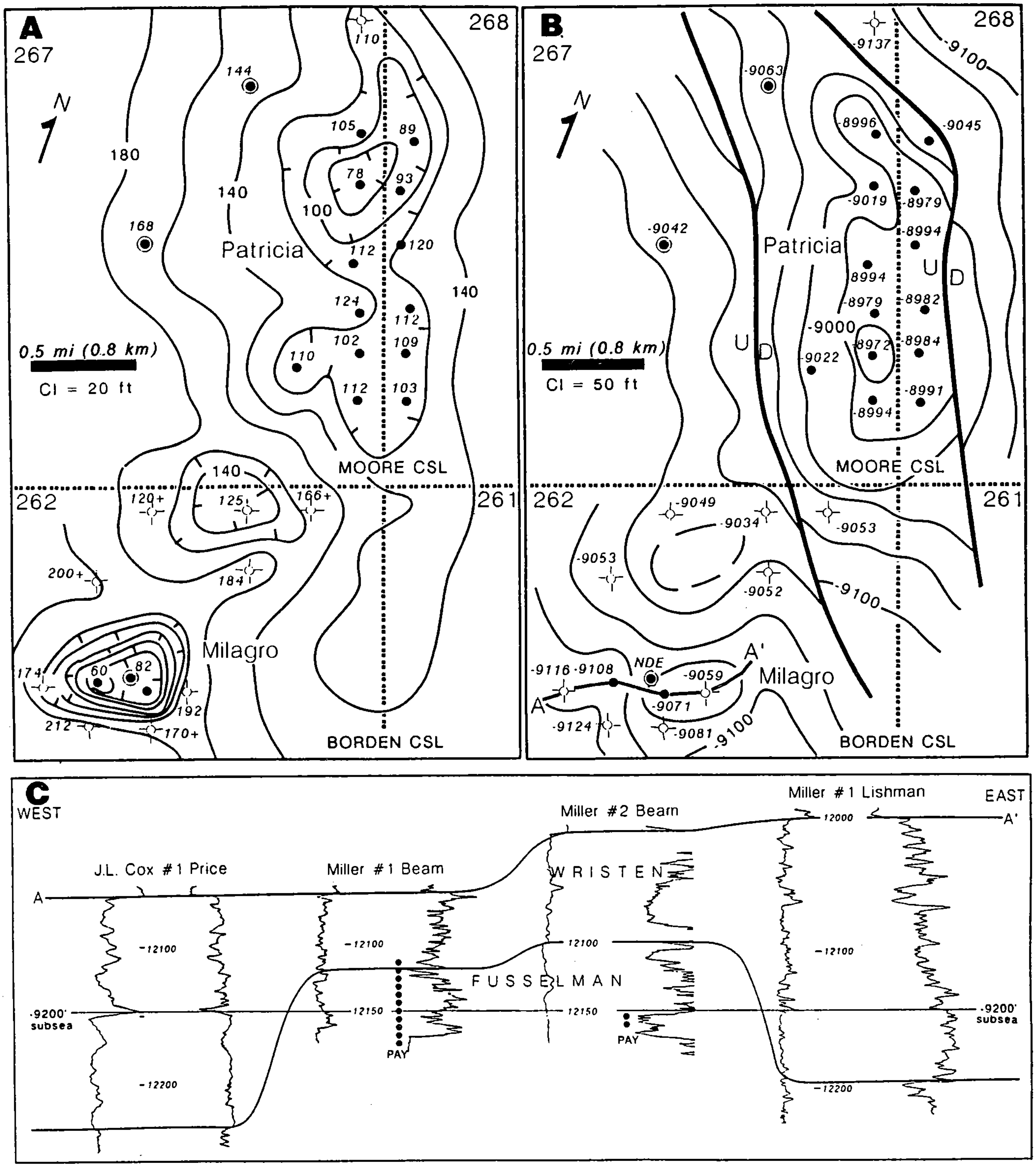

Figure 6. Patricia and Milagro fields, Dawson County, showing examples of buried-hill traps. *A*—Isopach map showing thinning of Wristen Formation over an inferred paleotopographic high on the Fusselman, as indicated by a structure map drawn on the base of the Woodford (*B*). *C*—Cross section through Milagro field illustrating buried-hill trap. Black dots on maps and section represent Fusselman production; production from shallower horizons is indicated by circled dots. Reprinted with permission of Permian Basin Section, SEPM (Society for Sedimentary Geology).

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General Aspects of the Hunton Group in the Western End of the Anadarko Basin

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ABSTRACT.—Recent wildcat drilling in the western Anadarko basin by Arrington CJM, Inc. has added significant new information on the Hunton Group in Hutchinson and Roberts Counties of the Texas Panhandle (Fig. 1). This drilling resulted in six dry holes and two field discoveries, the Arrington Hunton field and the Arrington West Hunton field. These two oil fields are the western-most Hunton oil production known to date in the Anadarko basin. The Hunton Group in this area is very similar to the Hunton in the rest of the Anadarko basin. Three porosity zones are present in the Hunton in most wells. The lower two porosity zones are stratigraphically controlled. The upper porosity zone is at the unconformity on top of the Hunton.

The oil in the Hunton Group in the two newly discovered fields was derived from Ordovician source rocks. The time of oil generation was between Hunton deposition and the Pennsylvanian Ouachita orogeny. The top seal of the Hunton reservoirs is Kinderhook shale, not the Woodford Shale.

GEOLOGIC SETTING

The western end of the Anadarko basin is bounded on the south by the buried Amarillo uplift. The Amarillo uplift was raised so high during Pennsylvanian time that the entire sequence of pre-Permian Paleozoic strata was eroded away; Precambrian granite subcrops below the Permian Brown Dolomite. To the north and west, the western end of the basin gradually merges with the northern and western shelf. To the east, it plunges into the main Anadarko basin.

The Hunton Group (Late Ordovician–Devonian) subcrops along the northern flank of the Amarillo uplift. Several large fault blocks are eroded down to the Ellenburger Group, or to granite, but they have a fringe of subcropping Hunton. The overlying rocks are granite wash (conglomerates and arkosic rocks) and thus there is no seal. The western Hunton subcrop extends under the old Panhandle field (acreage held by shallow production). The northern subcrop of the Hunton feathers out on the northern shelf. The potential for production from the Misener is small, because the Simpson sand is thin in the area and the Simpson subcrop is significantly farther to the north. Reworked Simpson sand, derived from the south, is a significant component of the Morrow cherts.

Two major unconformities are present in the western part of the Anadarko basin; the upper one is at the base of the upper Morrow, and the lower one is at the top of the Hunton. On the south flank of the western Anadarko basin, the upper Morrow unconformity truncates the Hunton unconformity. The two unconformities merge far up on the northern shelf.

HUNTON GROUP CHARACTERISTICS

The main aim of this report is to present data on characteristics of the Hunton Group in the western part of the Anadarko basin. A series of geophysical, drilling-time, and lithologic logs are presented along a line of cross section that is shown in Figure 1. The main focus is on the Arrington West Hunton field, where a number of wells have been drilled on the West Turkey Track lease.

In the Patterson No. 1-10 well, the Hunton is 156 ft thick (Fig. 2). The rocks consist of white to off-white limestone that is dolomitic in part. The upper porosity zone has been removed by erosion and the lower two porosity zones have merged. The formation has an oil show in the top but the lower porosity zone appears to be water wet.

A series of wells (Figs. 3–6) can be used to show a typical cross section across the Arrington West Hunton field. The Hunton averages ~198 ft thick. The rocks are primarily white to tan, limy dolomite. The northwest sector of the field contains more off-white limestone and dolomitic limestone than the rest of the field. All three porosity zones are present across the field, although in the northeast quadrant of the field the upper porosity zone has been re-

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moved by erosion. Across the field, the porosity ranges ≤20%. All three porosity zones are within the oil-production zone across the crest of the field. Figure 7 is a structure map of the field.

The discovery sequence was as follows: The Hunton discovery well was on the west flank of the field. It had an initial potential of 50 bbl oil and 350 bbl water. The second Hunton well encountered the Hunton 13 ft lower than the discovery well and had only 3 ft of pay; it was judged noncommercial. The third Hunton well encountered the Hun-

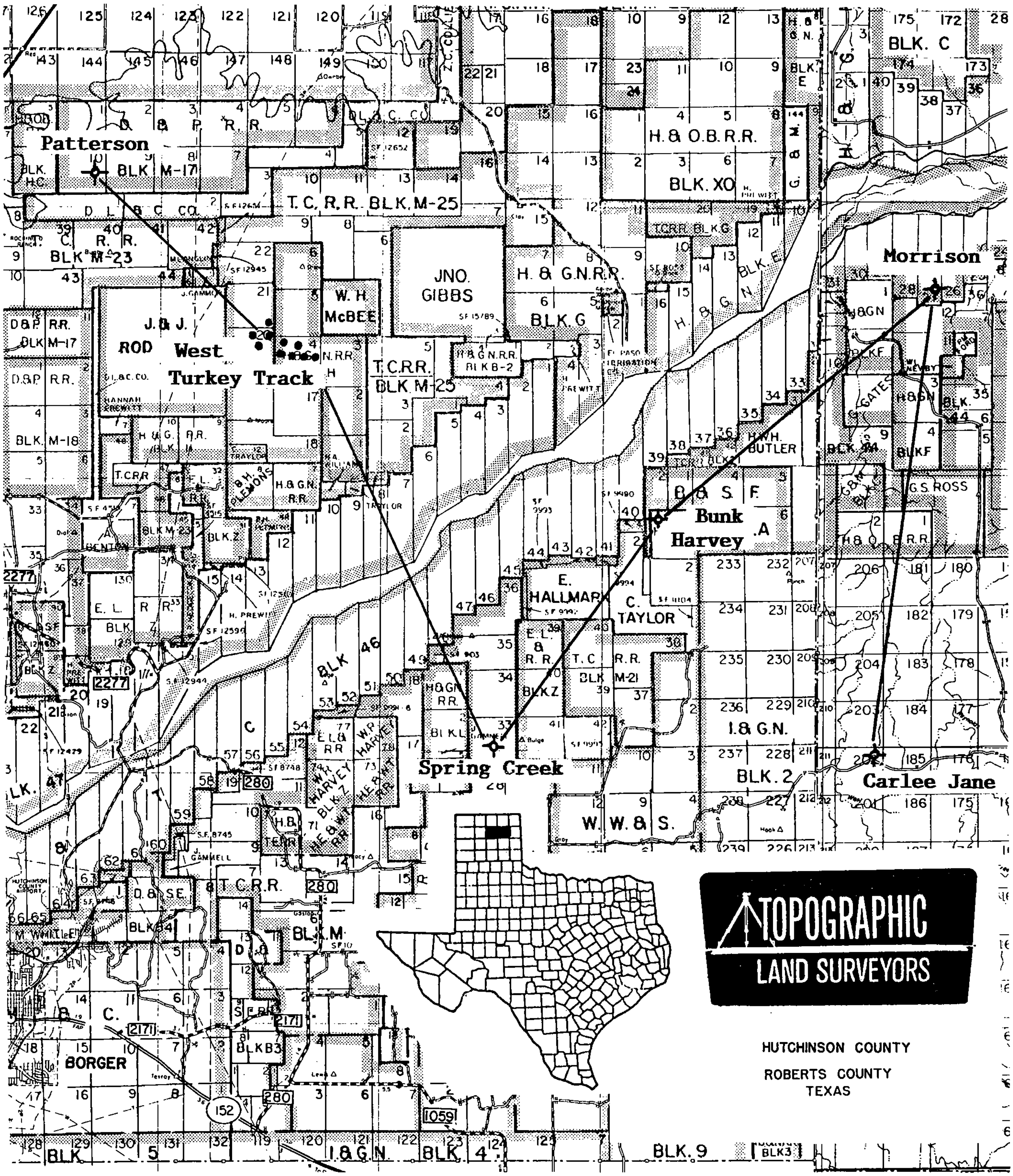

Figure 1. Location of study area in eastern Hutchinson and western Roberts Counties, Texas.

ton 99 ft higher than the discovery well. The second Hunton porosity zone was tested in this well and found oil productive. The fourth Hunton well encountered the Hunton 73 ft higher than the discovery well. The third Hunton porosity zone was tested and found oil productive (Fig. 4). The pay zone has a low resistivity on geophysical logs.

The Hunton oil here was fingerprinted by Chevron Oil Co. as part of their exploration program, and the data were shared with Arrington CJM. The oil was derived from Ordovician source rocks. A thin sequence of Mississippi lime over the structure indicates that the structure is older than the Mississippi lime.

Chevron also fingerprinted a heavy oil from the lower granite wash, a rock unit that probably is post-Cherokee in age. This oil is degraded, possibly an Ordovician-type oil. I personally believe this is a captured oil seep from the Hunton reservoir whose seal was slightly damaged during the Pennsylvanian Ouachita orogeny. A supporting indication of this is a completion test at the base of the third porosity zone in the West Turkey Track No. 9 well that still had a 1–5% oil cut at 30 ft below the oil/water contact of the field. This suggests the Hunton reservoir has been partially drained due to natural causes. However, part of the lost oil may be trapped in fractured Mississippi lime, and part of the lost oil may be trapped in the captured oil seep in the lower granite wash. Another indication of partial drainage is the almost total lack of casing-head gas associated with the Hunton oil production.

In the Spring Creek No.1 well, the Hunton is 188 ft thick (Fig. 8). The rock consists of bone-white chert at the top, grading down first into white cherty limestone and then into limestone with varicolored chert and some pink limestone. Shales above the Hunton are varicolored red, maroon, purple, yellow, green, and gray. The shales below the Hunton are varicolored light blue-green, khaki, olive, gray-green, light gray, and red. All three porosity zones are present, but they are of low quality and appear water wet. The above-described colors indicate that the Hunton here has been exposed to weathering and oxidation.

In the Bunk Harvey No. 1 well, the Hunton is 202 ft

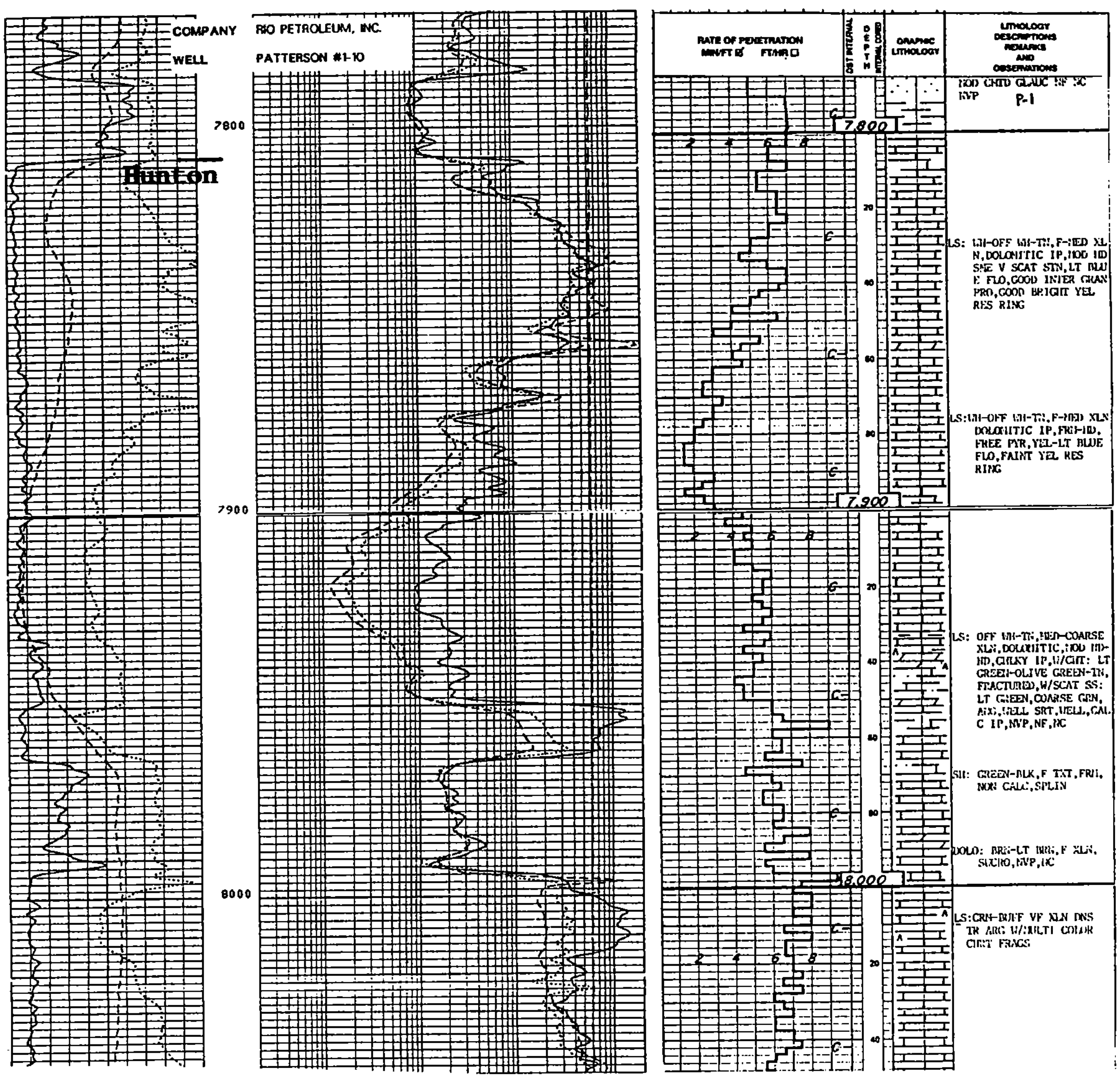

Figure 2. Logs of Rio Petroleum Patterson No. 1-10 well, sec. 10, Block M-17, D&PRR Co. Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 3,112 ft.

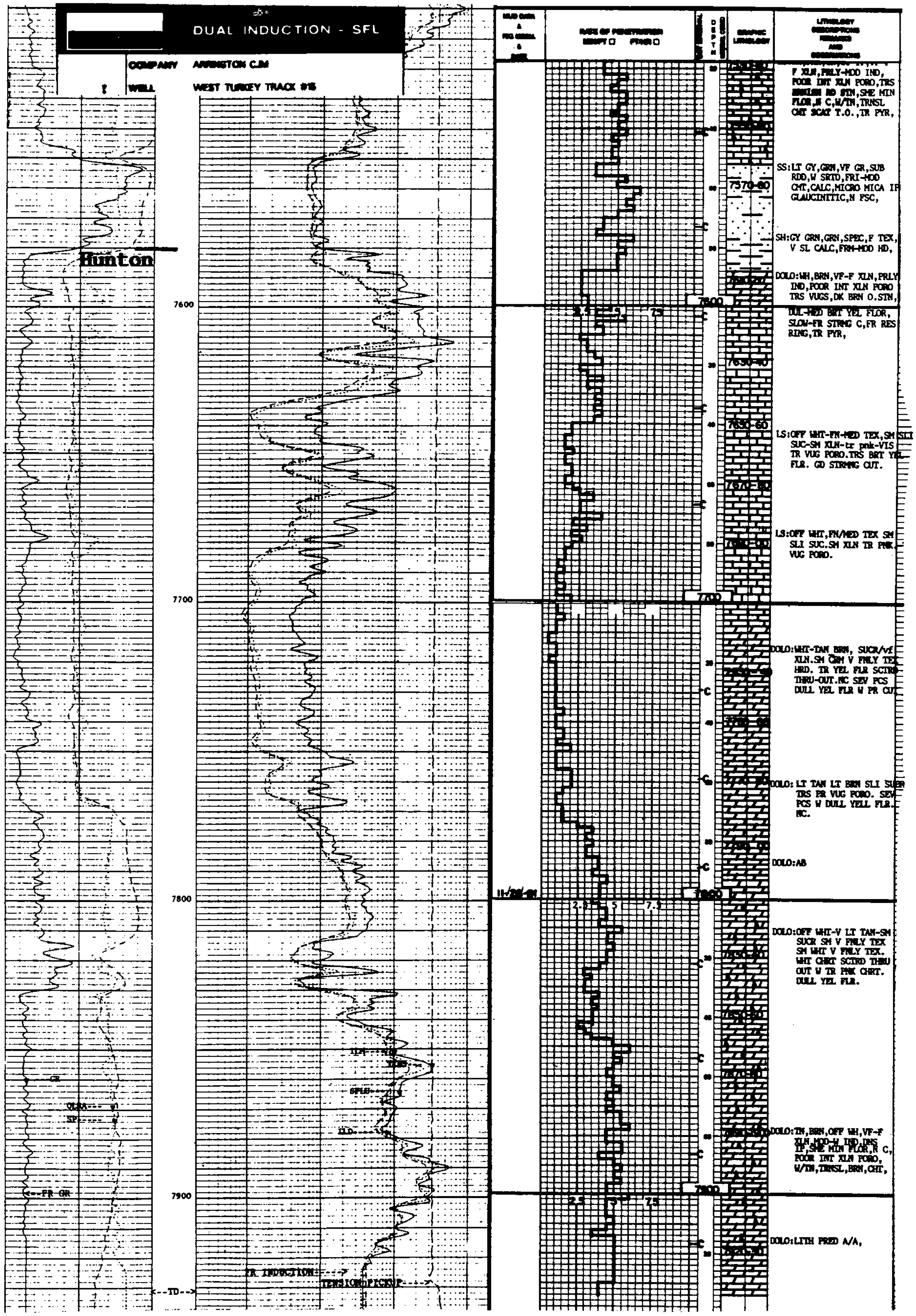

Figure 3. Logs of Arrington CJM West Turkey Track No. 15 well, sec. 15, Block M-25, TCRR Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,933 ft.

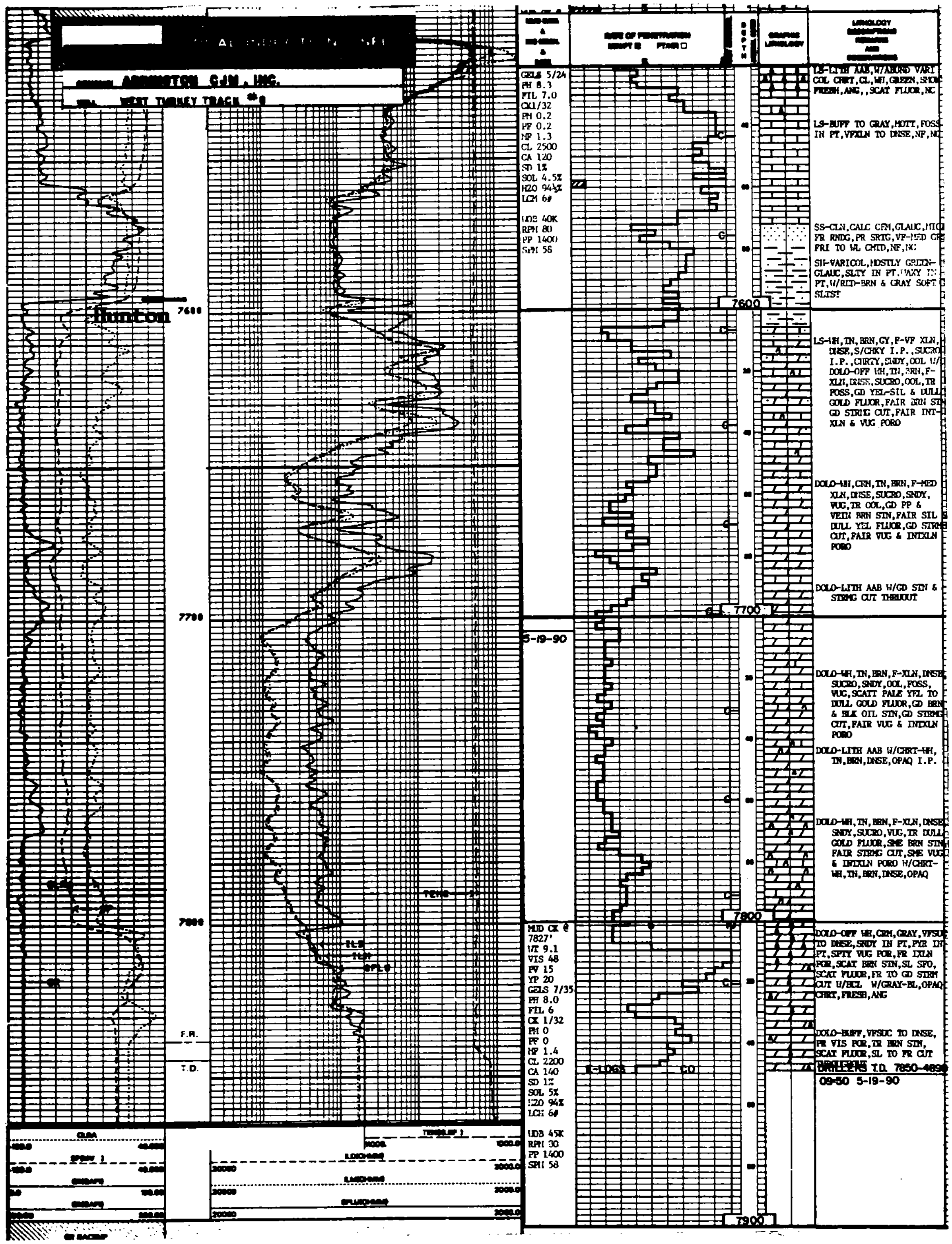

Figure 4. Logs of Arrington CJM West Turkey Track No. 8 well, sec. 20, Block M-25, TCRR Survey, Hutchinson County, Texas. Dual induction SFL, drilling time, and lithologic logs. Datum is kelly bushing, elevation 2,952 ft.

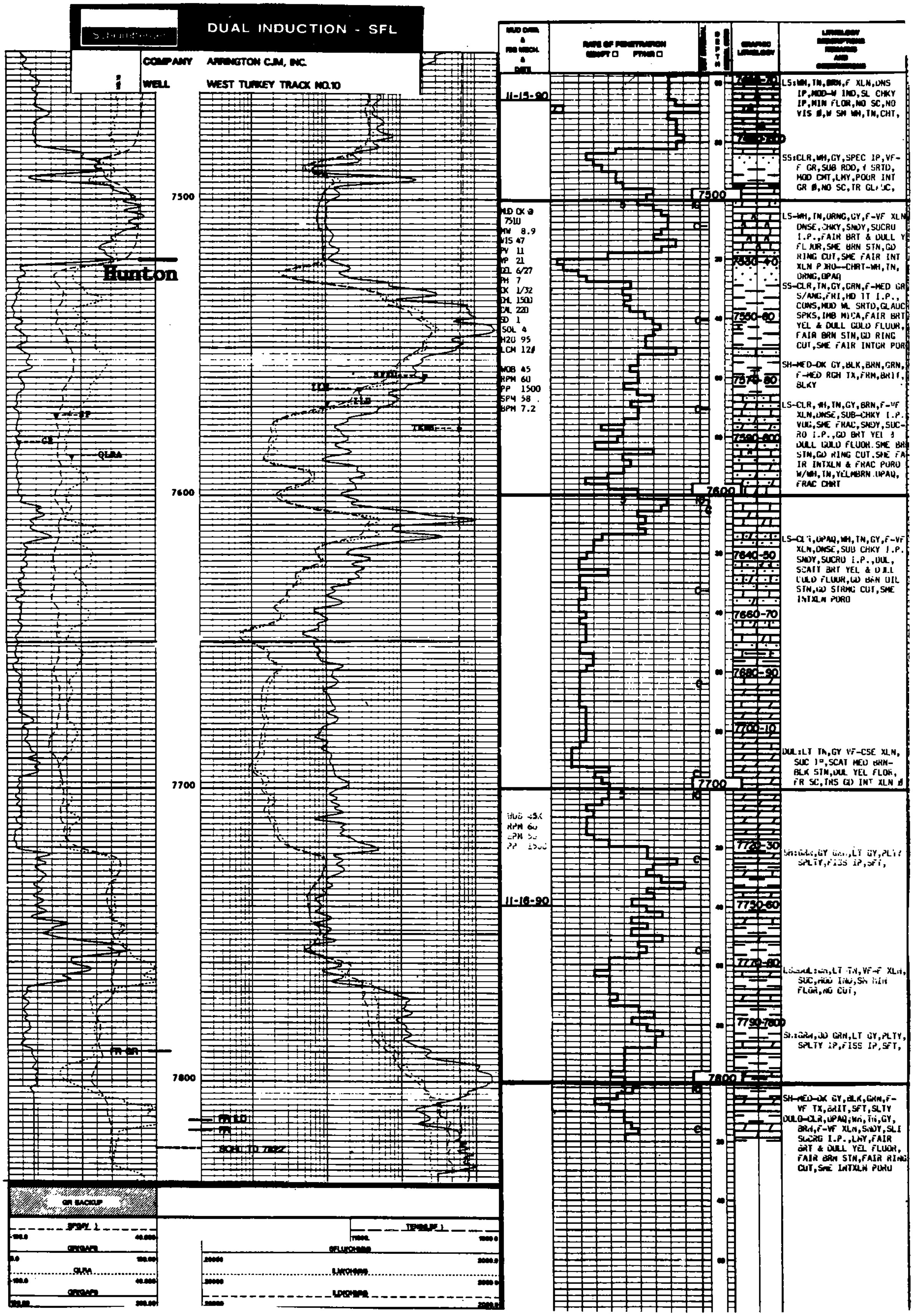

Figure 5. Logs of Arrington CJM West Turkey Track No. 10 well, sec. 4, Block H, H&GN Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,892 ft.

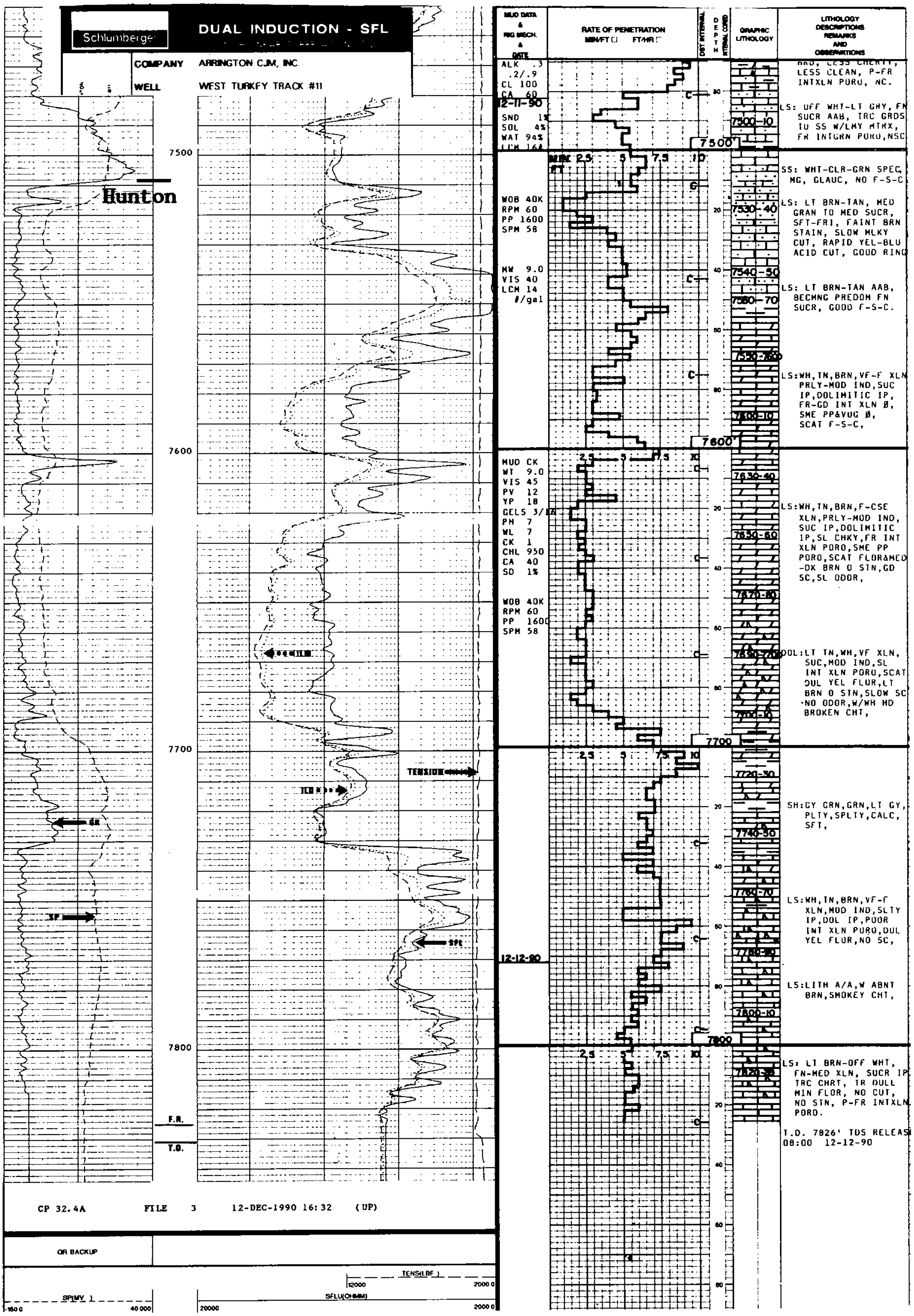

Figure 6. Logs of Arrington CJM West Turkey Track No. 11 well, sec. 4, Block H, H&GN Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,894 ft.

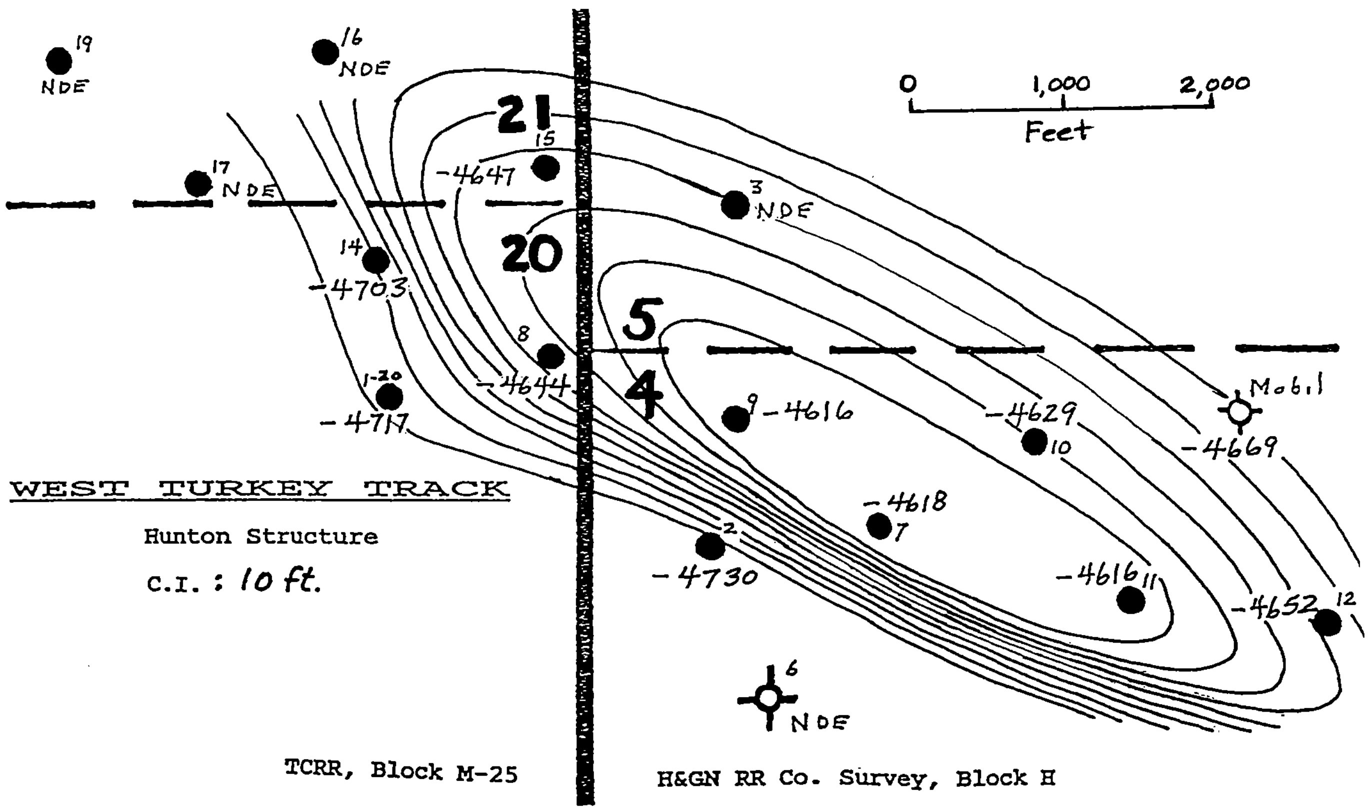

Figure 7. Structure map on top of Hunton Group in the West Turkey Track of the Arrington West Hunton field, Hutchinson County, Texas.

thick (Fig. 9). The rock consists of cream to tan limestone with small amounts of dolomite and chert. All three porosity zones are present. However, the oil show is slight and the porosity zones appear to be water wet; this was confirmed by a production test.

In the Morrison No. 1 well, the Hunton is 280 ft thick (Fig. 10). This well is located farthest east and toward the axis of the basin, therefore the Hunton is deeper and thicker. The rock consists of light-buff, tan, and light-brown limestone with scattered chert. All of the porosity zones are present in this well, and they are thicker than in the wells farther west. The oil show is slight and was not considered worth a production test. The electric logs indicate the formation is water wet.

In the Carlee Jane No. 202-1 well, the Hunton is 268 ft thick (Fig. 11). The rock consists of cream to buff dolomite with some dark- and light-gray chert. The upper porosity zone is quite thick and the lower two porosity zones have merged into one. Porosity is relatively low, without any drilling break. A drill-stem test was conducted over the Hunton, but only drilling mud was recovered in the test. The sample chamber contained drilling mud with a slight show of oil. The electric logs indicate the formation is water wet (Fig. 11).

The Carlee Jane and the Bunk Harvey wells contain a zone between the Hunton and the Viola Groups that is not typical Sylvan Shale (the Sylvan Shale normally separates the Hunton from the Viola). This anomalous Sylvan dolomite, or Maquoketa, or lower Hunton presents a stratigraphic problem similar to the basal-Hunton problem south of Oklahoma City.

A Schlumberger formation micro-scanner log was run in the Carlee Jane well (Fig. 12). This micro-scanner shows some of the detailed stratigraphic, sedimentologic, and structural features within the formation. The upper contact of the Hunton is very sharp. Other observable features include small-scale laminated bedding, bioturbation, vugs, fossil casts, cross bedding, and vertical fractures. Some of the vugs and fossil casts show dog-tooth calcite lining. The lower contact of the Hunton is an interbedded transition into the Sylvan shale.

CONCLUSION

The Hunton Group in the western end of the Anadarko basin is similar to the Hunton throughout the Anadarko basin. The lower contact is a depositional transition from Sylvan Shale to Hunton carbonates. The three main porosity zones extend throughout the area, but the porosity itself is variable. The unconformity at the top of the Hunton truncates the Hunton in other subsurface areas to the north, west, and south.

In many parts of the Anadarko basin, the Woodford Shale provides both the source and the seal for Hunton oil production. An accepted oil-industry myth during the early exploration activity of Arrington CJM was: "No Woodford, no Hunton oil production."

Established production of oil from Hunton reservoirs, where the oil is derived from Ordovician source rocks and the reservoirs are sealed by overlying Kinderhook shale, eliminates the need for the existence of Woodford Shale for Hunton oil production in the western Anadarko basin.

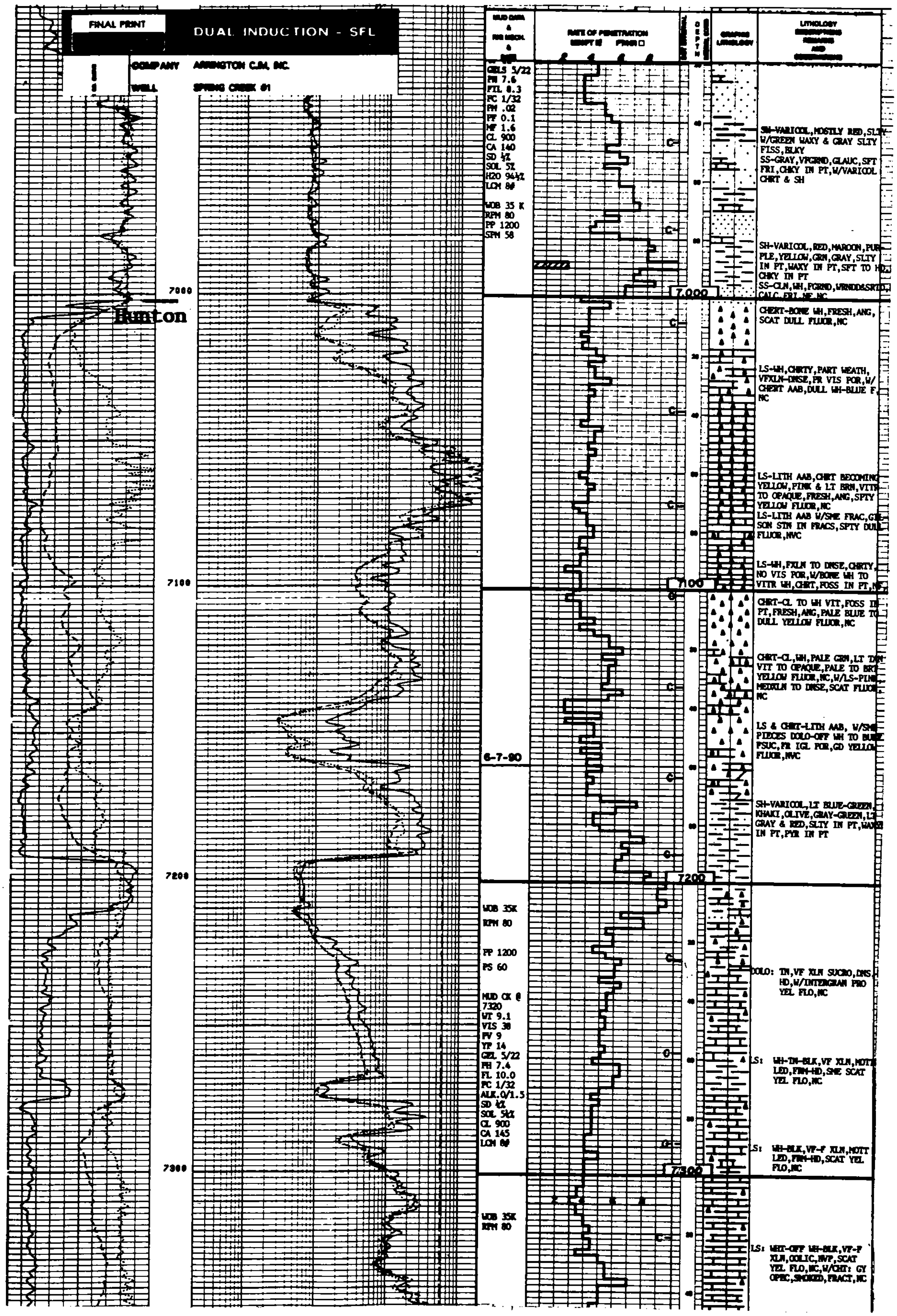

Figure 8. Logs of Arrington CJM Spring Creek No. 1 well, sec. 33, Block M-21, TCRR Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,982 ft.

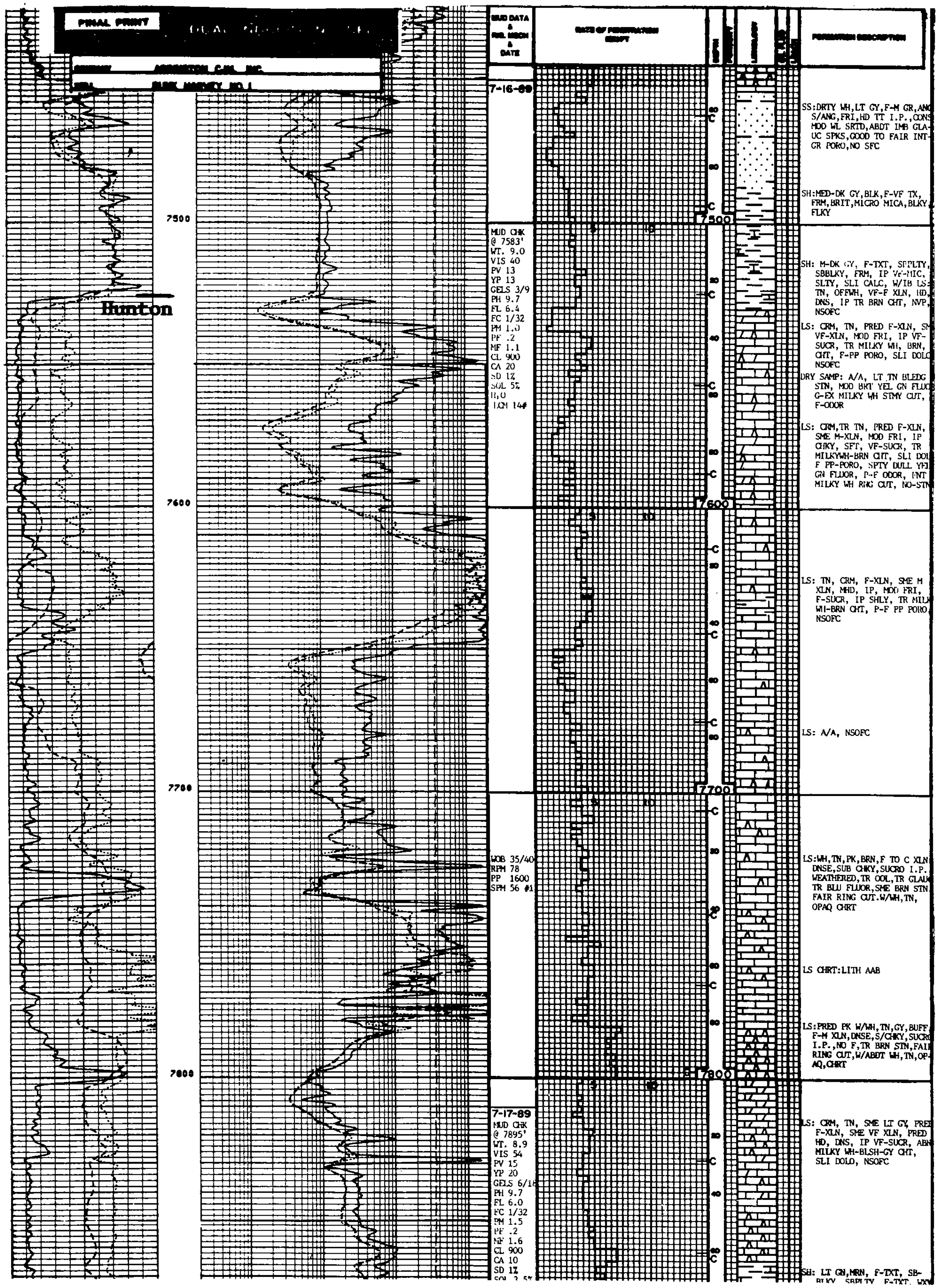

Figure 9. Logs of Arrington CJM Bunk Harvey No. 1 well, sec. 1, Bock LA, H&OBRR Survey, Hutchinson County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,759 ft.

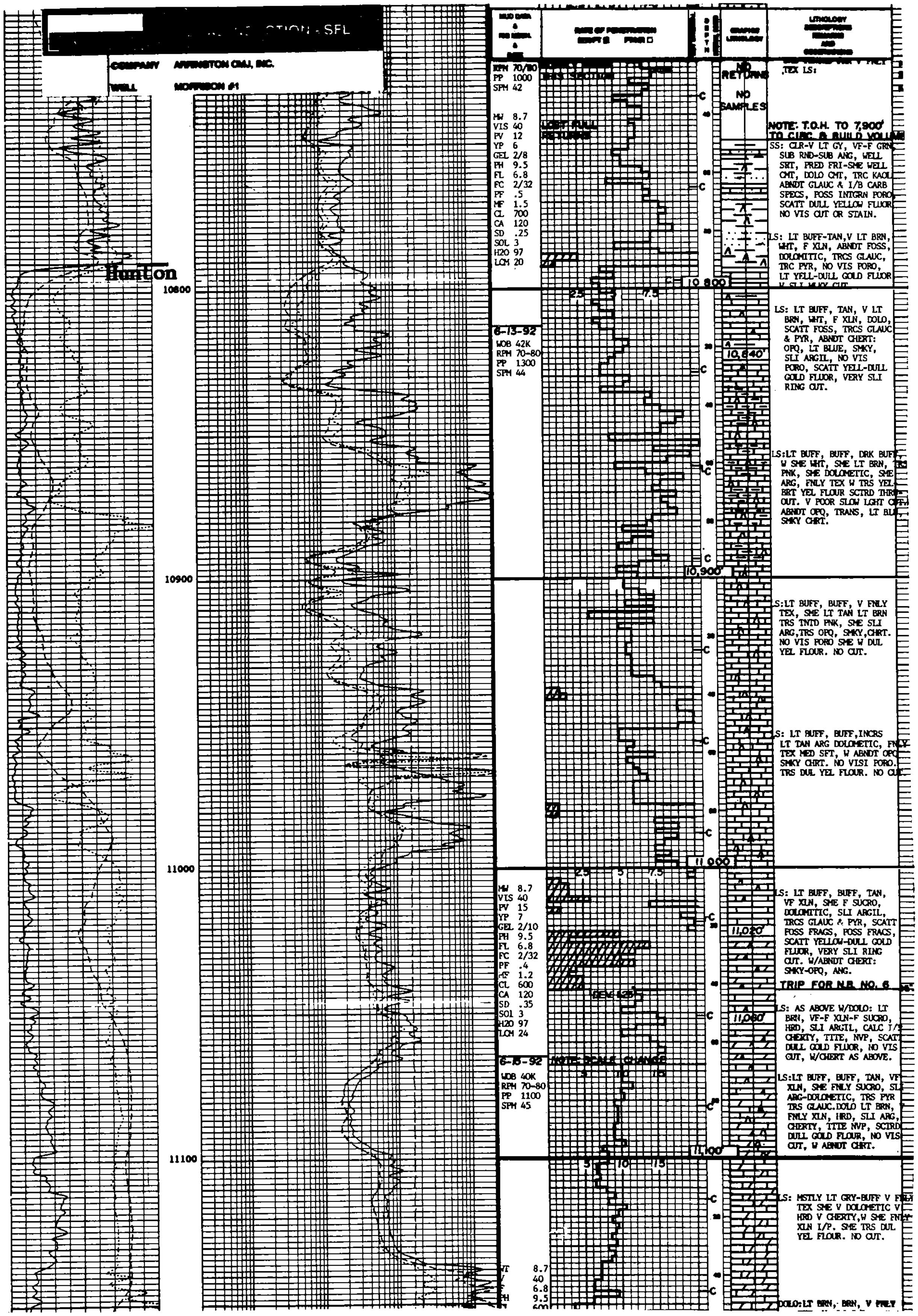

Figure 10. Logs of Arrington CJM Morrison No. 1 well, sec. 27, Block 46, H&TC Survey, Roberts County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 2,795 ft.

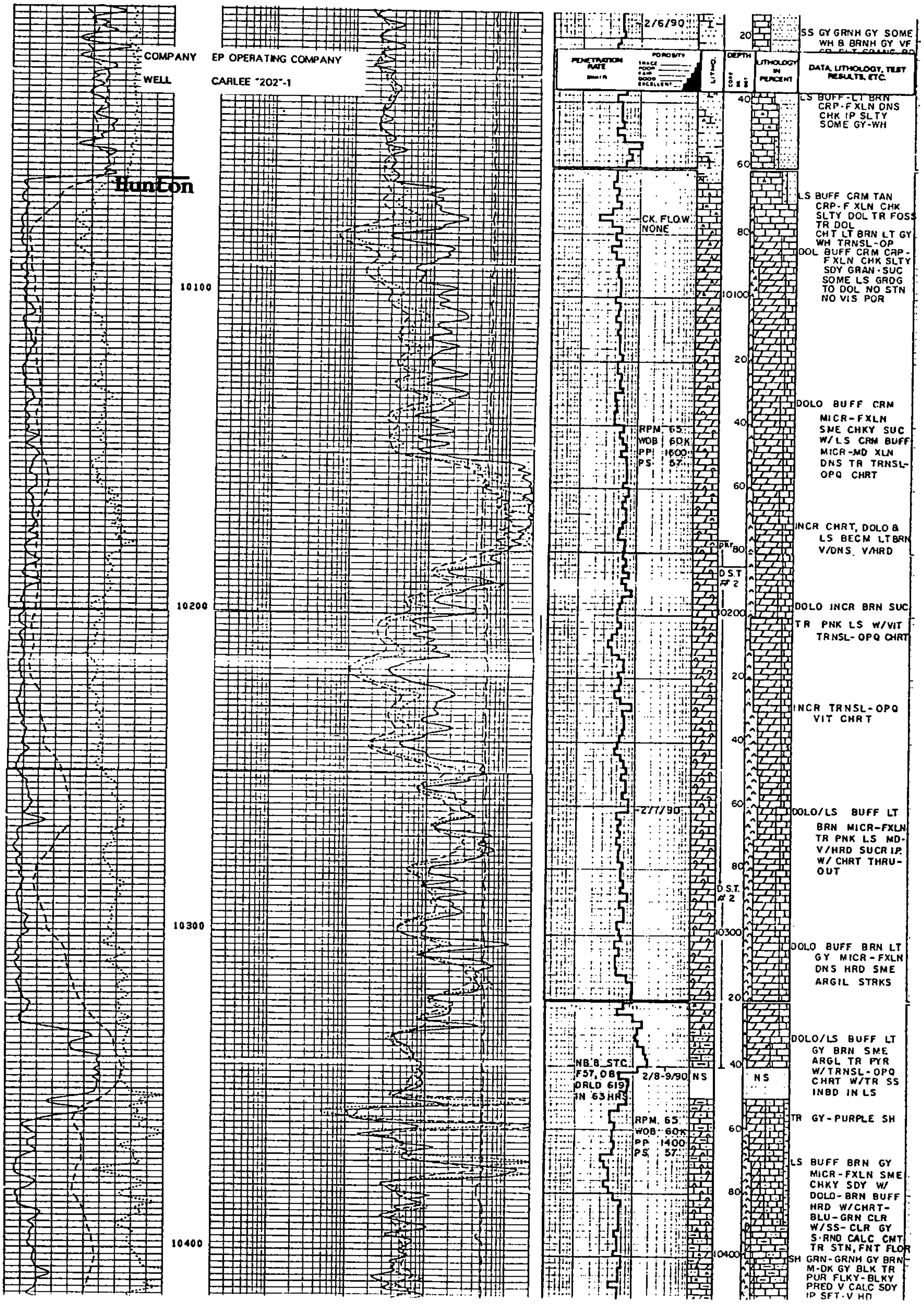

Figure 11. Logs of Arrington CJM (originally EP Operating Co.) Carlee Jane No. 202-1, sec. 202, Block 2, I&GN Survey, Roberts County, Texas. Dual induction SFL, drilling-time, and lithologic logs. Datum is kelly bushing, elevation 3,041 ft.

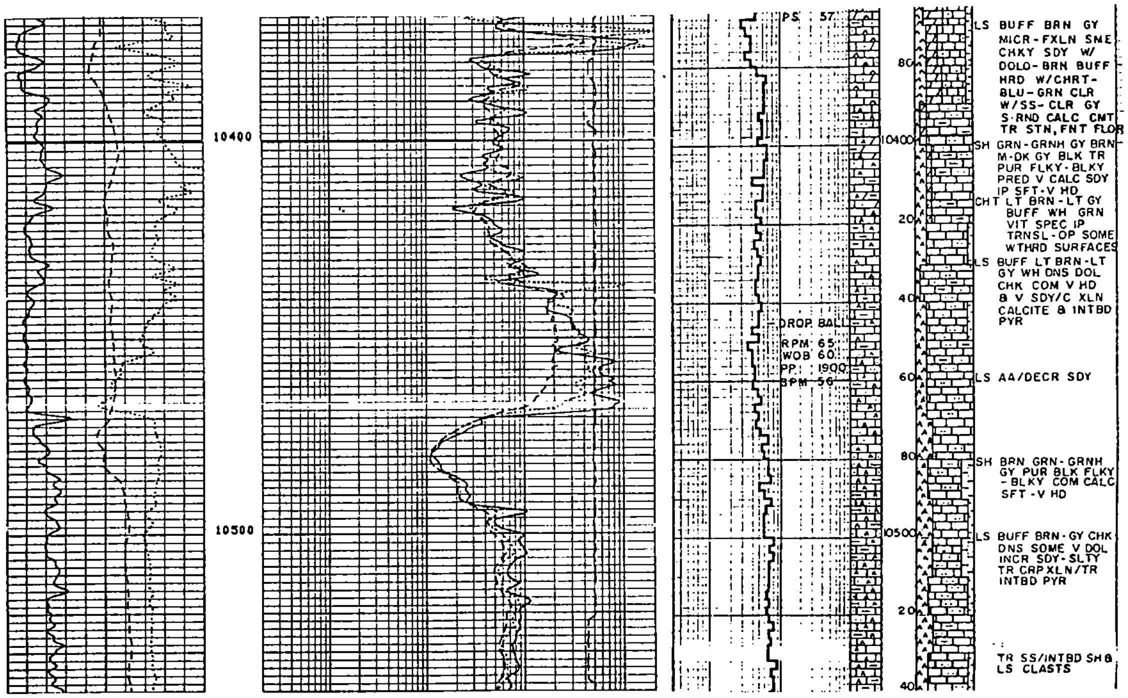

Figure 11. *Continued.*

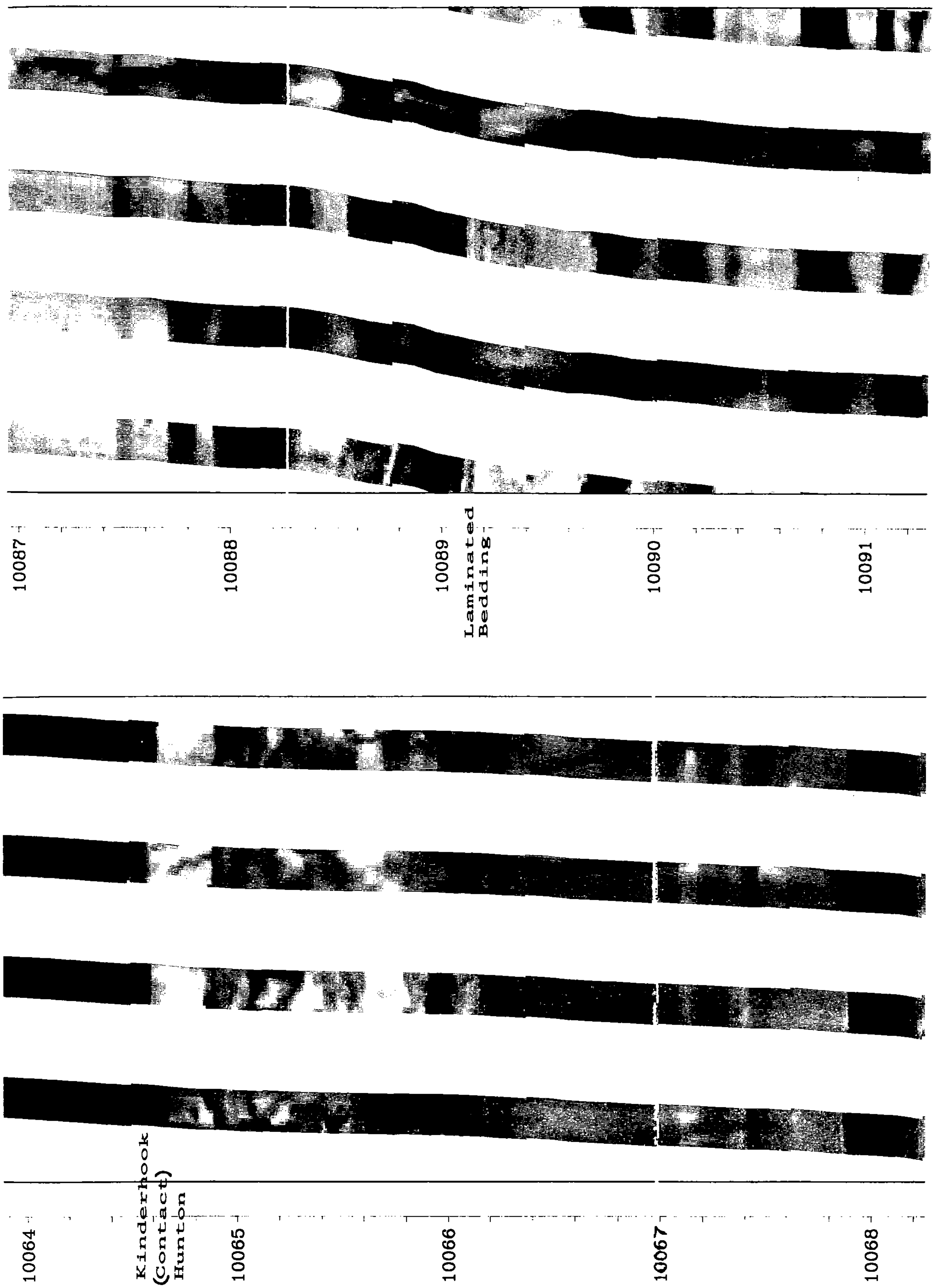

Figure 12. Formation micro-scanner log of Carlee Jane No. 202-1 well (see Fig. 11). Six selected depth intervals are shown; depths are in feet.

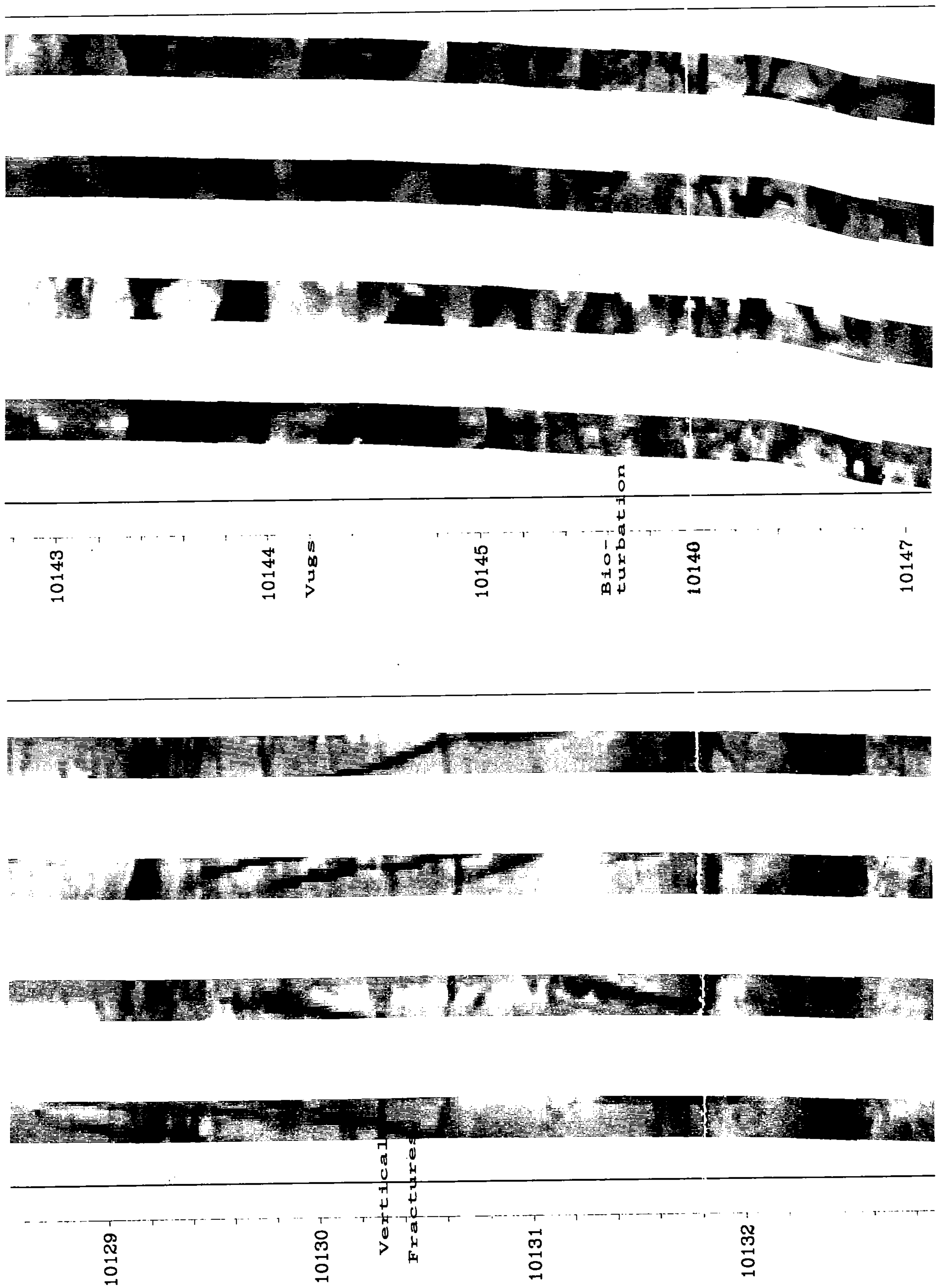

Figure 12. *Continued.*

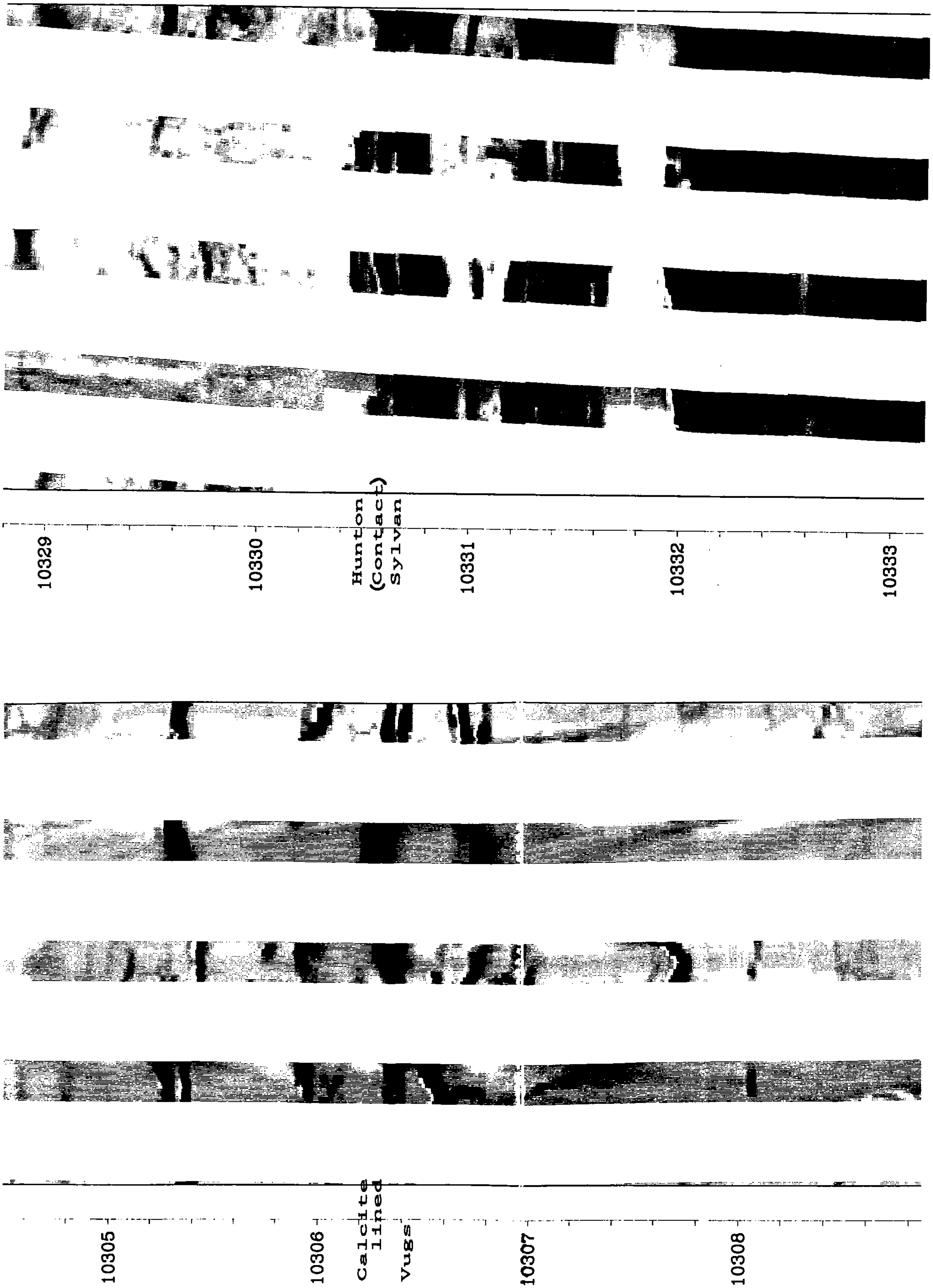

Figure 12. *Continued.*

A Regional Look at Hunton Production in the Anadarko Basin

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ABSTRACT.—Porous dolomites in the Hunton Group contain the most prolific gas reservoirs, on a per-well or per-acre basis, in the Anadarko basin of western Oklahoma. Aledo field, in Dewey County, will ultimately yield ~92 Bcfg (billion cubic feet of gas) in each square mile of Hunton reservoir, making it the most prolific field in the basin. The Phillips No. 1 Callie "B" well, located in the Aledo field, has cumulatively produced ~77 Bcfg from Hunton dolomite perforations, making it the largest gas well in the basin. Across the State line, in Hemphill County, Texas, the Gulf No. 1 Helton well was completed for 2.75 Bcfgd initial potential, calculated absolute open flow (IP CAOF), from Hunton dolomite perforations, making it the largest initial potential test in the basin.

A Hunton dolomite porosity fairway is present as a northeast-trending band across the Anadarko basin; it extends from the Hunton's truncational limit in Garfield County, Oklahoma, southwestward to the basin's southern fault limit in Wheeler County, Texas. This fairway roughly encompasses any Hunton Group reservoir, regardless of formation, where the principal lithology is dolomite. It is estimated that >80% of Hunton oil and gas production within this fairway comes from the uppermost 100 ft of Hunton, regardless of the specific formation that is present in the subcrop at its truncated top.

The Hunton dolomite porosity fairway, when superimposed on regional maps of either the Hunton structure or thickness, shows only a slight association. The closest association, however, is with a regional isopach map of the Woodford Shale, which unconformably overlies the Hunton. This emphasizes the importance that diagenesis, rather than subcrop formation or primary deposition, has on the presence of reservoir-porosity development.

PURPOSE

This paper presents regional lithological factors that affect oil and gas accumulations in carbonates of the Hunton Group of the Anadarko basin. Presented herein is a summary of a much more comprehensive regional study involving data assemblage over a 20-year period. Although anchored by subsurface control of more than 5,000 drill holes, virtually every tool available to the explorationist (i.e., seismic, surface geology, remote sensing, etc.) was utilized in understanding Hunton oil and gas fields and their relationship to regional lithologic trends.

OIL AND GAS PRODUCTION

Frank J. Adler (1971) estimated the known and future-potential Silurian/Devonian gas reserves of the Midcontinent at 16,900 Bcfg, or 38% of the total reserves. The Hunton Group, primarily the Hunton Group in the Anadarko basin, represents most of this potential. It is beyond the scope of this paper to deal with actual production numbers in the Anadarko basin, but a few highlights of Hunton exploration and production will be emphasized.

The most significant phase of Hunton drilling in the Anadarko basin occurred from 1958 to 1968, with early development of what is now encompassed in the "Sooner Trend" field. Reference is made to the Anadarko basin Hunton Group isopach map (Fig. 1), which shows the thickness and areal extent of the Hunton. The Sooner Trend production occurred (and is still ongoing, to a lesser extent) primarily in the extreme northeast corner of this map area, in Garfield, Kingfisher, and Major Counties. Here, in the Enid embayment, the Hunton has a distinct production advantage over where it occurs in the rest of the basin, due to a large regional trap formed by its truncational limit being in the structurally shallowest part in the basin.

Continued drilling to the south of the Sooner Trend revealed a gradational change in reservoir lithology; a change from porous dolomites to porous and/or fractured limestones. Continued drilling to the west and southwest of the Sooner Trend revealed that porous dolomites are present, but oil and gas traps are more structurally controlled and isolated. Thirty-five Hunton gas reservoirs, peripheral to the original Sooner Trend drilling, were studied for various geological signatures, in an attempt to define exploration techniques that could best be utilized away from the regional entrapment area of the Enid embayment. Some of the findings are listed below:

Howery, S. D., 1993, A regional look at Hunton production in the Anadarko basin, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 77–81.

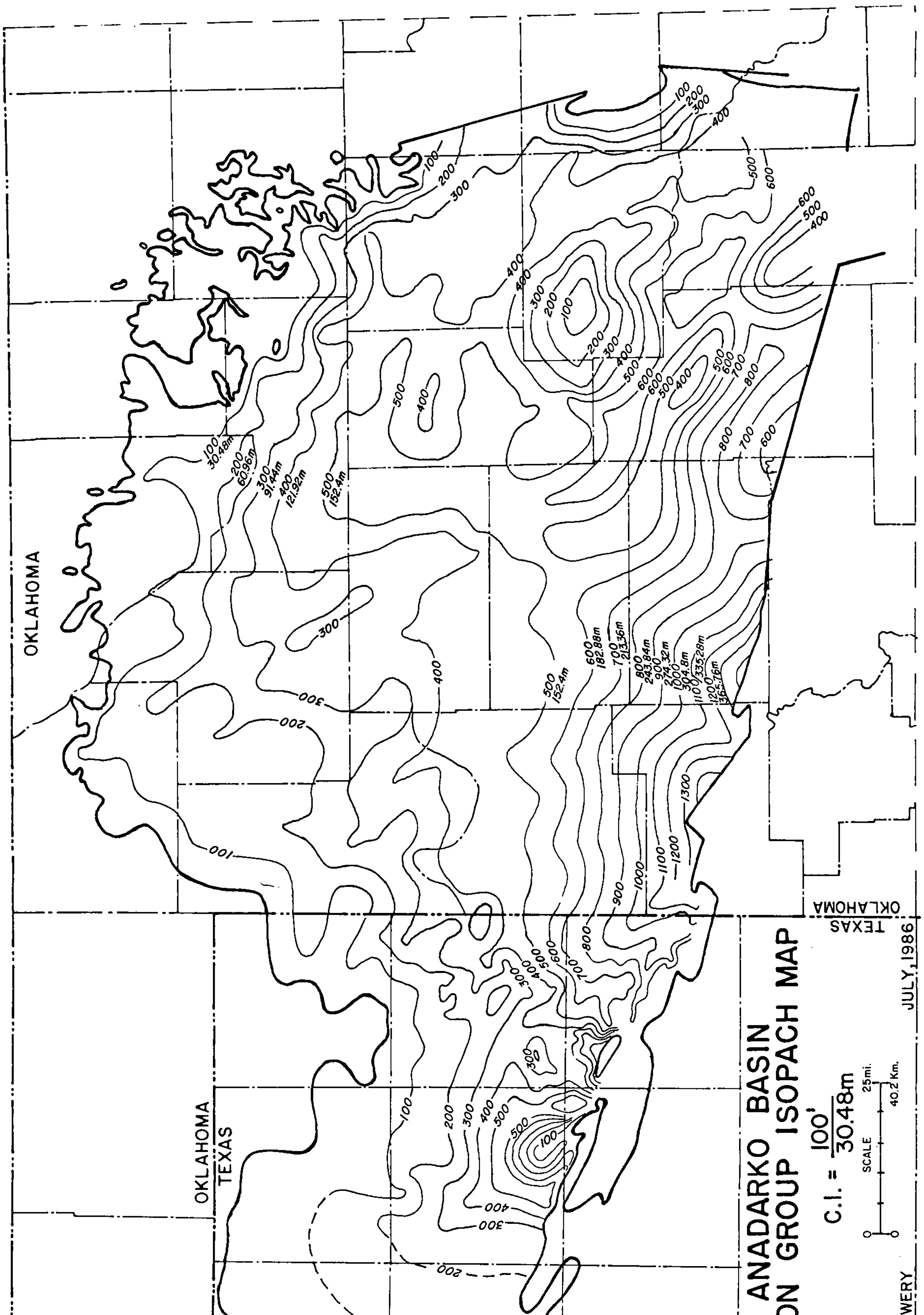

Figure 1. Thickness of the Hunton Group in the Anadarko basin.

1) The largest gas well in the basin produces from Hunton dolomite perforations (15,088-15,420 ft) in the Phillips No. 1 Callie "B" well, sec. 32, T. 16 N., R. 18 W., Dewey County, Oklahoma, which, from its completion in 3/18/69 until 1/1/93, has produced 77,323,483 Mcfg (thousand cubic feet of gas).

2) The largest initial potential in the basin is from Hunton dolomite perforations (19,512-19,910 ft) in the Gulf No. 1 Helton well, sec. 21, Block M-1, H&GNRR, Hemphill County, Texas, where, on 9/19/69, it tested CAOF 2.75 Bcfgd.

3) The average per-well ultimate recovery of gas for the Hunton Group, an average of 8.97 Bcf, is the highest of all producing intervals in the basin (Clark, 1980).

4) The most prolific reservoir, per acre, in the basin is the Hunton dolomite at Aledo field, centering in T. 16 N., R. 18 W., Dewey County, Oklahoma, where each square mile (640 acres) of reservoir will yield ~92 Bcfg.

5) The top 10 Hunton reservoirs, based on estimated ultimate recovery (EUR) of gas per 640 acres (not to be confused with spacing units), are presented in Table 1.

TABLE 1. — TOP 10 HUNTON RESERVOIRS IN THE ANADARKO BASIN

Field	County	EUR/640 acres (Bcf)
Aledo	Dewey, OK	92
Washita Creek	Hemphill, TX	64
Putnam	Dewey, OK	42
Gageby Creek	Wheeler, TX	41
Buffalo Wallow	Hemphill, TX	37
Mills Ranch/Mayfield	Wheeler, TX/ Beckham, OK	30
Arnett, N.E.	Ellis, OK	25
Cedardale, N.E.	Major, OK	22 (equiv.)
Pan Wheeler	Wheeler, TX	20
Mathers Ranch	Hemphill, TX	18

HUNTON DOLOMITE POROSITY FAIRWAY

The Woodford/Hunton (or source rock/reservoir rock) couplet in the Anadarko basin averages ~600 ft thick, in its preserved section. Considering the fact that this interval accumulated over a span of 65 ma (million years), this is an extremely slow sedimentation rate (9.2 ft/ma). By comparison, the entire Pennsylvanian interval accumulated over a span of 50 ma and averages ~8,000 ft thick and thus had a sedimentation rate of 160 ft/ma (17 times faster than Woodford/Hunton). This slow Woodford/Hunton sedimentation rate, including as many as six unconformities, is highly conducive to repetitive calcite/dolomite diagenesis. Accordingly, long-distance correlations by subsurface methods are extremely difficult.

The author chose early in the study to avoid time-consuming and questionable detailed stratigraphic correlations, and instead to focus on geologic signatures that gas fields have in common. These signatures-in-common would subsequently be the basis for a technique of detailed exploration. Probably the most obvious common aspect of the most prolific fields studied is their association within a "fairway" of dolomite porosity. This Hunton dolomite porosity fairway is defined basically as a continuous NE–SW-trending band, nearly orthogonal to the basin axis, that contains Hunton reservoir porosity, not specific to any formation, and where the dominant lithology is dolomite rather than limestone. It is estimated that >80% of known reservoirs within this fairway produce from the uppermost 100 ft of Hunton strata, regardless of the formation that is found at the subcrop. Production lower in the Hunton is limited, and then is associated only with those traps of highest structural relief.

In Figure 2, the Hunton dolomite porosity fairway is superimposed on a map showing Hunton structure and Hunton producing fields. The fairway, from its truncational limit in Garfield County (upper right corner of map), extends southwestward to the fault limit of the Hunton Group in Gray County, Texas (centered above the title block). With little exception, the aforementioned top 10 fields, based on recovery, are located along the central axis of this trend (the map scale does not permit specific field identification). This fairway is somewhat aligned along Hunton depositional strike, but is clearly at an angle to structural strike. When superimposed on the regional thickness of the Hunton Group (Fig. 1), the fairway has very little association.

Figure 3, showing the thickness of the Woodford Shale, depicts probably the strongest basis for the fairway's NE–SW trend. Note the close association of the northwest limit of the Hunton dolomite porosity fairway to the 25-ft isopach of Woodford Shale. Because there are no significant unconformities within, or close above, this interval, this isopach line is believed to portray the ancestral depositional strike of strata immediately above the truncated Hunton carbonates. The southeastward limit of the fairway would accordingly align with the Woodford shelf edge, where the Woodford abruptly increases in thickness into the basin. The stages of Hunton dolomitization and diagenesis are beyond the scope of this paper; however, the presence and trend of the Hunton dolomite porosity fairway are clearly depicted as being associated with an interval of time subsequent to Hunton deposition.

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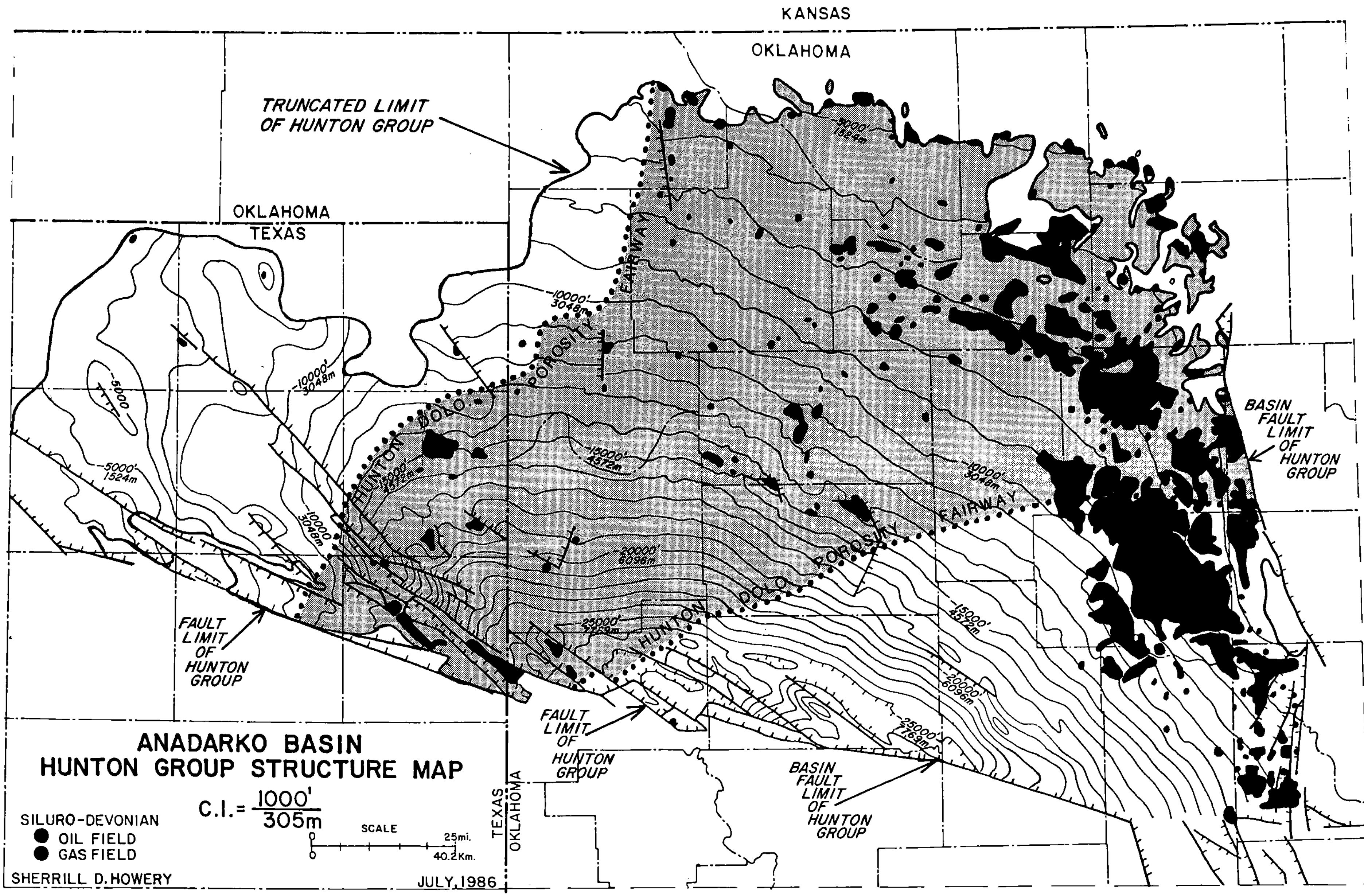

Figure 2. Structure on top of the Hunton Group in the Anadarko basin. Also shown is the Hunton dolomite porosity fairway (shaded) and Hunton oil and/or gas fields (black).

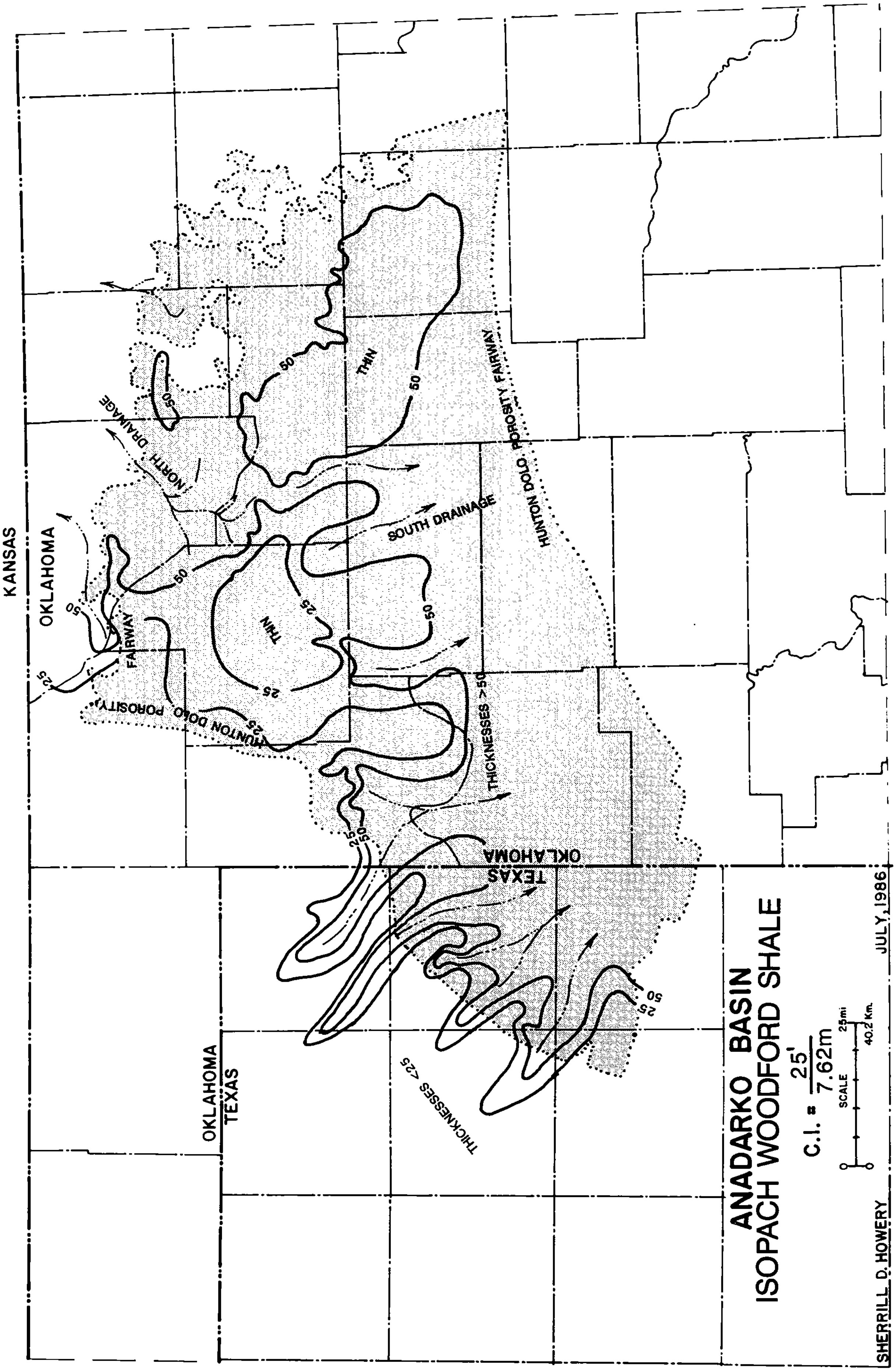

Figure 3. Thickness of the Woodford Shale in the Anadarko basin showing areas where the Woodford is >25 ft thick and the area of the Hunton dolomite porosity fairway (shaded).

Log-Derived SP Trends of the Hunton, with Possible Ramifications to Henryhouse–Chimneyhill Depositional Environments, Lincoln and Logan Counties, Oklahoma

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INTRODUCTION

The study area covers ~288 mi² in north-central Oklahoma. It is in Lincoln and Logan Counties, and includes T. 16–17 N., R. 2 E. through R. 2 W. Figure 1 shows the geographical location of the study area.

METHOD OF STUDY

Data for this study were compiled primarily from electric logs available from the Oklahoma City Geological Library. The SP curve on each log was selected as the primary means of correlation for the following reasons:

1) The SP curve is almost always included on any variety of open-hole logs, and therefore it is generally common to all.

2) Basic assumptions, such as shaliness, porosity, and permeability, can be interpreted from the SP curve.

3) Fluids within the Hunton do not affect the SP curve as dramatically as they affect the resistivity curves.

4) Only a portion of the logs available include the gamma-ray curves.

5) Within clastic sediments, the SP curve can be used to interpret basic facies or environments of deposition.

The author employs a fence-diagram technique for use in stratigraphic correlation. This method involves subdividing a large surface (e.g., a wall, large poster, or paper sheet) into map divisions, such as sections or sections, townships, and ranges. The logs to be used are then copied, annotated, and trimmed to include enough of the zones above and below the formation in question to assure not only complete inclusion of the formation, but also to determine any effects those zones above and below have on the formation in question. A stratigraphic datum common to all logs used is picked, and the logs are then secured to the fence diagram with the stratigraphic datum serving as the spot location on the grid. Figure 2 is a photograph of the fence diagram used for this study.

Approximately 460 well logs were employed on the fence diagram, with the remainder of the available logs then correlated to the fence diagram. The results of the correlations were compiled, calculated, and used to make the stratigraphic cross sections, structure map, and various isopach maps used in this paper and the Hunton core-workshop presentation.

PREVIOUS INVESTIGATIONS

The Hunton Group has been studied and described by many scholars since the earliest part of the century. The study area, either in whole or in part, has been included with several of these investigations. Amsden (1975) dealt with the Hunton in considerable detail in the subsurface, not only in the Anadarko basin but also in the surrounding provinces, including the study area. Hollrah (1978) studied the Hunton in greater detail in portions of Lincoln, Payne, and Logan Counties, which also includes the study area. Figure 3 is the stratigraphic column used for the surface exposures of the Hunton in the Arbuckle Mountains and Criner Hills areas (Amsden, 1975).

PURPOSE AND SCOPE

Evaluation of the apparent correlation of cross section A–A′, as seen on Figure 4 and whose position is shown on the location map of Figure 5, is the primary purpose of this paper. Even though the author's analysis of this area is incomplete at this time, certain correlations, trends, and interpretations are clear enough to warrant presentation at this workshop and the writing of this paper. It is the author's intent to stimulate discussions concerning the evaluation and implications of cross section A–A′ (Fig. 4). When this study area is linked to the 12,672 mi² regional study of the Woodford and Hunton in central and western Oklahoma that the author is currently pursuing for Beard Oil Co., then the questions raised here will be addressed more completely. Figure 6 is a location map showing the position of this western Oklahoma study, relative to the study area described in this paper. Figure 7 is a photograph of a portion of the fence diagram used for the larger western Oklahoma study.

Rottmann, Kurt, 1993, Log-derived SP trends of the Hunton, with possible ramifications to Henryhouse–Chimneyhill depositional environments, Lincoln and Logan Counties, Oklahoma, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 83–89.

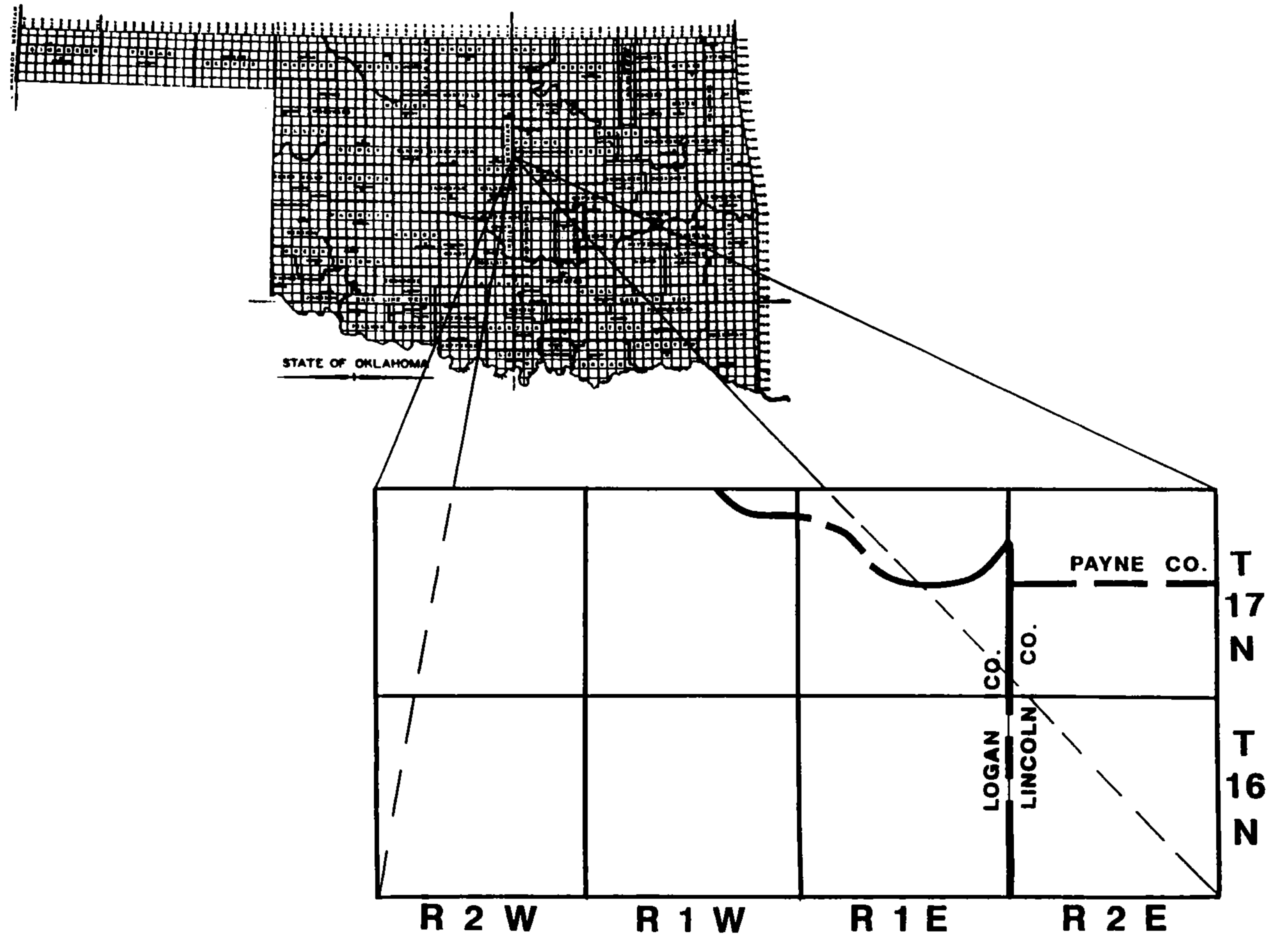

Figure 1. Location of study area in central Oklahoma.

HUNTON STRATIGRAPHY

Figure 8 is the electric log of the Universal Resources No. 3-17 Heupel well located in the NW Richland field of sec. 17, T. 13 N., R. 6 W., Canadian County, Oklahoma. Sample descriptions from the Hunton Group suggest that the following formations, in descending order, are present within this well:

Frisco Formation	8,806–8,842 ft
Bois d'Arc–Haragan Formation (undiff.)	8,842–8,920 ft
Henryhouse Formation	8,920–9,114 ft
Chimneyhill Subgroup (undiff.)	9,114–9,272 ft

(Note: The Chimneyhill will not be subdivided into its component formations in this report.)

The formation tops listed above are those established by the author and are based on sample descriptions and regional correlations to the Gulf Oil Corp. No. 1 Streeter, sec. 20, T. 13 N., R. 4 W., Oklahoma County, and the Sinclair Oil and Gas Co. No. 1 Frank Horlivy, Jr., sec. 19, T. 11 N., R. 5 W., Canadian County. Both of these wells have been cored, with core descriptions and formation boundaries having been described by Amsden (1975). There are many significant reasons why the No. 3-17 Heupel has been selected to represent the Hunton, all of which will be brought out in future discussions. For the purpose of this paper, however, the author would simply like to compare the general shape of the SP curves, as seen in Figure 9. Even though the thickness of the Hunton varies considerably, notice that the sigma shape (Σ) of the SP curve of the Heupel well is very similar, in overall character, to the sigma shape (Σ) SP curves of the wells in cross section A–A′ (Fig. 9).

Figure 2. Photograph of fence diagram used for current study.

DISCUSSION OF FIGURE 4

Figure 4 is an east–west stratigraphic cross section of the B. W. Blake No. 1 O. M. Morgan and the Ross O&G No. 1 State Land Commission wells, whose positions are spotted on the location map (Fig. 5). The overall Hunton section in the two wells has a strikingly similar SP shape. There is a strong, negative SP deflection at the top of the Hunton, which is marked as zone 1–1′. Point 2–2′ has a general curve similarity, and even seems to be at the same vertical stratigraphic point, as measured from the Sylvan datum. The lower third of the Hunton section within the two wells also exhibits similar SP characteristics. To a geologist correlating these two wells, it probably would be easy to assume similar environments of deposition, as well as equivalent correlations of the sequences zone 1–1′, point 2–2′, and zone 3–3′. The general correlation as shown would be reasonable, especially if these two wells were part of a larger, regional correlation.

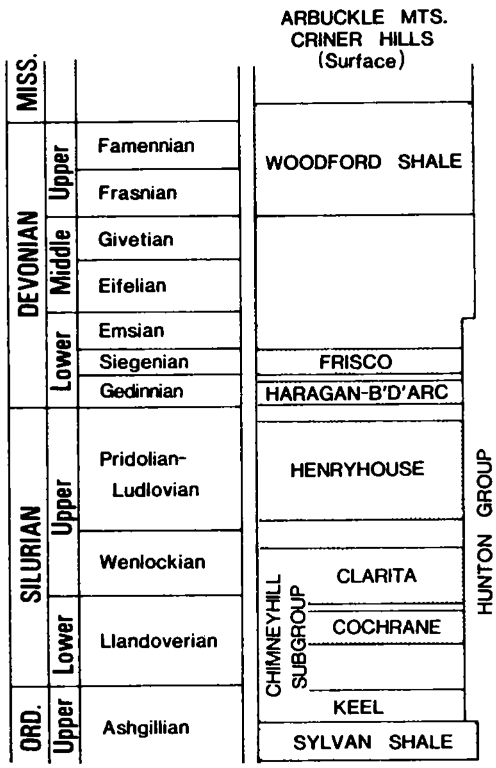

Figure 3. Geologic column of Ordovician, Silurian, and Devonian strata in Arbuckle Mountains and Criner Hills outcrops (modified from Amsden, 1975).

DISCUSSION OF FIGURE 10

Figure 10 is a series of three east–west stratigraphic cross sections within the study area, and their positions are shown on Figure 11. In general, these three cross sections reflect the overall correlation of the Hunton within the study area, as suggested from the dense well control on the fence diagram. The Woodford–Lower Mississippian/Pennsylvanian interface is not detailed on these cross sections. There also appears to be no erosion or modification of the Hunton–Sylvan contact, and this contact appears regionally conformable.

Cross sections B–B′, C–C′, and D–D′ all seem to have similar correlations. Notice the sequences labeled zone α on the eastern edge of all three cross sections. As you follow the zone α correlation westward, the base of zone α drops stratigraphically within the Hunton section, until, at the western edge of all three cross sections, zone α is at or near the bottom of the Hunton. Notice now the sequence labeled zone β on the eastern portion of cross section B–B′, C–C′, and D–D′. As you follow this correlation westward, zone β thins, until it is merely a remnant at the western edge of the cross sections. The Ross O&G No. 1 State Land Commission well from cross section A–A′ (Fig. 4) is on the western edge of cross section B–B′. The B. W. Blake No. 1 O. M. Morgan well from cross section A–A′ (Fig. 4) is a

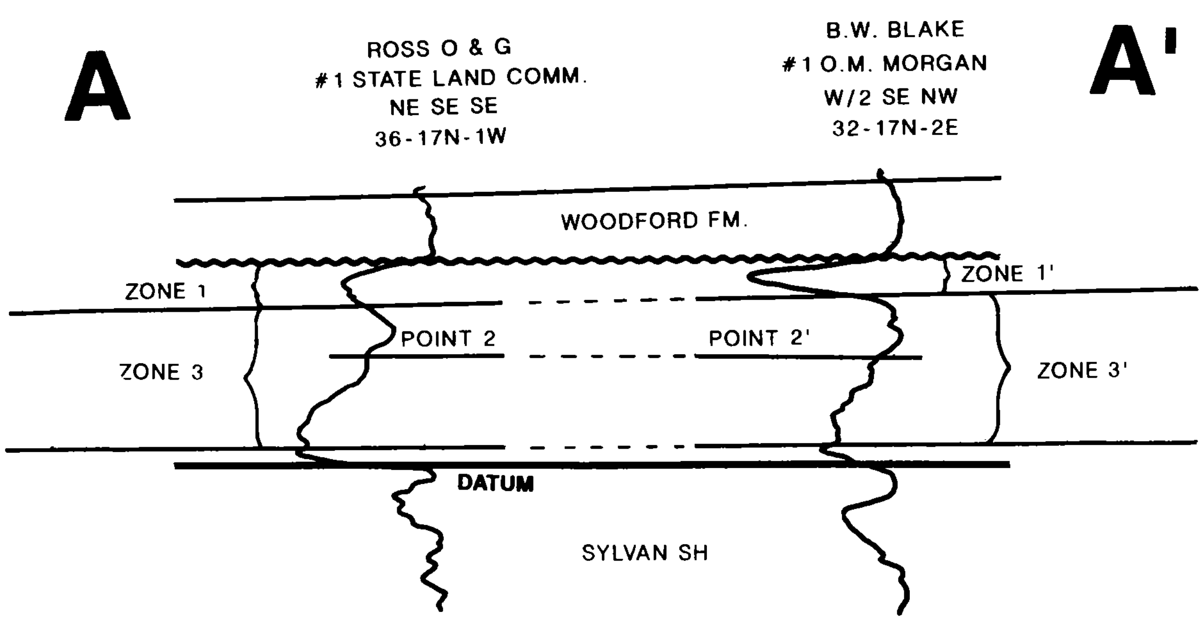

Figure 4. Stratigraphic cross section A–A′ (location map shown in Fig. 5).

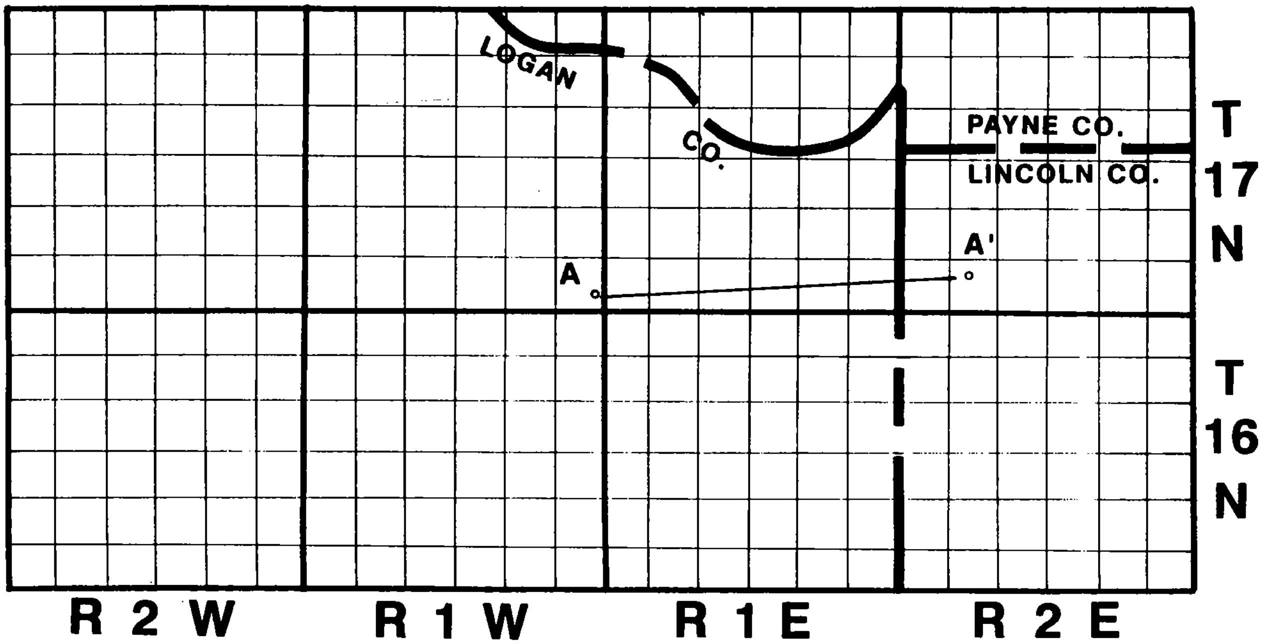

Figure 5. Location map for cross section A–A′ (see Fig. 4).

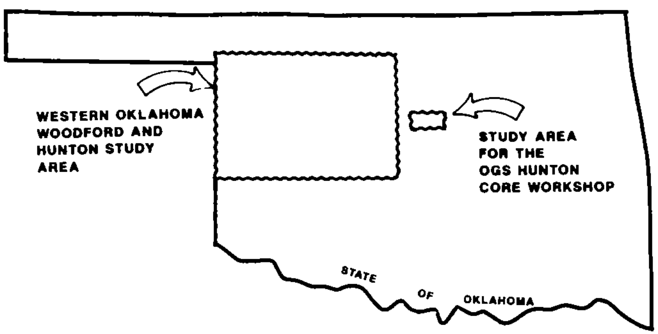

Figure 6. Location of study area for this report and larger area of western Oklahoma Woodford and Hunton study.

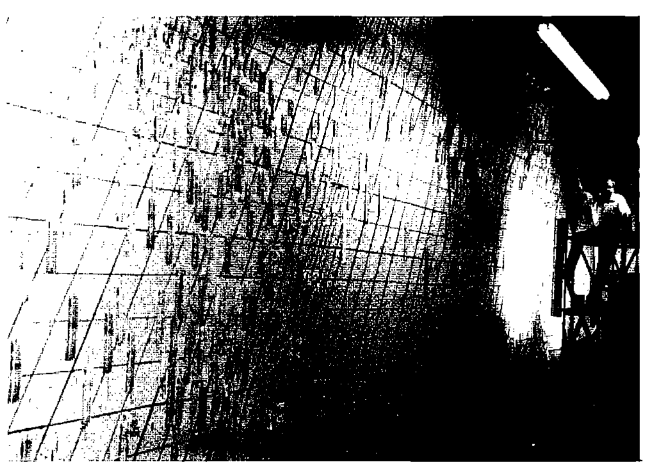

Figure 7. Photograph of a portion of the western Oklahoma Woodford and Hunton study area.

direct offset to the W. H. Martgan No. 1 Orner, which is on the eastern edge of cross section B–B′. These correlations, as suggested from cross sections B–B′, C–C′, and D–D′, are in direct conflict with the apparent correlation of zone 1–1′, point 2–2′, and zone 3–3′, as shown on cross section A–A′ of Figure 4.

DISCUSSION OF FIGURE 12

Figure 12 is a revised version of cross section A–A′ that is shown in Figure 4, incorporating all of the densely spaced Hunton well control in between. Zone α in the Ross well does not correlate stratigraphically to zone β in the Blake well, nor does point 2 correlate to point 2′. However, the similarity of the SP curves in the two wells has not changed. The apparent similarity of the SP shape perhaps represents a cyclical environment of deposition for the rock-stratigraphic sequences in these two wells. If the sequences zone α and zone β on revised cross section A–A′ (Fig. 12) are composed of allochthonous materials, then these sequences take on the appearance of more-typical, clastic-type depositional environments, such as progradational fan or deltaic environments. At the time of this writing, the core and sample work have not been completed. However, various scout tickets and sample logs have referred to portions of the Hunton within this area as being "detrital."

CONCLUSIONS

The author has noticed, over the years, that upon reviewing Hunton prospects and observing correlations annotated on electric logs, many geologists tend to equate the general sigma shape (Σ) character of the SP as an indication of the presence of the various Hunton formations. The top, negative-deflected portion is often attributed to the Frisco, Bois d'Arc, and/or Haragan Formations. The cen-

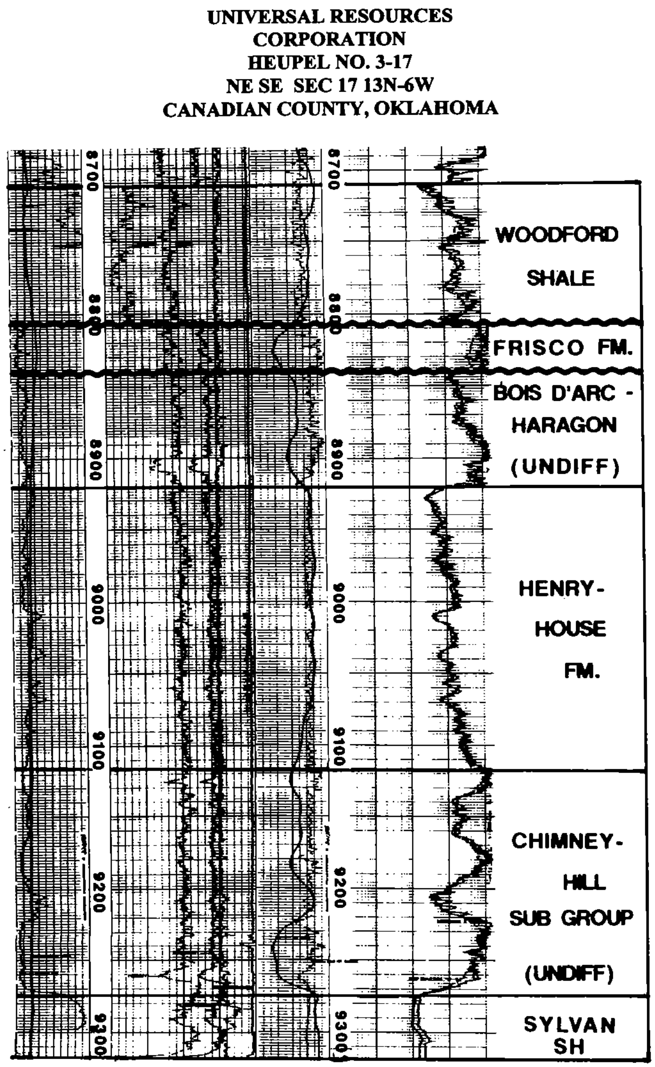

Figure 8. Example log of Woodford–Hunton–Sylvan section in central Oklahoma.

tral, positive-deflected portion is generally referred to as Henryhouse, and the bottom, negative-deflected portion is thought to be Chimneyhill. The No. 3-17 Heupel well (Fig. 8) is an example where this overall tendency is true. However, the SP shape of wells in cross section A–A′ (Fig. 12) also have this sigma shape (Σ) character, and indeed the author has noticed attempts by others to force various formation names to these curves on wells within the study area. The revised cross section A–A′ (Fig. 12) demonstrates a cyclical mode of deposition; it indicates that there are not several formations present, but that the entire Hunton section within the study area is probably only a portion of, or a member of, a single formation. No attempt will be made in this paper to speculate as to which formation is present within the study area.

The cyclical nature of deposition within the study area is strongly evident, based upon using a fence-diagram approach to regional stratigraphic correlations. It is the author's intent to connect the stratigraphic geometries in this study with those geometries observed in the regional study of the Woodford and Hunton of central and western Oklahoma, as shown on Figures 6 and 7.

ACKNOWLEDGMENTS

Sincere appreciation is extended to several individuals who have contributed to this paper: to Mr. Terry Hollrah, for his cooperation in obtaining cores and core information for the study area; to Mr. Guy Franson and Mr. Bob Powell of Universal Resources, for their cooperation in making their files available; and to Mr. Don Terrell and Mr. David

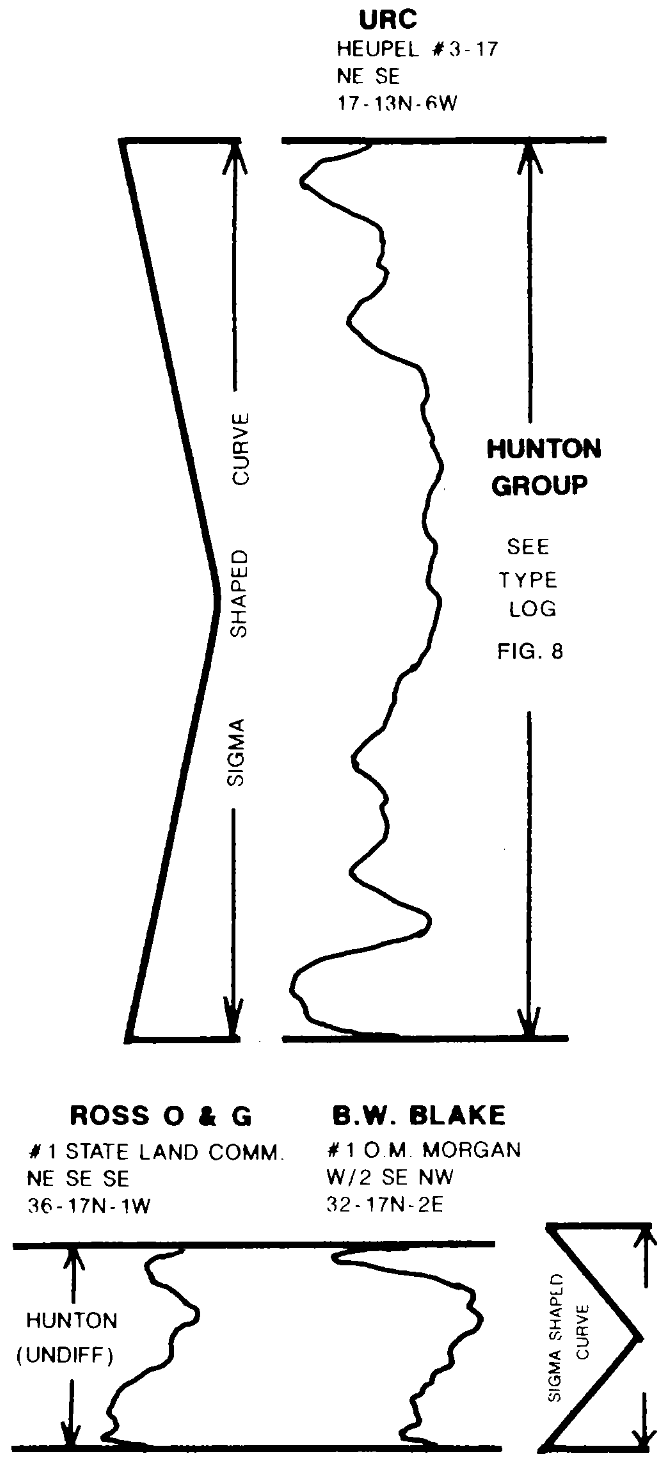

Figure 9. Comparison of sigma shape (Σ) of SP curves for wells shown in Figures 4 and 8.

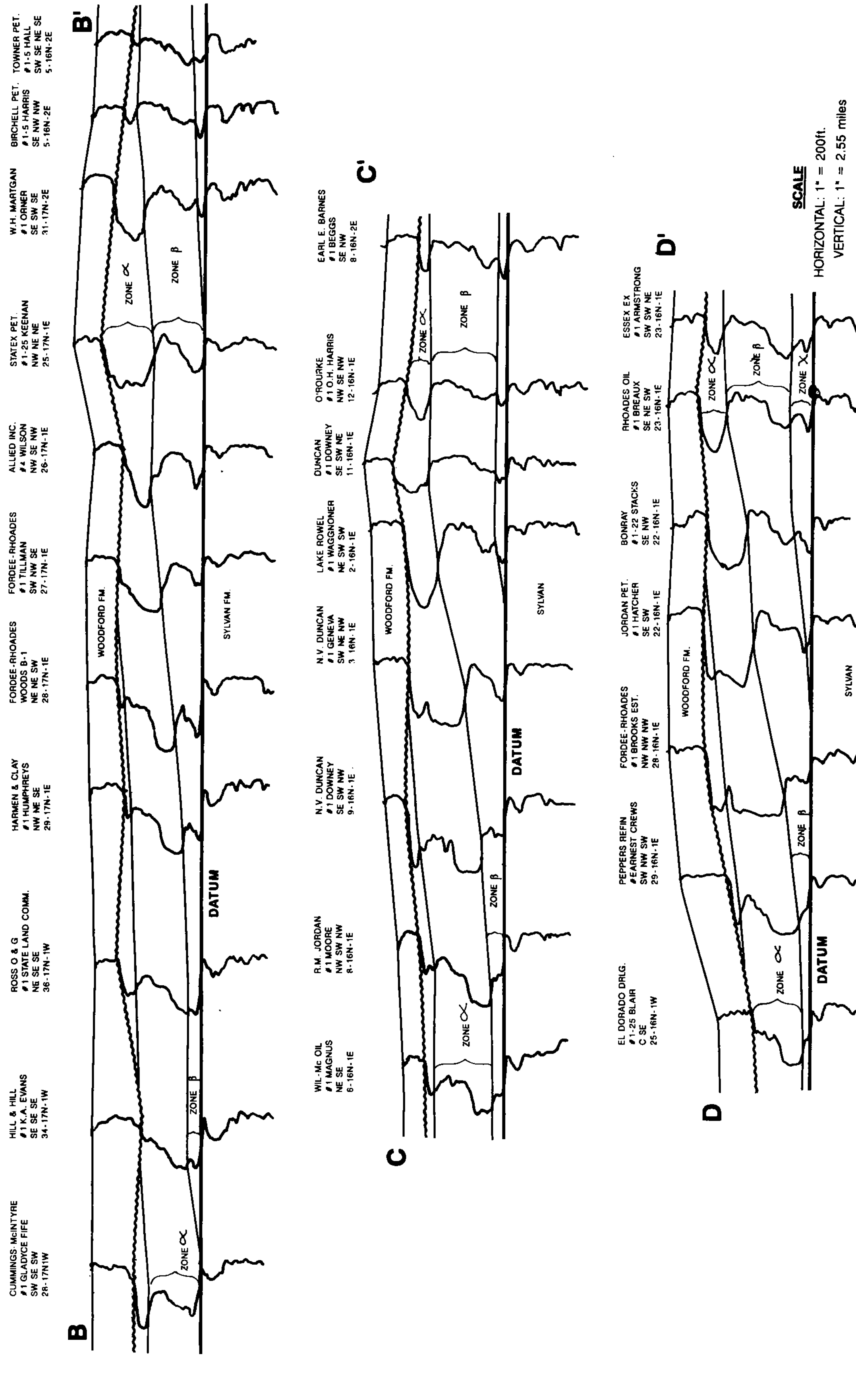

Figure 10. Cross section B–B′, C–C′, and D–D′ (location map shown in Fig. 11).

Crutchfield of Beard Oil Co., and Mr. Lee C. Lamar, retired, for their insight and review of this paper. I am deeply grateful to Mr. Walter Parrish of Beard Oil Co. for his patience and encouragement in pursuing these studies. And I am especially grateful to Mr. Bill Beard of Beard Oil Co. for his total support of these projects.

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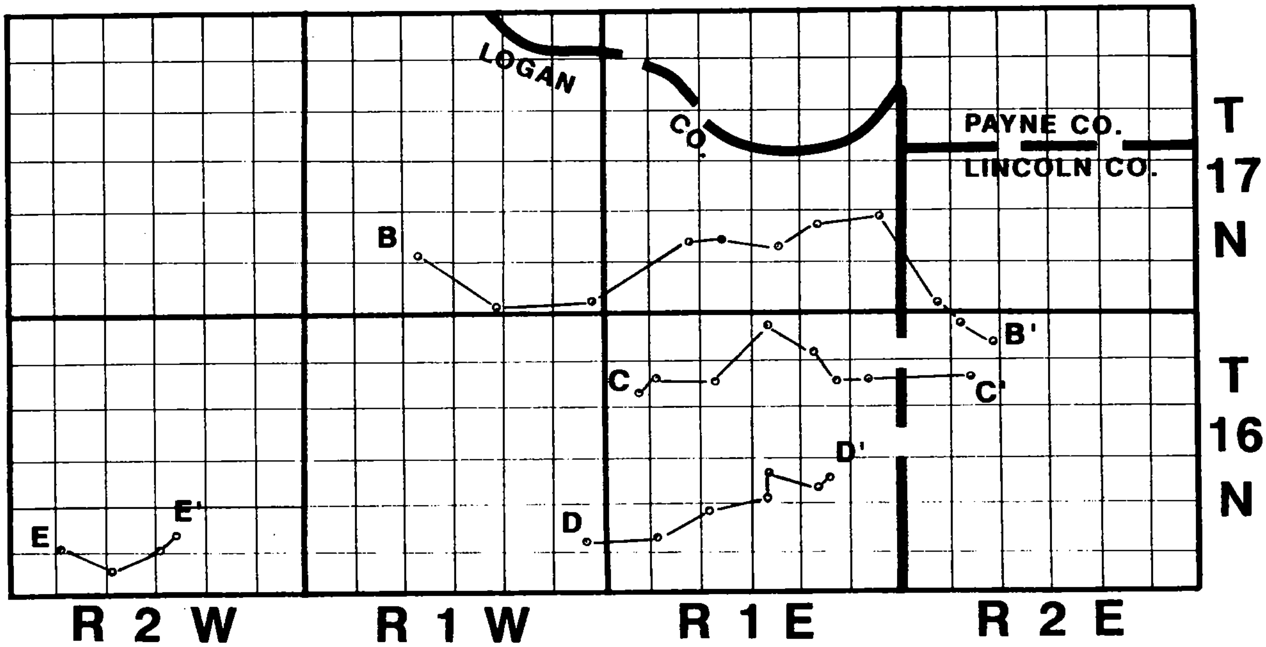

Figure 11. Location map for cross sections B–B′, C–C′, and D–D′ (see Fig. 10). Cross section E–E′ was discussed only at Hunton Group core workshop.

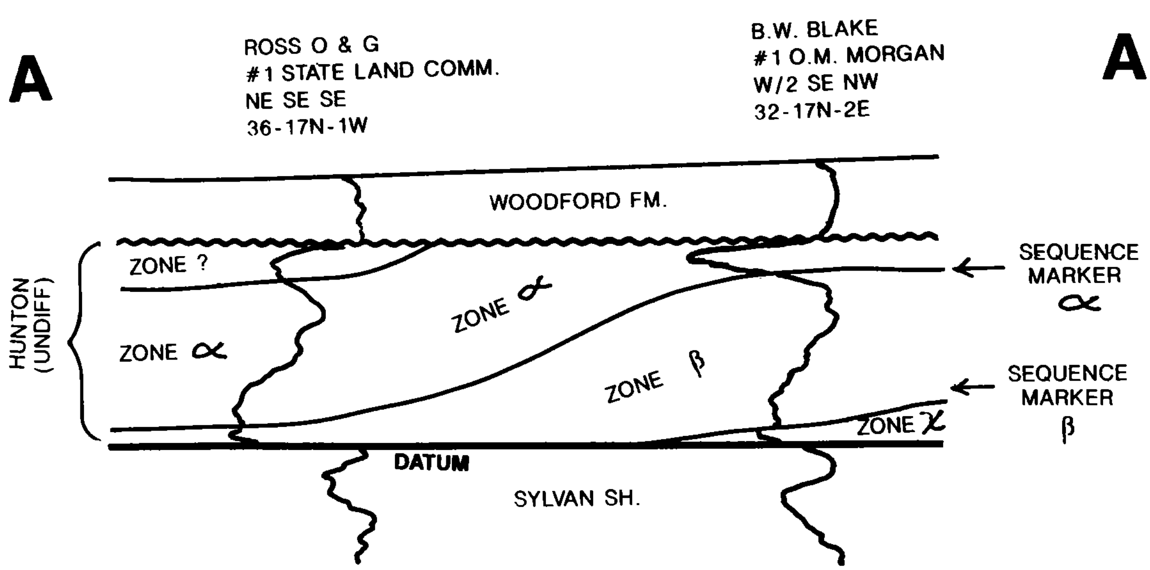

Figure 12. Revised interpretation of cross section A–A′ (see Figs. 4 and 5).

Depositional and Diagenetic Character of Hunton-Equivalent Rocks in the Permian Basin of West Texas

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ABSTRACT.—Hunton Group-equivalent rocks in the Permian Basin of West Texas and New Mexico constitute an important hydrocarbon-bearing succession. Production from nearly 650 reservoirs developed in these Silurian and Devonian carbonates and cherts totals almost 2 billion bbl of oil.

The Fusselman Formation, which forms the lowest part of the section, comprises Late Ordovician to Middle Silurian age rocks that were deposited on a stable, broad, low-gradient, shallow-water platform. Variations in depositional facies are principally the result of paleotopography. Reservoirs are principally developed in basal ooid grainstones and overlying pelmatozoan packstones, both of which are areally extensive. Multiple sea-level-fall events resulted in punctuated early diagenesis that has contributed to reservoir heterogeneity.

Middle Silurian drowning and downwarping of the Fusselman platform was followed by deposition of Wristen Group carbonates that include outer ramp/slope to basin mudstones (Wink and Frame Formations) and complex platform-margin and inner-platform suites of shallow-water carbonates (Fasken Formation). Reservoirs in these rocks are developed exclusively in platform margin, coral/stromatoporoid-buildup successions and in cyclic, laterally variable, inner-platform facies of the Fasken Formation. Porosity in inner-platform rocks is related to diagenesis associated with periodic, high-frequency sea-level fall and to long-term exposure during the Middle Devonian.

Early Devonian sea-level rise resulted in filling of the Middle Silurian basin by transgressive, slope/basin, spiculitic chert and progradational, highstand, skeletal carbonate of the Thirtyone Formation. Thirtyone Formation reservoirs are developed in cherts whose distribution is the result of cyclic sea-level rise and fall and gravity mass transport, and in aggradational, platform-top, grain-rich carbonates that underwent leaching during regional Middle Devonian uplift.

Hunton Group equivalents in the Permian basin resemble their Oklahoma counterparts, but differ in several significant ways. Comparative study of the two temporally equivalent successions can greatly improve the understanding of the depositional and diagenetic processes that have contributed to reservoir development in each.

INTRODUCTION

Carbonate and chert rocks of Silurian and Devonian age constitute a thick (≤2,000 ft thick) hydrocarbon-bearing succession in the Permian basin of West Texas and New Mexico (Fig. 1). Nearly 2 billion stock-tank barrels (STB) of oil have been produced from the approximately 650 reservoirs developed in these rocks (Fig. 2), and current estimates suggest that another 2 billion bbl of mobile oil remain (Ruppel and Holtz, in press).

Hunton Group-equivalent rocks in West Texas and New Mexico consist of a thick succession of carbonate and chert rocks, and a generally much thinner interval of overlying shales that are assigned to the Woodford Formation (Jones, 1953; Wilson and Majewski, 1960; McGlasson, 1967). Although the rocks that make up the Silurian and Devonian reservoir succession in West Texas and New Mexico are equivalent in age to the Hunton Group (Fig. 3), they exhibit important differences in lithology and reservoir development. This report describes these rocks and

Ruppel, S. C., 1993, Depositional and diagenetic character of Hunton-equivalent rocks in the Permian basin of West Texas, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 91–106.

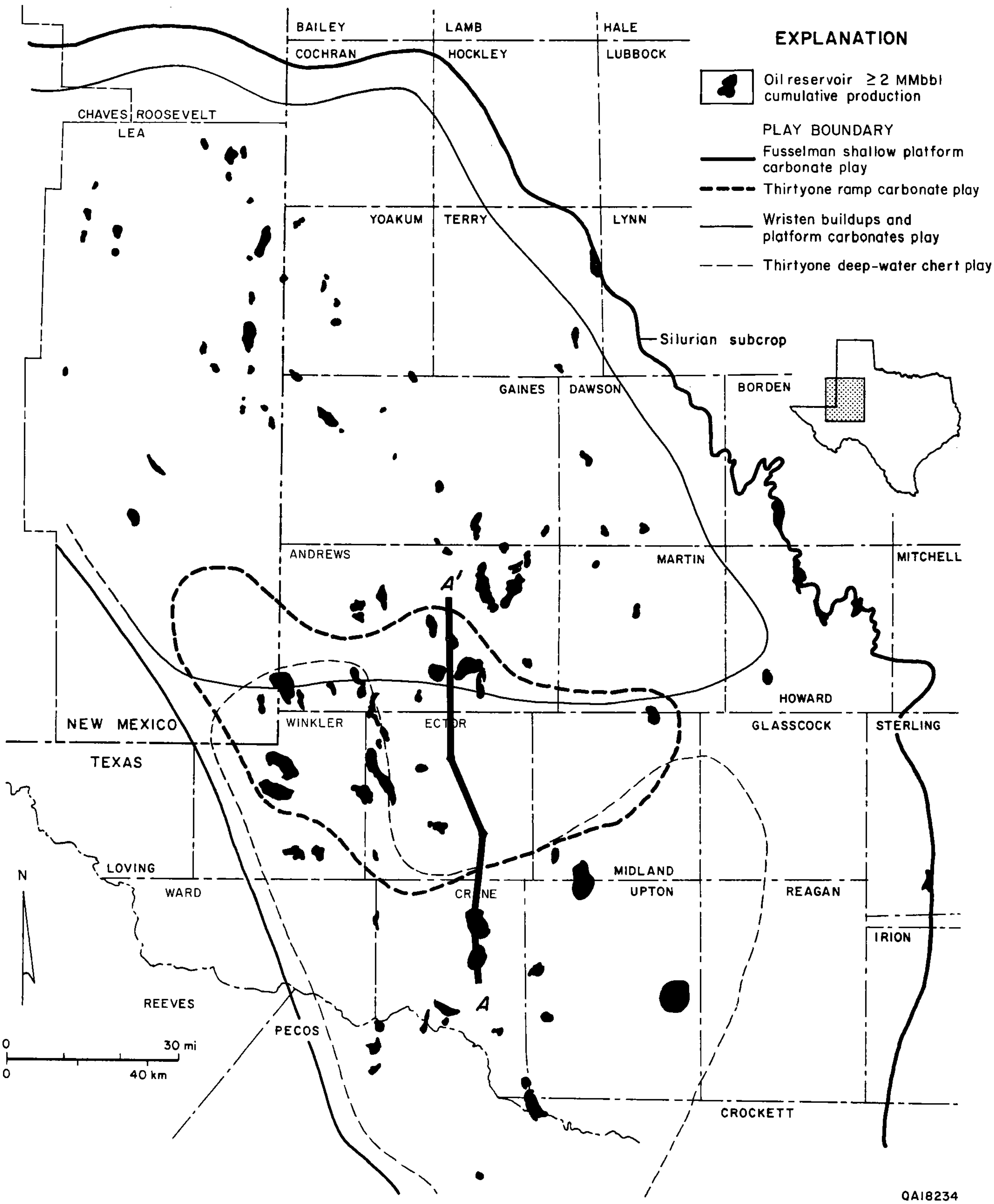

Figure 1. Map of West Texas and New Mexico showing the distribution of major hydrocarbon-producing reservoirs in Hunton-equivalent strata. Cross section A–A′ is shown in Figure 4.

presents models for their deposition, diagenesis, and reservoir development. Three cores are displayed to illustrate the key features of each major reservoir unit: the Fusselman Formation (Keel–Cochrane equivalent), the Fasken Formation (Henryhouse equivalent), and the Thirtyone Formation (Haragan–Frisco equivalent).

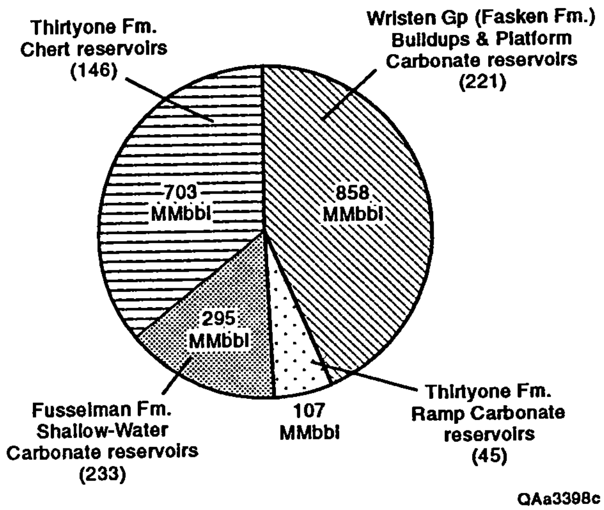

Figure 2. Cumulative production from Hunton-equivalent reservoirs in the Permian basin of West Texas and New Mexico.

STRATIGRAPHY AND FACIES

Only the Hunton Group-equivalent, hydrocarbon-bearing, carbonate and chert sections of Late Ordovician to Early Devonian age are dealt with in this report; for a recent treatment of the Upper Devonian Woodford Formation, see Comer (1991). Pre-Woodford Silurian and Devonian rocks in the Permian basin can be divided into three major lithostratigraphic units: the basal Fusselman Formation, the overlying Wristen Group, made up of the Fasken, Frame, and Wink Formations, and the Thirtyone Formation (Figs. 3,4). Reservoir development in Hunton-equivalent rocks in West Texas and New Mexico is restricted to the Fusselman Formation, Fasken Formation (Wristen Group), and Thirtyone Formation. The Frame and Wink Formations (Wristen Group) are deeper water, shaly, mud-rich equivalents of Fasken platform carbonates and are not productive.

Fusselman Formation

Although traditionally assigned to the Silurian (Wilson and Majewski, 1960; McGlasson, 1967; Harbour, 1972), recent studies (Barrick and others, 1993) have shown that the Fusselman Formation comprises rocks of Late Ordovician and Early Silurian age, and it is equivalent to the Keel and Cochrane Formations of the basal Hunton Group (Fig. 3).

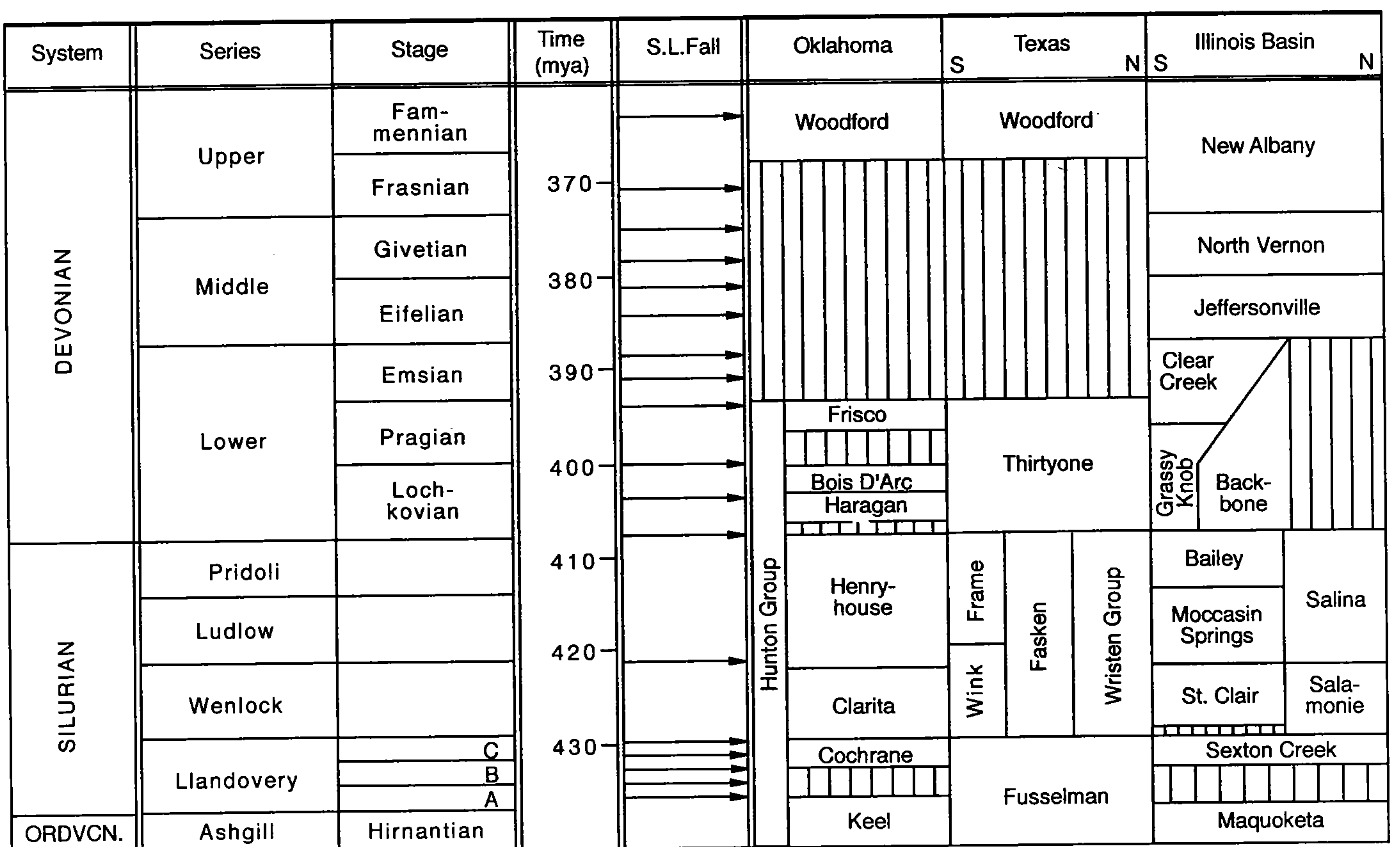

Figure 3. Correlation of Silurian and Devonian strata in West Texas with the Hunton Group of Oklahoma. Age dates are from Harland and others (1989).

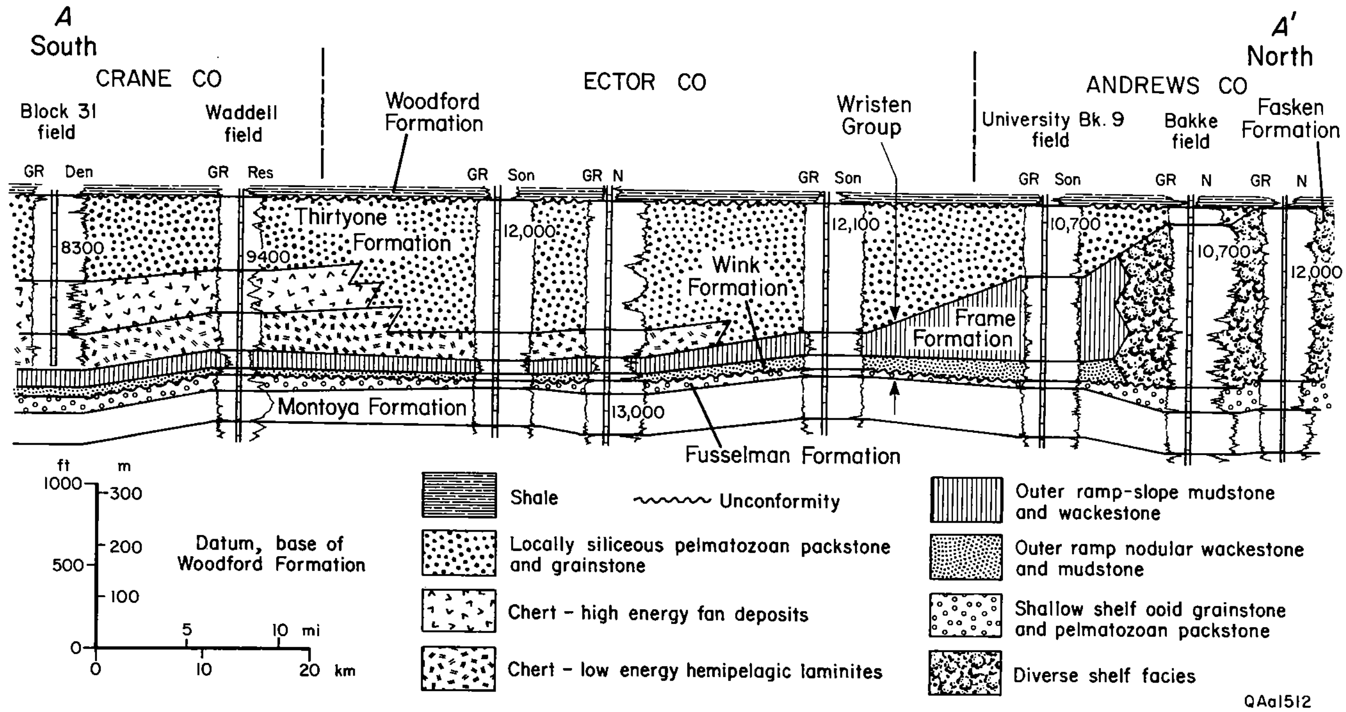

Figure 4. Cross section showing general stratigraphic relationships of Silurian and Devonian rocks in West Texas and New Mexico. Line of section is shown in Figure 1.

Facies

The Fusselman comprises a diverse succession of shallow-water carbonate facies (Fig. 5); however, reservoirs are principally developed in just two of these: a basal, thin (<10 ft thick) ooid grainstone to packstone, and an upper, thicker interval of pelmatozoan grainstone and packstone. The basal ooid grainstone is lithologically similar to the Keel Formation of the Hunton Group, and recent biostratigraphic work suggests that the two are equivalent (Barrick and others, 1993). These rocks are typically well sorted, locally cross bedded, and commonly display well-preserved primary interparticle porosity and good permeability (Fig. 6A). Meniscus and pendant cements partially fill pore space (Fig. 6B). These cements, combined with overlying fenestral mudstone that locally contains shale-filled solution/erosion pits, document exposure and meteoric diagenesis at the top of the lower Fusselman. Locally, especially near the eastern subcrop margin, the ooid facies grades into mud-dominated rocks associated with paleotopographic highs (Geesaman and Scott, 1989; Garfield and Longman, 1989).

The upper Fusselman reservoir unit contains well-sorted and locally cross-bedded, grain-dominated packstone composed predominantly of pelmatozoans and subordinate bryozoans, brachiopods, ostracodes, corals, and mollusks (Fig. 6C). The basal part of this unit locally grades laterally into skeletal wackestone containing pelmatozoans, brachiopods, and ostracodes. These deposits are similar to correlative rocks in Oklahoma (Cochrane and Clarita Formations; Amsden and others, 1980), indicating the development of a continuous, broad, shallow platform during the Early Silurian. Multiple successions of sediment- and cement-filled geopetal voids exhibiting cross-cutting relationships are locally common within the upper Fusselman (Fig. 6D). These voids were likely produced by secondary leaching and subsequent sediment infill and cementation.

Depositional Environment

The Fusselman Formation documents deposition on an open-marine, shallow-water carbonate platform that probably formed during Late Ordovician time. Underlying Montoya Group deposits (Fig. 4) represent the earlier development of this platform, which extended across West Texas and New Mexico during the Late Ordovician and Early Silurian. Basal Fusselman ooid grainstones represent deposition in relatively high-energy tidal-flat to shallow-subtidal conditions. The extent of these deposits indicates that the platform was broad and flat, and extended across much of the southern Midcontinent. Vertical facies successions suggest an overall deepening through Fusselman deposition. Upper Fusselman pelmatozoan facies document more open-marine conditions across this broad platform.

Stratigraphic and diagenetic patterns in the Fusselman indicate that deposition was punctuated by episodic rise and fall of relative sea level. Johnson (1987) documented four major sea-level falls during the Llandovery (Fig. 3). Comparison of these data with the biostratigraphically well-constrained Chimneyhill Subgroup in Oklahoma suggests that: (1) the lower ooid grainstone member of the Fusselman (Keel equivalent) is separated from the upper pelmatozoan facies (Cochrane equivalent) by a long-duration hiatus (approximately 3–4 m.y.), and (2) the upper Fusselman pelmatozoan facies experienced at least two significant falls in relative sea level (Fig. 3; Amsden and Barrick, 1986). Both of these conclusions are supported by diagenetic features recognized in Fusselman core succes-

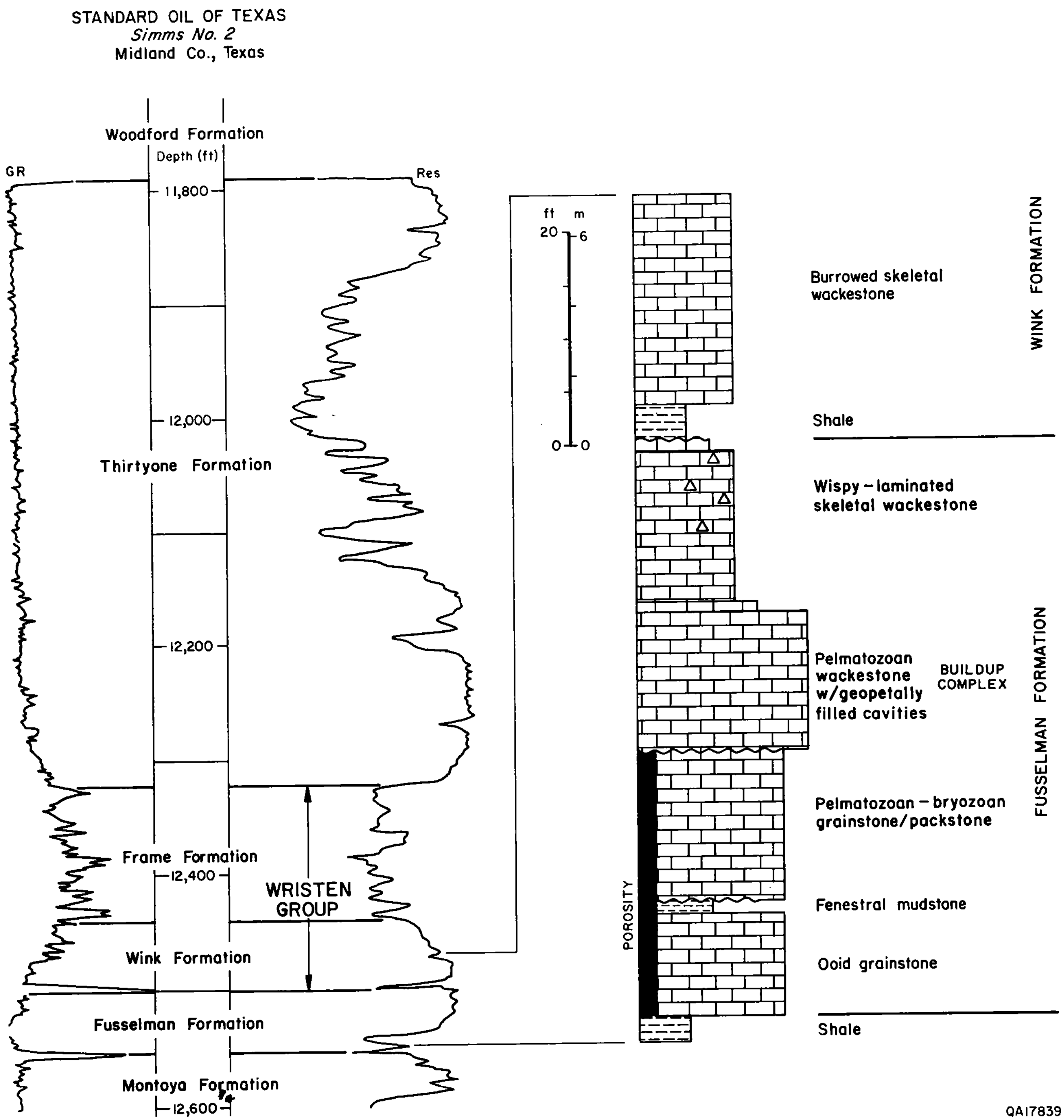

Figure 5. Stratigraphy and facies of the Fusselman Formation in West Texas.

sions. The multiple generations of geopetally filled vugs in the pelmatozoan facies are likely a record of these late Llandoverian sea-level falls.

Reservoir Development

Fusselman reservoirs are associated with three main types of porosity development: primary intergranular, secondary intercrystalline to vuggy, and cavernous. Primary intergranular porosity is restricted to basal Fusselman ooid facies. Other pore types are primarily associated with the upper Fusselman pelmatozoan facies. Secondary intergranular porosity is developed in the upper Fusselman due to leaching of mud in packstones. These pores and generations of cross-cutting vugs (Fig. 6D) may have developed at repeated sea-level-fall events during Fusselman deposition, or may be related to post-Fusselman longer-duration hiatuses. Major karsting associated with the development of cavernous porosity in the Fusselman (e.g., Mazzullo and Mazzullo, 1992) is the result of meteoric diagenesis, probably associated with one or more of these longer-duration, post-Fusselman exposure events. The top of the Fusselman commonly displays a variety of dissolution features where the unit is overlain by post-Silurian rocks. Top seals for Fusselman reservoirs are formed by

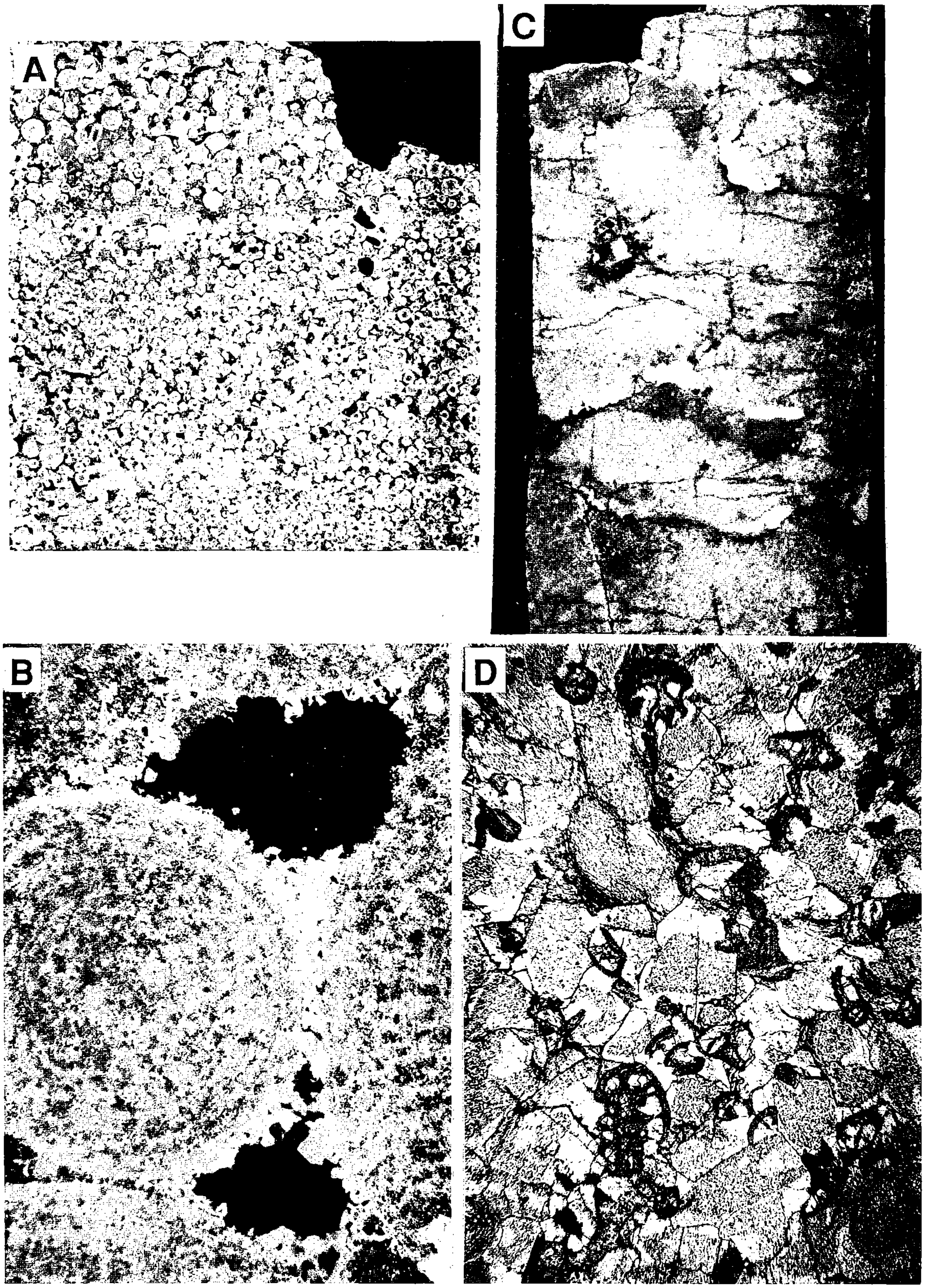

Figure 6. Typical Fusselman facies in West Texas. *A*—Basal Fusselman ooid grainstone, similar to the Keel Formation. Slab is 6 cm wide. Standard of Texas, Simms No. 2, Midland County, Texas. *B*—Cathode-luminescence photomicrograph of A showing interparticle pore development and fringing cement. Field of view is 2 × 3 mm. *C*—Photomicrograph of upper Fusselman pelmatozoan packstone facies. Sinclair, Emma Cowden "C" No. 31, Andrews County, Texas. Field of view is 3.5 × 7 mm. *D*—Slab photograph of upper Fusselman pelmatozoan packstone facies showing abundant geopetally filled vugs. Many of these vugs are filled with sediment and cement, but some are open and contain hydrocarbons. Seaboard Meiners No. 1, Upton County, Texas. Depth: 12,690 ft. Slab is 8 cm wide.

deep-water Wristen Group (Silurian) mudstones and shales, or by the Upper Devonian Woodford Formation. Reservoirs are located over late Paleozoic structures or along the truncated Silurian subcrop margin.

Fasken Formation

The Fasken Formation, which represents the carbonate-platform facies part of the Wristen Group (Fig. 4; Ruppel and Holtz, in press), ranges from <200 to 1,400 ft in West Texas. Throughout most of the region, the Fasken is overlain by the Upper Devonian Woodford Formation, except where the latter has been removed by late Paleozoic erosion. Although dating of the Fasken is imprecise, it is probably a shallow-water equivalent of the Henryhouse of the Hunton Group (Fig. 3). Lithologically similar platform equivalents of the deeper-water Henryhouse have apparently been removed by erosion in Oklahoma.

Facies

The Fasken Formation comprises a highly diverse assemblage of carbonate-platform and platform-margin lithofacies. The unit can be subdivided into two general facies complexes: (1) platform-margin skeletal wackestones to grainstones and boundstones; and (2) inner-platform mudstones, pellet and skeletal wackestones, and grainstones. A typical platform-margin succession is illustrated in Figure 7. The base of the succession comprises dark-colored skeletal wackestones that contain poorly sorted skeletal debris, including stromatoporoids, corals, and pelmatozoans. These rocks pass upward into a section of wackestones that contains more abundant and larger fragments of stromatoporoids and corals (including *Halysites*). Stromatoporoids comprise both broken hemispherical forms that display all orientations, and thin, laminate stromatoporoids in growth position (Fig. 8A). Overlying these rocks are pelmatozoan/stromatoporoid packstones (Fig. 8B). Pelmatozoan debris is well sorted, but stromatoporoids are variable in size throughout the unit. These rocks are succeeded by a thin interval of relatively well-sorted coral/bryozoan mudstone that exhibits considerable interparticle porosity (Fig. 8C). The Fasken is capped in this section by coral framestone composed of small stick corals, bryozoans, and relatively uncommon stromatoporoids. These deposits contain geopetal cavities filled with sediment and, less commonly, cement (Fig. 8D). Elsewhere, such buildups are locally capped by ooid grainstone. Fasken platform-margin buildups in Texas and New Mexico are similar to well-described outcropping Silurian buildups in the Illinois basin, in terms of both facies patterns and fauna (Lowenstam, 1950; Ingels, 1963; Wilson, 1975).

Inner-platform Fasken deposits exhibit a broad spectrum of facies successions, most of which are highly cyclic. Low-accommodation successions are characterized by cyclic tidal-flat sections, in which porosity is associated with dolostone at cycle tops. Like shallow-water platform rocks in the Permian, porosity evolution in these rocks is the result of early diagenesis associated with exposure during sea-level-fall events. Cyclicity is also apparent in more-open-marine, subtidally dominated, inner-platform successions. These subtidal cycles are typically capped by grain-dominated facies or buildups. Porosity in these rocks is usually associated with leaching of grain-dominated facies.

Depositional Setting

The Wristen Group documents drowning of the Early Silurian platform in West Texas and New Mexico during the Middle Silurian. Fasken Formation rocks record renewed platform growth in the northern part of the region, whereas equivalent, nonproductive Frame and Wink Formations accumulated in basin and slope settings to the south. Vertical and lateral variations in Fasken facies patterns are a function of the combined effects of high-frequency sea-level rise and fall and paleotopography.

Reservoir Development

Reservoir porosity in Fasken buildups is typically developed as intergranular pores in capping grain-dominated facies (skeletal and ooid grainstones), and in leached, dolomitized intervals at the top of the succession. These reservoirs, which are mostly located within ~20 mi of the platform margin, have accounted for only a relatively small percentage of total Fasken production (Ruppel and Holtz, in press).

Most Fasken reservoirs are developed in inner-platform successions where porosity has been created (meteoric leaching) or preserved (by dolomitization) at cycle tops. These reservoir zones are usually fabric selective. Locally, there is also cavernous porosity development associated with karsting and dissolution at the top of the Fasken section. Although many Fasken karst successions have been subsequently filled by collapse, sediment infill, and cementation, some karst-related porosity remains (Entzminger and Loucks, 1992; Troschinetz, 1992a,b).

Thirtyone Formation

The Thirtyone Formation is as much as ~1,000 ft thick in the Permian basin. The unit thins to the north due to Middle Devonian erosion and is truncated at about the position of the Silurian platform margin (Fig. 4). Most faunal data suggest that the Thirtyone Formation is Early Devonian (Lockhovian to Pragian) in age (Wilson and Majewski, 1960; Barrick and others, 1993; Ruppel and Holtz, in press). Equivalent Hunton Group rocks in Oklahoma include the Haragan, Bois d'Arc, and Frisco Formations of the upper Hunton Group (Fig. 3; Amsden, 1980).

Facies

Thirtyone strata contain both chert and carbonate facies. Chert facies are most abundant downsection and in the southern part of the area; carbonate facies are more abundant upsection and to the north (Fig. 4).

Cherty Thirtyone lithofacies are particularly well developed in the depocenter of the Thirtyone Formation. The stratigraphic section in this area (Fig. 9) displays a spectrum of lithofacies that encompasses most of the range seen in the Thirtyone across the basin. Generally, this section comprises an upward-shallowing succession that is chert-rich at the base and carbonate-rich at the top. Four facies, listed in order of generally decreasing water depth and chert/limestone ratio, can be recognized: (1) basal, dark-colored, chert/carbonate laminite; (2) light-colored, thick-bedded, laminated to massive chert; (3) burrowed/laminated chert; and (4) skeletal packstone.

Dark-colored, abiotic, chert/carbonate laminites form the basal Thirtyone deposits in the deepest parts of the basin where they typically overlie mudstones and shales of the Silurian Frame Formation (Fig. 4). These lowermost Thirtyone deposits typically display centimeter-thick interbeds of structureless or finely laminated chert and carbonate mudstone (Fig. 10A). The succession of characteristically nonporous, chert/carbonate laminites reaches thicknesses of nearly 300 ft in the central area of the Thirtyone depocenter, and thins toward the margins.

Thick-bedded, laminated to massive chert, in striking contrast to the dark-colored chert and mudstone interval

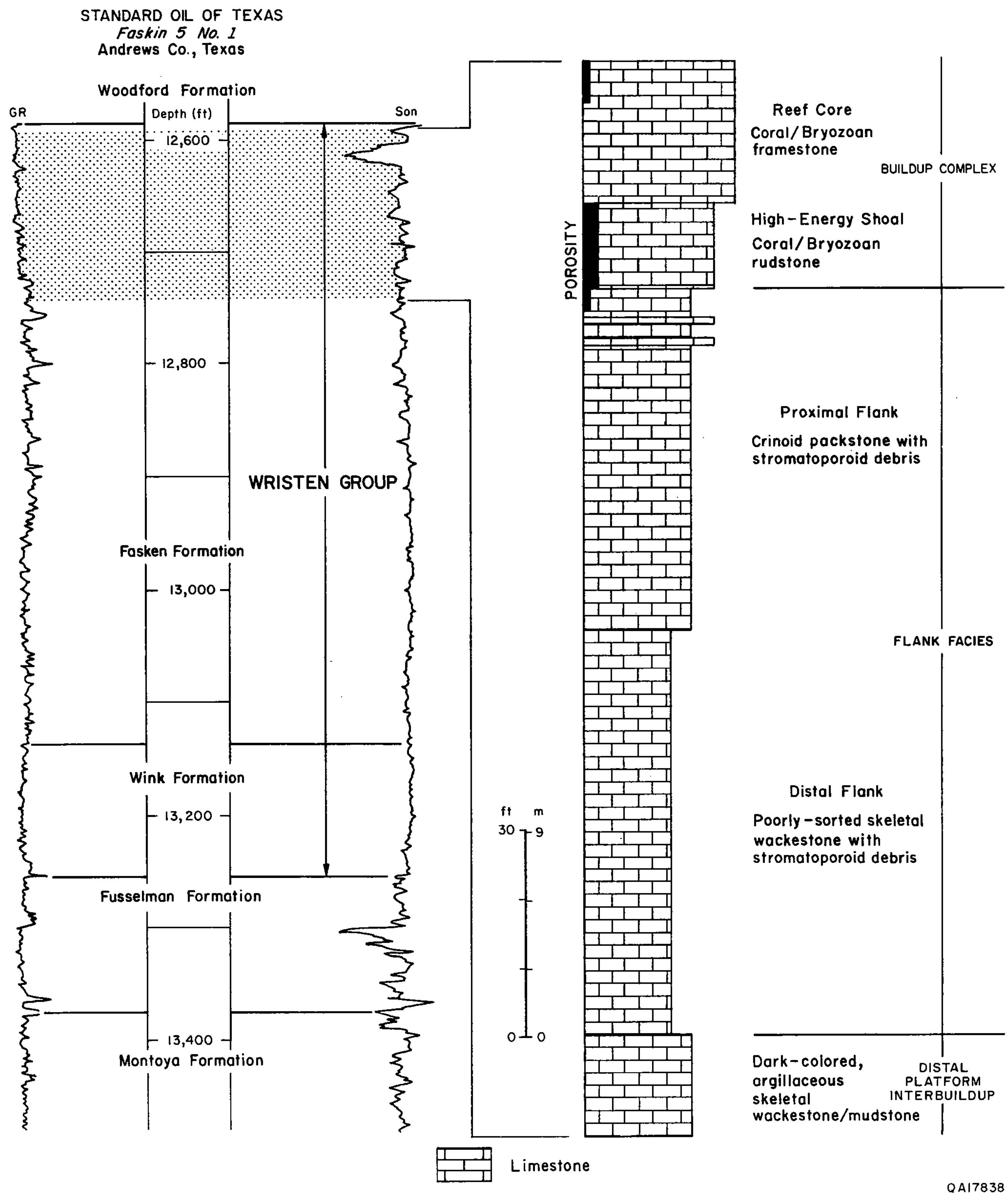

Figure 7. Stratigraphy and facies of the Fasken Formation (Wristen Group) near the Middle Silurian platform margin.

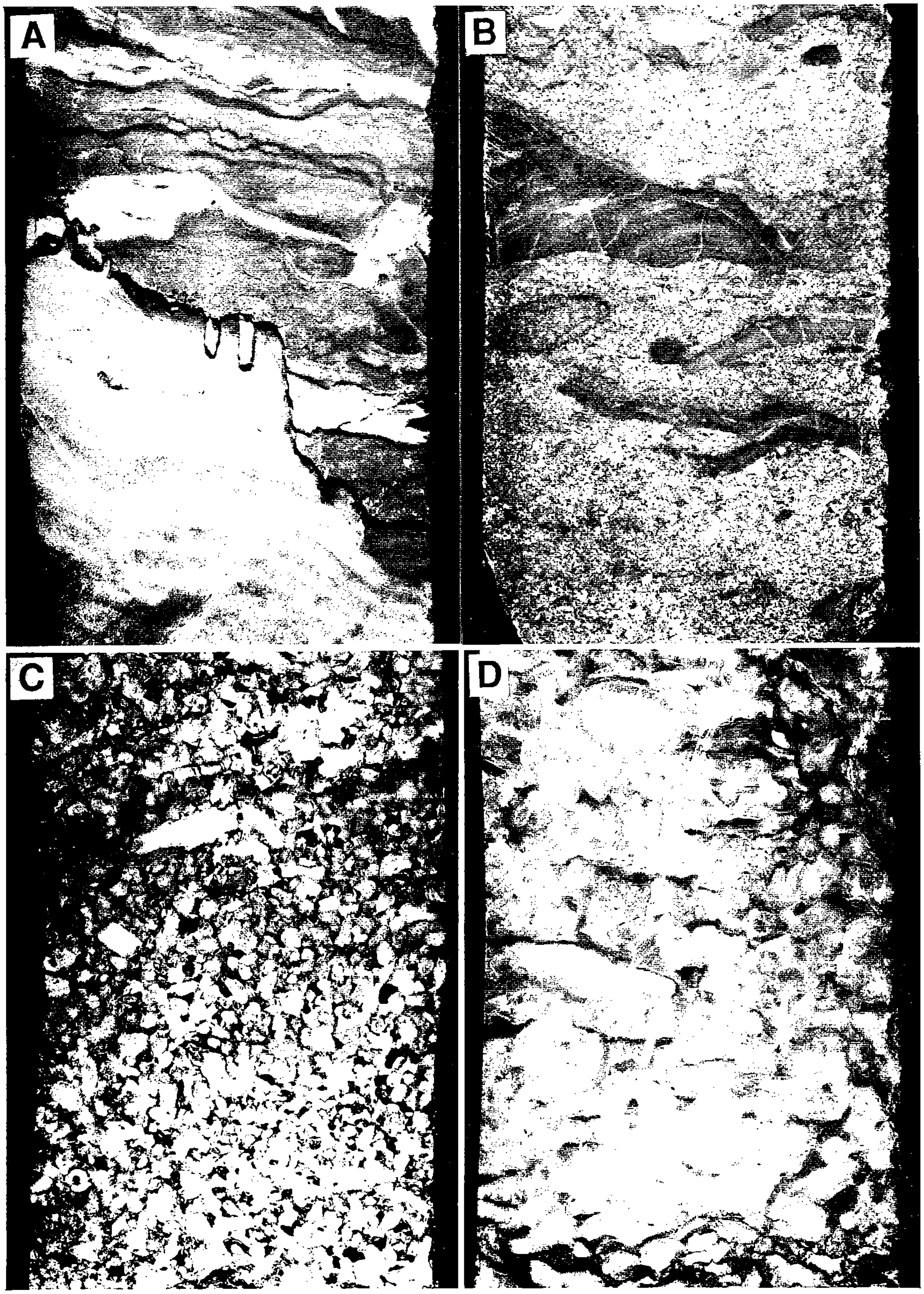

Figure 8. Representative facies in the Fasken Formation. *A*—Stromatoporoid/coral packstone/wackestone containing abundant, poorly sorted, skeletal debris deposited along distal buildup flanks. Note large transported, hemispherical stromatoporoid and smaller, thin, in-situ, laminate stromatoporoids. Depth: 12,632 ft. *B*—Pelmatozoan/stromatoporoid packstone. Moderately well-sorted pelmatozoan debris forms the matrix for large clasts of hemispherical stromatoporoids in this deposit, which formed on the proximal flanks of Wristen buildups. Depth: 12,629 ft. *C*—Coral/bryozoan rudstone containing stick corals and ramose bryozoans that were deposited as a coarse, high-energy lag on buildups. Locally, these rocks exhibit well-developed primary, intergranular porosity. Depth: 12,620 ft. *D*—Coral boundstone composed primarily of stick corals and less-common stromatoporoids. Buildup deposits commonly exhibit numerous sediment-filled geopetals, but generally they are nonporous. Depth: 12,597 ft. All photographs from Standard Oil of Texas, Fasken 5 No. 1, Andrews County, Texas. Slabs are 8 cm wide.

it typically overlies, is light colored, highly porous, and nearly massive in some locations. Dark-colored, organic-rich laminations are common, especially at the tops of beds, which are as much as 3 ft thick (Fig. 10B). Fluid-escape structures and vague, wispy laminations are locally common. Carbonate in these rocks takes the form of small, irregular patches of calcite, some of which can be identified as scattered, corroded fragments of skeletal debris, and isolated dolomite rhombs. Thick-bedded, laminated chert, which has been referred to as "tripolitic chert" by many workers, constitutes the most important reservoir facies in the Thirtyone Formation. At many localities these deposits contain abundant sponge spicules, which are well sorted and preserved both as open and quartz- or chalcedony-cemented molds (Fig. 10C). Successions of thick-bedded, laminated chert reach thicknesses of as much as 150 ft.

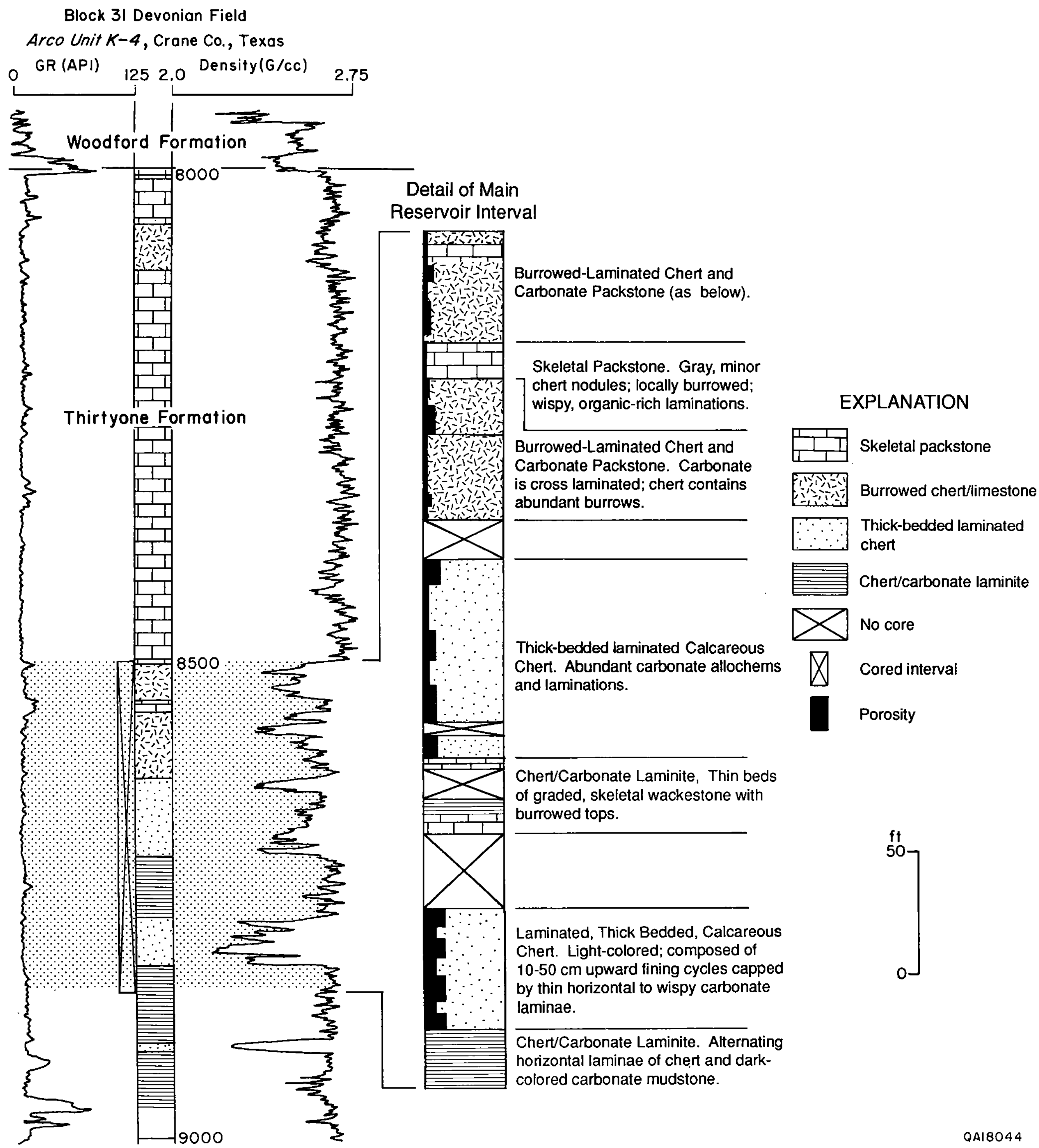

Figure 9. Typical downdip stratigraphic succession in the Thirtyone Formation, Crane County, Texas.

Burrowed/laminated chert includes both fine-laminated (2–3-mm laminations) chert and burrowed chert (Fig. 10D). These two facies are interbedded at several scales. Laminated chert is more common in the lower parts of the Thirtyone, where it is interbedded with the thick-bedded, laminated chert facies. Burrowed chert is increasingly abundant upsection, where it is commonly interbedded with carbonate packstone.

Carbonate packstone (Fig. 10E) dominates the upper part of the Thirtyone throughout all but the extreme south-

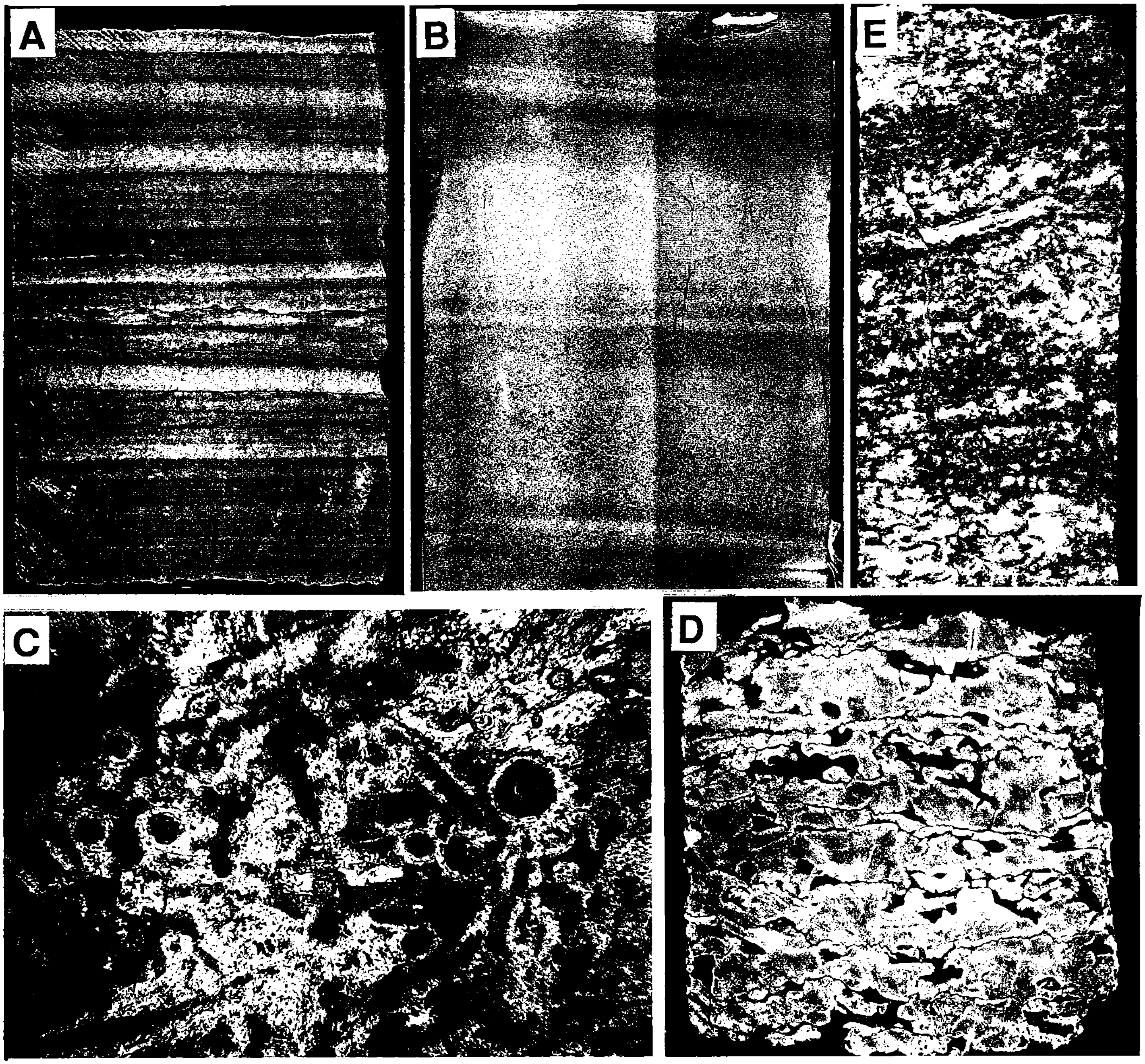

Figure 10. Representative facies from the Thirtyone Formation. *A*—Chert/carbonate laminite facies illustrating alternating layers of carbonate and siliceous mudstone, representing hemipelagic deposition in a quiet-water, below-wave-base setting. Depth: 8,850 ft. *B*—Thick-bedded, laminated chert facies composed of cycles having nearly massive, light-colored chert bases and thin, laminated tops. Some beds show chaotic, disrupted laminations, probably caused by dewatering or soft-sediment deformation. Laminations are organic-rich and may contain minor amounts of carbonate or siliceous mudstone. Depth: 8,800 ft. Photographs A and B from Arco Block 31 Unit No. K-4, Crane County, Texas. *C*—Photomicrograph showing grain-supported, spiculitic chert of the thick-bedded, laminated chert facies, Thirtyone Formation. Porosity (31%) is developed as obvious, large (50–100 μm) spicule molds and smaller (<5 μm) intercrystalline pores. Permeability is 27 md. Field of view is 1 × 2 mm. Depth: 8,155 ft. Unocal, Dollarhide 46-5-D, Andrews County, Texas. *D*—Burrowed chert facies. Dark patches are carbonate mudstone that is the result of burrowing and/or soft-sediment deformation. Arco Block 31 Unit No. K-4, Crane County, Texas. *E*—Coarse-grained skeletal packstone of the Thirtyone Formation in the northern part of its distribution. These deposits contain large, well-preserved bryozoans, pelmatozoans, and brachiopods that are commonly poorly sorted, suggesting largely in-situ deposition. Depth: 10,508 ft. Arco University 9B No. 4, Andrews County, Texas. All core slabs are 8 cm wide.

ern parts of its distribution, where the upper part of the section has been removed by erosion. These deposits are primarily composed of fine-grained, well-sorted, grain-supported, skeletal packstones that contain abundant pelmatozoans and locally common bryozoans and ostracodes. Typically, these deposits are burrowed, although some are laminated and locally display normal grading. Grain size and sorting increases northward.

Thirtyone facies exhibit cyclic-stacking patterns of at least two scales. The lower part of the section in the Thirtyone depocenter is characterized by cyclic alternations between chert/carbonate laminites and laminated, thick-bedded, calcareous chert (Fig. 9). Upsection, analogous cycles are composed of burrowed/laminated chert and thick-bedded chert. Yet farther upsection, cycles are composed of burrowed/laminated chert and skeletal wackestone/packstone. Higher-frequency cycles are apparent within this scale of cyclicity. In the chert/laminate facies, high-frequency cyclicity is reflected by the thin, centimeter-scale alternations of chert and carbonate mudstone (Fig. 9). Laminated, thick-bedded, calcareous chert units are composed of thin, 10–50-cm-thick cycles characterized by grain-dominated bases and laminated, mud-dominated tops. These rocks also display porosity trends that suggest a lower-order (longer-duration) cyclicity may also be present (Fig. 9).

Depositional Setting

Thirtyone chert-dominated successions represent deposition in a deep-water, slope-to-basin setting. This is indicated by the restricted faunal complement, the scarcity of platform-derived carbonate, sedimentary structures, and regional stratal geometries. Upward-fining successions and water-escape structures suggest rapid deposition of grain-dominated (spiculitic) sediments by mass gravity-transport processes, such as those associated with submarine-fan complexes. Much of the carbonate in the Thirtyone accumulated similarly by downslope transport of platform-derived sediment. Vertical and updip increases in allochem grain size, and more-normal-marine faunal elements, document progradation and infilling of basin/slope topography. More poorly sorted skeletal carbonate facies at the top of the Thirtyone in the basin and updip represent shallow-water, fully aggraded carbonate-platform deposition, and document complete infilling of depositional relief.

Reservoir Development

Thirtyone reservoirs are developed in both chert and carbonate successions, although the former constitute by far the greatest hydrocarbon resource (Fig. 2). Chert-reservoir development is a function of both diagenesis and original depositional facies. Reservoir distribution reflects original basin geometry; reservoirs are essentially confined to a relatively narrow axis through the basin depocenter. Porosity in this trend is commonly spicule-moldic (Fig. 10C) or microcrystalline (Ruppel and Hovorka, 1990; Saller and others, 1990a, b; Ruppel and Holtz, in press). Both pore types appear to be a response to differential silica diagenesis soon after deposition, but porosity development is commonly greatest in higher-energy, grain-dominated chert facies.

Carbonate reservoirs in the Thirtyone Formation are developed in grain-dominated successions that have undergone meteoric diagenesis associated with exposure events. Most of this diagenesis is associated with leaching and dolomitization at the top of the Thirtyone section. These reservoirs are, for the most part, restricted to the updip margins of the Thirtyone subcrop overlying the Silurian platform margin. Porosity is developed as interparticle and intercrystalline pores in leached and dolomitized packstones. In contrast to the Fusselman and Fasken shallow-water successions, there is no evidence of penecontemporaneous sea-level fall and exposure resulting in porosity evolution in the Thirtyone.

DEPOSITIONAL HISTORY

Regional facies relationships suggest the following depositional model for West Texas and New Mexico during the Late Ordovician to Early Devonian. Fusselman Formation carbonate deposits accumulated on an extensive shallow-water carbonate platform that originated during the Late Ordovician (Fig. 11A). The similarity of depositional facies across the region during this time indicates that conditions were relatively uniform over great distances. Equivalent rocks in Oklahoma, for example, are lithologically similar (Amsden, 1980; Amsden and Barrick, 1986). Depositional histories of Fusselman and worldwide stratigraphic equivalents demonstrate that the accumulation of these rocks was punctuated by rise-and-fall cycles of absolute sea level of ~2 m.y. duration (McKerrow, 1979; Johnson, 1987). In the Fusselman, these eustatic events are reflected in unconformities and associated diagenetic alteration related to episodes of exposure.

The Fusselman shallow-water platform was drowned by sea-level rise, probably associated with tectonic downwarping of the southern parts of the region during the Wenlockian (Fig. 11B). Regional stratigraphic relationships indicate that this event was widespread along the southern margin of the North American craton (Becker and Droste, 1978; Droste and Shaver, 1987; Amsden and Barrick, 1988).

Drowning of the Fusselman platform was followed by establishment of deep-water-ramp/outer-platform deposition in southern West Texas (Wink and Frame Formations) and by continued shallow-water deposition to the north (Fasken Formation) (Fig. 11C). Deeper-water deposits document a classic drowning succession, with outer-ramp nodular wackestones and mudstones (Wink Formation) being overlain by more distal, deeper-water, carbonate mudstones and shales (Frame Formation). The increasing abundance of skeletal debris, in many cases in the form of carbonate turbidites, upward and northward in the Frame attests to the aggradation of the Fasken platform to the north, and downslope transport of platform material southward into the basin.

The Middle to Late Silurian was punctuated on the Wristen platform by episodic rise and fall of relative sea level, similar to that which affected Fusselman deposition. This cyclicity is reflected in both repetitive, upward-shallowing facies-stacking patterns, recording sea-level rise, and selectively leached cycle tops, associated with sea-level fall.

During the Late Silurian or Early Devonian, a major rise in relative sea level took place in West Texas and New

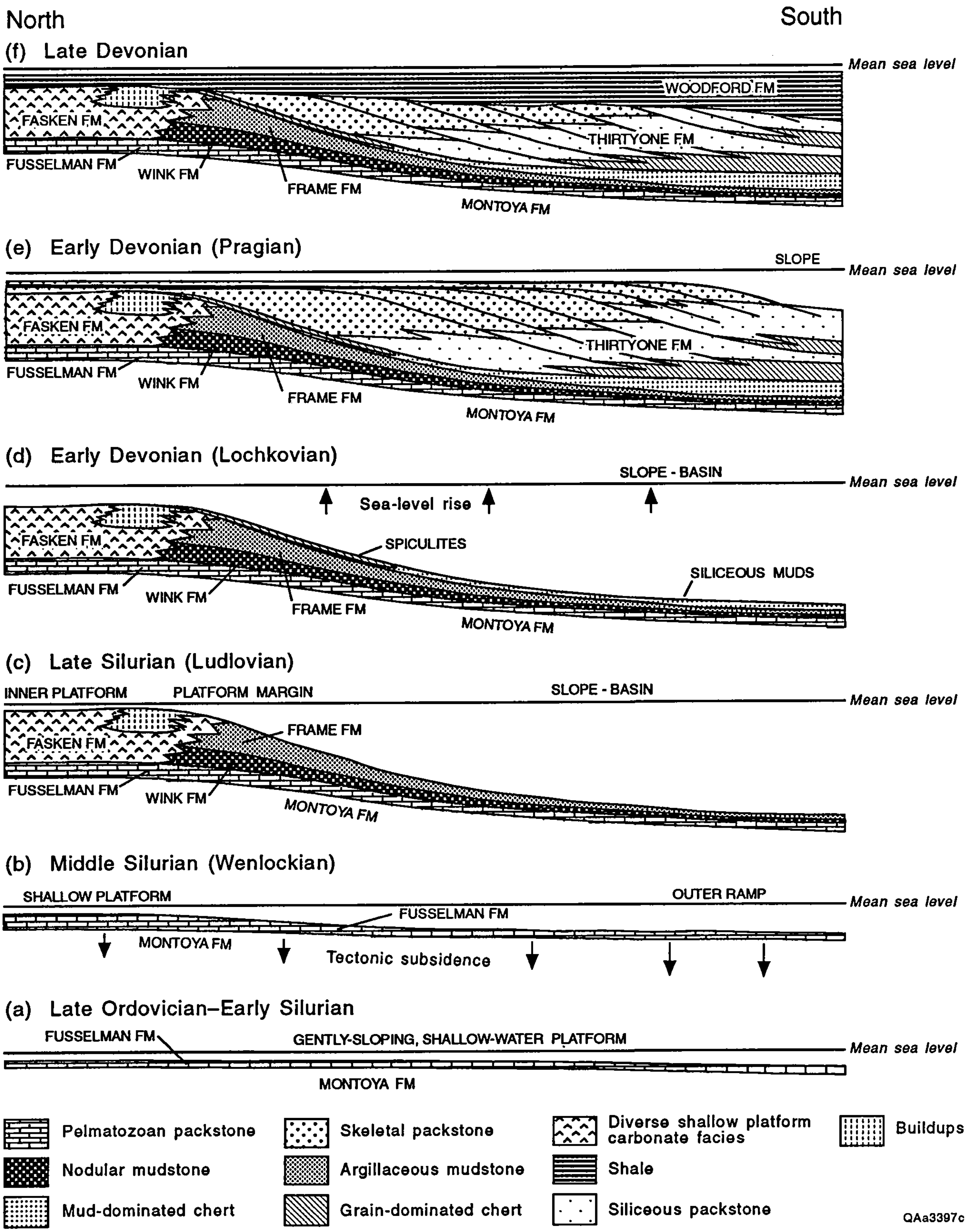

Figure 11. Depositional history of Late Ordovician–Devonian carbonate and chert deposition in West Texas and New Mexico.

Mexico. Relative rise is evidenced by the distinctly deeper-water character of basal Thirtyone Formation deposits, compared with immediately underlying Wristen Group rocks. Evidence for global sea-level rise at this time is equivocal (cf., Vail and others, 1977; McKerrow, 1979). Deepening at the Silurian–Devonian boundary in West Texas may in part be related to a second pulse of foreland deformation associated with the continued convergence of the North American and South American/African plates (Walper, 1977).

Early Devonian sedimentation (Fig. 11D) was marked initially by pelagic-mud accumulation throughout the Thirtyone depocenter. The distribution of these early, mud-dominated siliceous deposits, which represent transgressive deposits formed during sea-level rise, was controlled by basin geometry inherited from Silurian (Wristen) deposition. Along the proximal, northern extent of the deep-water axis of the basin, accumulation of siliceous muds rapidly gave way to submarine-fan deposition of grain-rich, spiculitic sands (Fig. 11D). During accumulation of these high-energy chert deposits, mud-dominated sediments continued to accumulate on the lower-energy flanks of the northern axis and farther south in the deeper, more distal parts of the basin. With decreasing rates of sea-level rise, carbonate production on the shelf resulted in rapid aggradation and subsequent basinward progradation.

Coincident with highstand progradation of the carbonate platform, high-energy, grain-dominated cherts began to be deposited in submarine fans in front of the advancing carbonate wedge (Fig. 11E). Initially, geometries and textures of siliceous sediments were related to submarine-fan depositional processes, with high-energy, grain-dominated sediments concentrated along fan/channel axes, and the more mud-dominated deposits accumulating in interchannel areas. With continued progradation of the carbonate platform, fan deposition was interrupted by episodic influx of carbonate debris derived from the north. Continued platform progradation produced progressive shallowing of the basin and terminated chert accumulation.

As was the case during the Silurian, Devonian Thirtyone deposition was markedly cyclic. Submarine-fan chert deposits exhibit cyclic-facies stacking patterns that may be due to shifting axes of deposition. These cycles are associated with different energy regimes that are reflected in changes in mud content. Cyclic porosity development in these rocks (Fig. 9) appears to reflect similar variations in current energy and resultant facies. Although these cycles may have been the result of normal submarine-fan deposition, it is likely that the ultimate cause of these depositional cycles may have been high-frequency rise and fall of sea level.

Truncation of the Thirtyone Formation, as well as that of underlying Wristen Group and Fusselman Formation strata, probably occurred during the Middle Devonian. Regional studies suggest that uplift and truncation of much of the craton of the southwestern United States took place at this time (Ham and Wilson, 1967). Work by Johnson and others (1985) also indicates a global sea-level lowstand during the late Pragian. Much of the carbonate-platform equivalents of the Thirtyone Formation were removed by erosion at that time, although several erosional remnants still survive within the thick, predominantly Silurian, Wristen section of carbonate rocks present in the northern part of the West Texas (Wilson and Majewski, 1960; J. E. Barrick, personal communication, 1992). Leaching of uppermost Thirtyone Formation rocks at several localities indicates that significant diagenetic alteration took place during the Middle Devonian uplift. Much of the Wristen platform-carbonate succession (Fasken Formation) and the Fusselman underwent additional, overprinting meteoric diagenesis at this time as well.

West Texas and New Mexico, along with most of the North American craton, was inundated in the Late Devonian (Frasnian) by an anoxic, shallow-water sea in which substantial thicknesses of organic muds and silts accumulated. In the Permian basin area (Fig. 11F), as much as 650 ft of Woodford Formation shales now overlie the Silurian and Devonian carbonate section (Ellison, 1950; Comer, 1991). The Woodford was locally truncated during Pennsylvanian–Permian regional tectonism of the area associated with collision and the assembly of Pangea.

SUMMARY AND CONCLUSIONS

Collectively, Silurian and Devonian carbonate strata in the Permian basin exhibit a high degree of facies complexity that reflects a varied depositional and diagenetic history. There are, however, distinctive patterns in the development of these Hunton-equivalent facies that reflect major stages of basin evolution during the deposition of this succession. Because Upper Ordovician–Lower Silurian Fusselman rocks were deposited on a stable carbonate platform of regional extent, facies display widespread continuity. In contrast, Middle–Upper Silurian Wristen Group rocks, which accumulated during platform drowning, and Upper Devonian Thirtyone rocks, which were deposited during basin filling, display considerably greater facies diversity. As a result, reservoir development in Fusselman rocks is more significantly a response to diagenesis, whereas Wristen (Fasken Formation) and Thirtyone reservoirs reflect both diagenetic and depositional controls.

Common to all of these deposits is the unmistakable imprint of episodic rise and fall of relative sea level. In the Fusselman and the Fasken, diagenesis and porosity development are associated with both moderate- and long-duration exposure events. Shallow-water Thirtyone carbonate rocks owe their reservoir development to the same long-term exposure event. In Fasken rocks, sea-level rise and fall exerted a strong control over facies geometries as well as diagenesis. Despite their much different depositional setting and lithological character, Thirtyone chert deposits also display evidence suggestive of sea-level control of facies development. Diagenesis in these rocks, however, probably is largely decoupled from sea-level fluctuations.

An understanding of the controls and processes responsible for development of reservoir facies is crucial for efficient exploration and exploitation of remaining hydrocarbon resources in the Upper Ordovician–Lower Devonian strata of the Permian basin. Models developed for the West Texas succession may aid in understanding the evolution of the Hunton Group and may also suggest new approaches to exploiting the remaining resource in these rocks.

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Stratigraphy and Petroleum Potential of the Hunton Group (Silurian–Devonian) in the Forest City Basin Area of Kansas and Nebraska

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ABSTRACT.—The Hunton Group carbonates in Kansas and Nebraska range in age from Middle Silurian to Late Devonian. Devonian dolomite is productive on anticlines within the Forest City basin. On the flanks of both the Nemaha uplift and the Sedgwick basin, production is from combined stratigraphic/structural traps with seals of either Chattanooga Shale or basal Pennsylvanian conglomerate that overlie the eroded Hunton. In most Hunton reservoirs, porosity has been enhanced by vugs and vertical fractures. Oil gravity ranges from 29 to 36° API, and nearly all fields have a strong water drive.

INTRODUCTION

The Hunton Group, as used in the western Midcontinent, is a sequence of carbonate rocks bounded above and below by shale. In the southern Midcontinent, the overlying shale is the Woodford (Upper Devonian) and the underlying shale is the Sylvan (Upper Ordovician). In the central Midcontinent, the overlying shale has been termed Chattanooga (Upper Devonian) and the underlying shale is the Maquoketa (Upper Ordovician). Thus the lithologic sequence is similar across the area. However, the age of the subdivisions of the intervening carbonate varies.

In Oklahoma (Amsden and Barrick, 1988), the Hunton carbonate ranges in age from Late Ordovician, through Silurian, and into the Early Devonian , with minor time hiatus (Fig. 1). Middle Devonian rocks are absent. The Woodford Shale of Late Devonian age unconformably overlaps the Lower Devonian, and in some areas contains a basal sandy zone (the Misener sand).

In Kansas and Nebraska, the equivalent Hunton carbonate package consists of Middle Silurian dolomite unconformably overlain by Middle Devonian dolomite and limestone. In Kansas, the Upper Devonian consists of a shale facies (Chattanooga) that grades northward towards Nebraska into carbonate (Lime Creek) overlain by shale (Sheffield/Maple Mill).

STRATIGRAPHY

Figure 2 is a north–south structural cross section across eastern Nebraska and Kansas. The depositional center for much of the Silurian and Devonian was the North Kansas basin, as emphasized by the thickness of the major units (Fig. 3). The Hunton Group is represented by carbonate of both Silurian and Devonian ages.

Silurian rocks were deposited widely across North America. Regional uplift prior to Devonian deposition allowed removal of most of these carbonates across the central Midcontinent, except where preserved within areas such as the North Kansas basin (Fig. 3D). Deposition, however, was nearly continuous into the Early Devonian in the southern Midcontinent. It wasn't until the Middle Devonian that the seas advanced westward and southward into the central Midcontinent, depositing carbonates with thin shale lenses (Fig. 3B). During the Late Devonian, shale deposition was widespread; dark-gray shale of the southern Midcontinent merged northward into green-gray shale, and finally into a carbonate facies (Fig. 3A). Much of the Midcontinent was emergent late in Devonian time, prior to the major transgression of Early Mississippian seas.

Thickness of the various units was strongly affected by uplift and erosion during Pennsylvanian time, particularly

Carlson, M. P.; and Newell, K. D., 1993, Stratigraphy and petroleum potential of the Hunton group (Silurian–Devonian) in the Forest City basin area, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 107–115.

SERIES	STAGES	OKLAHOMA	FOREST CITY BASIN	TERMINOLOGY
Upper Devonian	Famennian	Woodford	Chattanooga/Sheffield	
	Frasnian		Lime Creek	
		//////////////////////////		
Middle Devonian	Givetian	//////////////////////////	Cedar Valley	
		//////////////////////////		
	Eifelian	//////////////////////////	Wapsipinicon	
		//////////////////////////		
	Emsian	Sallisaw	//////////////////////////	HUNTON
Lower Devonian	Siegenian	Frisco	//////////////////////////	
	Gedinnian	Haragan-Bois d'Arc	//////////////////////////	
	Pridolian	Henryhouse	//////////////////////////	
Upper Silurian	Ludovian		//////////////////////////	
	Wenlockian	Clarita	Grower	
Lower Silurian	Llandoverian	Cochrane	Hopkinton-Scotch Grv.	
Upper Ordovician		Keel		
	Ashgillian	Sylvan	Maquoketa	

Figure 1. General correlation chart showing terminology and age of the Hunton Group in Oklahoma, Kansas, and Nebraska.

over the trend of the Nemaha uplift (Fig. 3). All of the lower and middle Paleozoic rocks were removed over portions of this structural high. The Forest City basin is the eastern half of the depositional basin (North Kansas basin) which was present during the middle Paleozoic. The thickness map of the Silurian (Fig. 3D) depicts the thickening into the North Kansas basin, and locates the erosional zero edges (zero thickness) created prior to Devonian deposition. The zero edges over the Nemaha uplift were created in Pennsylvanian time when this feature separated the North Kansas basin into the Salina basin to the west and the Forest City basin. A similar history and resulting patterns are obvious on the Devonian thickness maps. Figure 3C illustrates the current relationship of tectonic features in the Forest City basin area.

OCCURRENCE AND CHARACTER OF PETROLEUM IN THE HUNTON GROUP

Characteristics of Crude Oil

Studies of the geochemistry of the crude oils in the Forest City basin have focused on Ordovician source rocks (Decorah Formation and Simpson Group). This focus is due in part to the requirements for crude-oil generation—organic content and relationship to an "oil window." Newell and others (1987b) determined that crude oils in the older Paleozoic of the Forest City basin area have nearly identical chemical signatures, as revealed by gas chromatography. They also point out that these oils are very similar to bitumen extracted from shale in the Middle Ordovician Simpson Group. Chemical characteristics of Ordovician oils (Hatch and others, 1985) include a relatively high abundance of n-alkanes with carbon numbers less than 20, a strong predominance of odd-numbered n-alkanes between C_{10} and C_{20}, and relatively small amounts of branched and cyclic alkanes.

Possible pathways for migration of Ordovician oil into the Hunton are the fault and fracture zones associated with many of the structural traps, and the various unconformities that intersect the Hunton surface. Other potential source rocks are the Maquoketa Shale, underlying the Hunton, and the overlying Chattanooga Shale. Data obtained by Lambert (1992) indicate that the Chattanooga is sufficiently rich in organic matter to be an effective source rock. However, analysis of temperature data and vitrinite reflectance indicate that this shale is only marginally mature. The possibility does exist that regional thermal events or local thermal anomalies could have created temporary "oil windows" for shales younger than the Simpson.

The Hunton carbonate has produced oil in three areas: (1) over small anticlines within the Forest City basin, (2) stratigraphic traps under the Pennsylvanian unconformity, on or near the Nemaha uplift, and (3) in stratigraphic traps underlying the Chattanooga unconformity on the north flank of the Sedgwick basin (Fig. 3C). Scattered small fields have also been found in the Misener sandstone at the base of the Chattanooga Shale. Table 1 summarizes the production data for fields producing from the Hunton in Kansas and Nebraska.

Production in the Forest City Basin

The post-Mississippian Ouachita orogeny created the Salina and Forest City basins (Fig. 3C) by uplift along the Nemaha structural zone; this bisected the Paleozoic depocenter, the North Kansas basin. Although similar rocks are present on both sides of the Nemaha uplift, no Hunton production has been found in the Salina basin to the west. The Forest City basin fields are located on small anticlines that were created by mid-Pennsylvanian reactivation of basement features during the Ouachita orogeny (Carlson, 1989a).

The Hunton discovery at the Falls City field (Richardson County, Nebraska) in 1939 was the initial Paleozoic production in the Forest City basin (Carlson, 1989b). The series of Hunton fields in the basin occur on northwest-trending anticlines with 100–150 ft of closure. The very strong water drive caused many of the early wells to flow. There is no defined

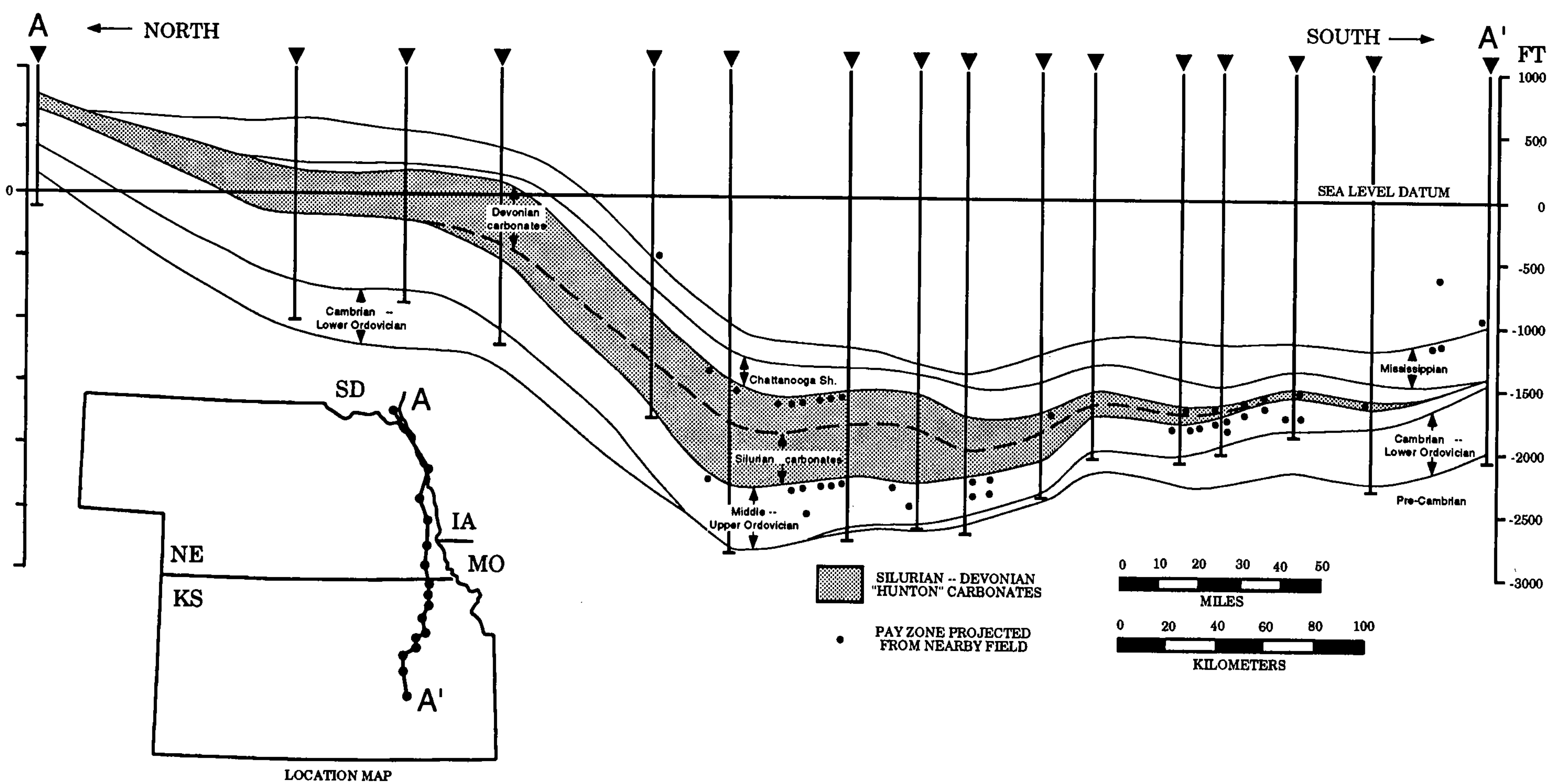

Figure 2. North–south structural cross section of the lower and middle Paleozoic rocks in eastern Nebraska and Kansas.

Table 1. — Characteristics of Selected Oil Fields Producing from the Hunton in Kansas and Nebraska

FIELDS WITH SOLELY OR MOSTLY "HUNTON" OIL PRODUCTION

FIELD NAME	DISCOVERY DATE[1]	INITIAL PRODUCTION (bopd)	API GRAVITY	AREA (acres)	PRODUCING WELLS (1991)	1991 PRODUCTION (bbls.)	CUM. PRODUCTION[2] (bbls.)	OTHER PAYS	DEPTH (ft)
Nemaha uplift									
Alta Vista	(1977)	5	29	40	0	0	NA	Perm. (gas)	1,928
Yaege	1959	36	33	1,600	40	55,911	3,247,263	Penn. cg	1,682
Forest City basin									
Barada	1941	680	29	1,500	13	12,408	3,202,634	----	2,450
Bosch	1973	25	40	280	7	9,228	414,677	----	2,748
Bushong	1950	257	29	40	0	0	20,640	----	2,952
Dawson	1941	55	24	680	4	1,700	515,400	Viola, Simpson	2,200
Falls City	1939	700	30	1,080	10	14,002	5,659,909	----	2,250
Kizler North	1972	48	23	200	0	0	34,364	---	2,964
Leach	1963	26	32	240	6	5,142	516,783	Viola	2,713
Livengood	1944	85	26	120	15	2,245	203,369	Viola, Simpson	2,580
Sabetha	1950	319	25	280	7	4,412	213,689	Viola	2,898
Shubert	1940	300	23	400	4	2,081	186,125	----	2,430
Strahm	1948	54	22	80	2	3,437	382,112	Viola	2,879
Sedgwick basin									
Buzzi	1958	148	42	100	0	0	443,718	----	3,436
Doyle Creek	1958	65	NA	80	0	0	46,907	----	2,891
Gingrass	1957	96	40	40	1	2,752	236,411	Burgess	3,226
Goessel	1958	12	33	120	3	7,596	1,081,723	----	3,407
Goessel Southeast	1980	40	NA	40	1	965	34,567	----	3,351
Goessel-Branch	1981	30	39	40	1	2,345	101,420	----	3,344
Graber	1934	870	41	1,200	30	15,285	16,705,760	Mssp, Misener, Simpson	3,274
Graber East	1961	25	38	40	0	0	1,232	----	3,404
Haley	1967	156	31	40	1	1,151	117,492	K.C.	3,426
Harmac	1955	33	40	120	3	3,010	935,327	----	3,521
Harmac East	1956	10	NA	40	0	0	NA	Mssp	3,522
Hollow-Nikkel	1931	520	NA	1,890	63	64,680	28,153,957	L.-K.C., Mssp, Misener, Simpson, Arbuckle	3,507
Krehbiel	1973	14	40	40	1	954	45,300	----	3,350
Lalouette	1957	60	NA	40	1	106	36,720	----	2,231
Marion Townsite	1985	10	NA	40	0	0	791	----	2,512
Sperling	1934	1,624	NA	200	5	2,839	1,192,665	Mssp, Simpson	3,279
Stroud North	1982	46	NA	40	1	922	25,863	Simpson	3,734
Unger	1955	122	NA	870	29	29,232	8,351,036	Mssp, Misener	2,809
Wenger	1947	50	37	40	1	966	1,079,413	----	2,770
Whitewater Northwest	1970	25	36	200	5	3,433	152,149	Mssp	2,675
Winsinger	1968	100	42	700	14	16,337	1,049,562	Mssp, Maquoketa	3,354
Winsinger West	1975	15	41	120	3	1,784	115,338	Mssp	2,959

FIELDS IN KANSAS WITH MINOR "HUNTON" OIL PRODUCTION

FIELD NAME	DISCOVERY DATE[1]	INITIAL PRODUCTION (bopd)	API GRAVITY	AREA (acres)	PRODUCING WELLS (1991)	1991 PRODUCTION (bbls.)	CUM. PRODUCTION[2] (bbls.)	OTHER PAYS	DEPTH (ft)
Forest City basin									
Davis Ranch	1949	NA	29	400	10	59,313	8,392,173	K.C., Viola	2,929
Wilmington	1959	NA	23	320	8	17,954	480,764	Viola, Simpson	2,860
Sedgwick basin									
Bentley	(1982)	15	39	80	0	0	23,362	K.C.	3,842
Butwick	1949	NA	NA	130	4	2,943	133,694	Mssp, Misener, Simpson	3,166
Burrton	1931	NA	42	9,450	315	400,501	74,616,047	L.-K.C., Mssp, Simpson, Arbuckle	3,583
Covert-Sellers	1920	NA	38	1,170	39	24,481	2,077,508	K.C., Viola	2,397
Cross	1929	NA	NA		0	0	175,426	K.C.	3,574
Elbing	1918	NA	35	1,820	89	92,216	11,497,460	K.C., Mssp, Viola	2,428
Fairplay	1957	91	NA	120	3	826	592,010	Mssp, Viola	2,458
Fairview	(1979)	95	NA	200	5	2,887	1,041,656	K.C., Burgess, Mssp, Viola, Simpson	3,353
Florence	1920	NA	32	280	7	11,909	593,966	Misener, Viola	2,300
Friendship	1941	NA	NA	200	5	7,098	251,348	K.C., Mssp, Viola	NA
Goodrich	1928	NA	39	1,080	27	14,636	6,562,585	L.-K.C., Mssp	3,335
Greenwich	1929	NA	NA	840	21	33,572	14,193,322	Mssp, Viola, Simpson, Arbuckle	3,210
Halstead West	1960	108	26	440	11	11,311	1,315,694	K.C., Mssp, Simpson	3,160
Peabody	1919	NA	NA	1,450	29	35,272	2,180,786	L.-K.C., Mssp, Viola	2,560
Powerline	1984	NA	NA	40	1	1,660	15,833	Mssp, Simpson	3,205
Valley Center	1928	NA	NA	1,400	35	92,289	24,375,189	K.C., Mssp, Misener, Viola, Simpson	3,360

FOOTNOTES

[1]Date in parentheses indicates date of Hunton pay discovery

[2]Records on production are kept only after January 1, 1944

oil/water contact within the Hunton, and water production is high in proportion to oil production.

The occurrence of oil in the Hunton (Devonian) rocks of the Forest City basin was described by Carlson (1988). The productive interval has been the uppermost carbonates, probably of Lime Creek age (Fig. 1). The reservoirs are vuggy limestone and dolomite, quite often with vertical fractures, and pay thicknesses of up to 36 ft. Oil is produced from depths of 2,240–2,870 ft; it has a strong water drive, but no associated gas.

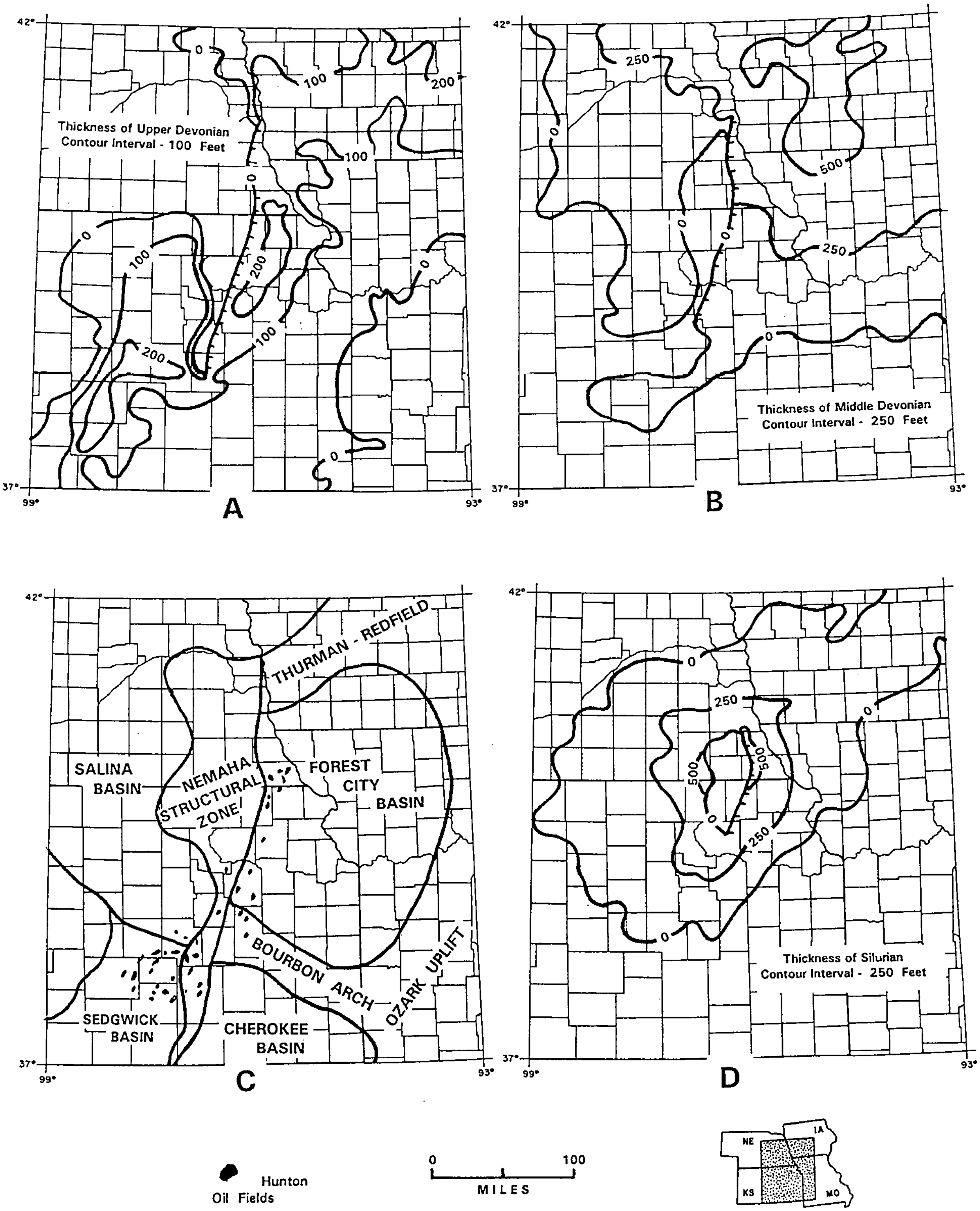

Figure 3. Maps covering parts of Kansas, Nebraska, Iowa, and Missouri. *A*—Distribution and thickness of Upper Devonian rocks. *B*—Distribution and thickness of Middle Devonian rocks. *C*—Major tectonic features and general location of oil fields producing from the Hunton. *D*—Distribution and thickness of Silurian rocks.

Production Related to the Nemaha Uplift

Several oil fields are located on the crest and along the flanks of the Nemaha uplift. Production is from stratigraphic traps in various subdivisions of the Hunton. These reservoirs were created by uplift, erosion, and then burial under the basal Pennsylvanian unconformity.

In the Yaege field in Riley County, Kansas, both the truncated Hunton and overlying Pennsylvanian conglomerate produce oil of ~33° API gravity with a strong water drive (Goebel, 1960). The Hunton reservoir is a dense to finely crystalline dolomite with vugular porosity and an estimated average porosity of 18%.

Both the Fairplay (Marion County, Kansas) and Gingrass (Harvey County, Kansas) fields produce from a Hunton reservoir overlain by Chattanooga Shale. The fields are structural traps related to the Nemaha uplift (Stubbs and Wright, 1960; Johnson, 1960). The Hunton reservoir in the Fairplay field is a fine- to medium-crystalline dolomite; one log analysis showing an average porosity of 8%. In the Gingrass field the Hunton contains an updip facies change from dolomite to limestone. In both fields, there is a strong water drive and water production continually increases as the reservoir is depleted.

Production on the Northern Flank of the Sedgwick Basin

Hunton production has been found on the north flank of the Sedgwick basin in truncated Silurian and Devonian carbonates (Newell and others, 1987a). A west–east trend of stratigraphic traps was created in Late Devonian time due to erosion and removal of the Hunton in the "McPherson Valley," followed by deposition of the Chattanooga Shale (Lee, 1956).

The Unger field in Marion County, Kansas, is characteristic of this producing trend: the Hunton is 0–50 ft thick (Brown, 1960); and the underlying seal is formed by the Maquoketa Shale, whereas the upper seal is the Chattanooga Shale. The lithology varies from dense, fractured dolomite, with primary porosity averaging ~12% to finely crystalline vuggy dolomite with primary porosity averaging ~19%. There is a very strong water drive with continual encroachment of the water.

Core Representing Selective Rock Characteristics in the Hunton

Very little core has been preserved, even in consideration of the small amount of coring done within the Hunton. Much of the early drilling activity was by cable-tool and open-hole completion. Several core samples are selected to illustrate reservoir-rock type and lithologic relationships. These cores are illustrated on Figure 4 as follows:

Cities Service No. 1-A Steele-Griffee, NW¼NW¼SE¼ sec. 25, T. 5 S., R. 12 E., Nemaha County, Kansas, core diameter 3.5 in. *Upper left photograph*: depth 2,536.5 ft; base of dark-gray Chattanooga Shale separated from tan-gray dolomitic limestone by ~1 in. of pyritic siltstone. *Upper right*: depth 2,542 ft; large vugs with staining in fractured dolomitic limestone.

Shaffer No. 1 Sandrock, NE¼NW¼NW¼ sec. 20, T. 1 N., R. 16 E., Richardson County, Nebraska, core diameter 2.75 in. *Middle left*: depth 2,253 ft; tan-gray, dense dolomitic limestone. *Middle right*: depth 2,266 ft; fine-crystalline, vuggy dolomite with heavy oil stain.

Skelly No. 3 Davisson, SE¼NW¼NE¼ sec. 20, T. 1 N., R 16 E., Richardson County, Nebraska, core diameter 2.5 in. *Lower left*: depth 2,280 ft; fine-crystalline, laminated dolomite with heavy oil stain.

Skelly No. 2 Steinbrink, W½E½NE¼NW¼ sec. 36, T. 3 N., R 16 E., Richardson County, Nebraska, core diameter 3 in. *Lower right*: depth 2,447 ft, gray, dense dolomite with near-vertical fractures and staining on fracture faces.

SUMMARY

The oil production from the Hunton carbonates in the Forest City basin area of Kansas and Nebraska occurs in Devonian-age rocks. Only where truncation has occurred, either beneath the pre-Chattanooga or pre-Pennsylvanian unconformities, has production been discovered in the remnants of Silurian-age rocks. Even in the traps which are primarily structural, there is often a stratigraphic component created by a facies change from dolomite to limestone.

The dolomite reservoir rock of the Hunton varies from light-gray and dense to tan and sucrosic crystalline—the variations resulting from both original fabric and recrystallization. Porosity is significantly increased by the presence of pinpoint to thumb-sized vugs. In many fields, the reservoir yield has been greatly enhanced by fracturing, primarily vertical. The Hunton fields are characterized by strong water drive, leading to high water-to-oil ratios.

Exploration should lead to additional discoveries in structural traps within the Forest City basin in both Kansas and Nebraska. Few wells have explored the potential of the Silurian dolomite. The north flank of the Sedgwick basin has been extensively drilled, and the potential exists for an extension of the "McPherson valley" stratigraphic relationships into the Forest City basin. The subcrop of both Silurian and Devonian rocks under the pre-Pennsylvanian unconformity should be further explored along the Nemaha uplift in both Kansas and Nebraska.

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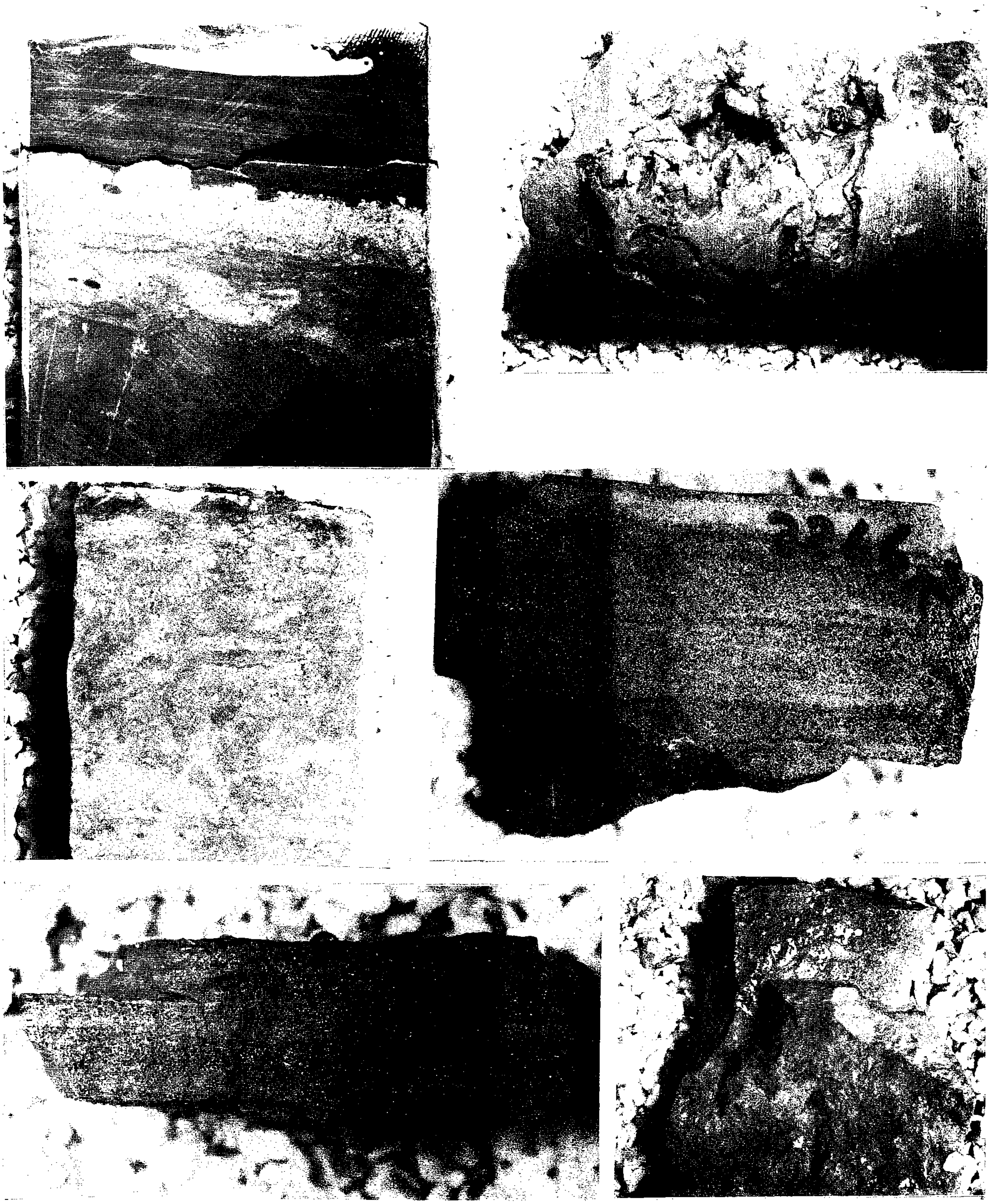

Figure 4. Representative cores of Hunton carbonates (see text for discussion).

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Misener Sandstone: Distribution and Relationship to Late/Post-Hunton Unconformities, Northern Shelf, Anadarko Basin

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INTRODUCTION

The Misener sandstone of the Midcontinent has been the focus of many investigations since its stratigraphic ambiguity and economic importance were first realized in the early 1900s. Since that time, geologic study of the formation has ranged from purely stratigraphic, paleontologic, mineralogic, and diagenetic, to paleogeomorphic and depositional, with varied interpretations of environment (Table 1).

MASERA Corp. conducted a detailed petrostratigraphic study of the Misener in north-central Oklahoma (Fig. 1). The study included examination of numerous cores, samples, and outcrops in eastern Oklahoma and northern Arkansas, and correlation of petrographic and diagenetic analyses to well-log signatures for facies recognition and sequence-stratigraphy analysis. A depositional model of the Misener was constructed and utilized in regional/local correlation and mapping.

Understanding the Misener's relationship to post-Hunton Group (including the pre-Frisco Formation unconformity) and pre-Woodford/Misener sequence unconformities, and resultant paleogeomorphology, is a critical key in determining the depositional environment, overall geological setting, stratigraphy, petrography, source, and distribution/geometry of the Misener reservoirs.

GENERAL GEOLOGIC SETTING AND STRUCTURE

The Misener sandstone is Middle to Late Devonian in age and closely associated with the Hunton Group carbonates along the northern shelf of the Anadarko basin. It is associated also with one of the most prominent unconformities and transgressive sequences developed in the Midcontinent (Fig. 2).

In terms of overall setting for the Misener, Oklahoma during the Devonian was characterized by three major tectono-sedimentologic sequences: (1) the aulacogen of southern Oklahoma, which was in a subsidence stage (as opposed to an earlier rifting and a later deformation stage), (2) a continental marginal basin in the area of the Ouachita system, and (3) the craton–shelf in the areas north of the previously noted regimes (Fig. 1).

The area of detailed study is situated along and west of the Nemaha uplift (Fig. 1) where strata generally dip south–southwest, unless affected by local structures or by faulting and folding related to the strike-slip movement of the Nemaha system. The Nemaha uplift has been an active structural feature since the early Paleozoic, although most of the movement occurred in Pennsylvanian time with the Wichita and Arbuckle orogenies. The Nemaha uplift appears also to have been active during the Devonian, as is evident by the fact that the Frisco Formation of the Hunton Group was deposited on a structural paleohigh around what is now the Oklahoma City field.

Regional and local structure appears to indirectly effect deposition of the Misener by controlling the paleodrainage system developed on the pre-Woodford/Misener surface (Fig. 3).

STRATIGRAPHY

The Misener, in relation to eustatic sea-level fluctuations, may represent the initial part of a transgressive/regressive megacycle that began in middle Givetian (Middle Devonian) (Fig. 4) (Johnson and others, 1985; Mankin, 1987). The previous megacycle, from Lochkovian–Pragian (Early Devonian) to middle Givetian, was characterized by an overall rise in sea level. The beginning of this cycle corresponds to the initiation of the Frisco deposition in Oklahoma and to the earliest phase of the transgression over the unconformity (or over the top of shallowing-upward marine beds). The unconformity at the top of the Frisco represents the end of the oldest transgressive/regressive cycle in the Early Devonian–Middle Devonian megacycle (I of Johnson and others, 1985; Fig. 4). The significance of these relationships are: (1) the Misener and Woodford are parts of the same megacycle, and (2) the Frisco is part of a megacycle that is separate from older units of the Hunton.

In the study area, the Misener is commonly underlain by the Sylvan Shale or Viola Group, and, to a lesser extent, by the Hunton Group and Simpson Group; it is overlain everywhere by the Woodford Shale (Figs. 2,5). Although

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TABLE 1. — SELECTED PREVIOUS MISENER INVESTIGATIONS SHOWING AUTHOR AND DATE, AREA, SCOPE OF INVESTIGATION, AND INTERPRETATION OF DEPOSITIONAL ENVIRONMENT

Author(s)/Date	Area	Scope of Investigation	Interpretation
Penrose (1891)	Northern Arkansas	Outcrop/Reconnaissance	?
Hopkins (1893)	"	"	?
Williams (1900)	"	"	?
Taff (1905)	"	"	?
Ulrich (1911)	"	"	?
Aurin & Clark (1921)	"	"	?
White & Green (1924)	Northeast Oklahoma	Outcrop/Stratigraphic	Eolian
White (1926)	"	"	"
Croneis (1930)	Northern Arkansas	"	?
Cram (1930)	Cherokee & Adair Co., Okla.	"	Marine
McKnight (1935)	Northern Arkansas	"	?
Rau & Ackley (1939)	Keokuk Pool, Seminole Co., Okla.	Field Study	Marine
Borden & Brandt (1941)	E. Tuskegee Pool, Creek Co., Okla.	"	Near-shore Marine
Imbt (1941)	Zenith Pool, Stafford Co., Kansas	"	"
Swanson & Landis (1962)	Northern Arkansas	Outcrop/Stratigraphic	Transgressive
Freeman & Schumacher (1969)	"	Outcrop/Biostratigraphic	Marine
Krumme (1969)	Oklahoma	Geomorphologic	Alluvial (Channel)
Amsden & Klapper (1972)	North-central Oklahoma	Depositional/Stratigraphic	Near-shore marine
Kochick (1978)	Payne & Lincoln Co., Oklahoma	Depositional/Exploration	Alluvial/shallow-marine
Baurenfiend (1980)	Lincoln & Creek Co., Oklahoma	Depositional/Petrologic	Tidal-flat/fluvial
Horner & Craig (1984)	North-central Arkansas	Outcrop/Phosphate Origin	Transgressive
Walker (1984)	Garfield & Grant Co., Oklahoma	Depositional/Diagenetic	Esturine/ tidal-channel
Mansfield & Breckon (1985)	"	Depositional/Petrologic	Shallow-marine (tide/wave)
Cameron (1986)	"	Depositional/Diagenetic	Marine
Bockhorst (1987)	"	Mapping/Depositional	Alluvial
Francis & Mansfield (1987)	"	Depositional/Petrologic	Shallow-marine
Pogue (1987)	"	"	Marine (strandline)
Busanus (1988)	"	Depositional/Paleogeomorphic	Marine
Francis (1988)	"	Depositional/Petrologic	Shallow-marine & storm
Pittenger (1988)	N. E. Oklahoma & N. W. Arkansas	Outcrop/Provenance	Near-shore marine
MASERA (1990)	North-central Oklahoma	Petro-stratigraphic/exploration	Shallow-marine/estuary
Cunningham & Newell (1991)	Zenith Field, Stafford Co., Kansas	Field study/sequence strat.	Low-stand/transgressive ?

the general stratigraphy of the Misener is relatively simple, the internal stratigraphic relationships tend to be complex. In several fields, reservoir heterogeneity, including lateral changes in lithofacies, is reflected in the relative production.

The Misener, which appears to be genetically related to the Woodford Shale, corresponds to the "lower Woodford cycle," which overlies and interfingers with Woodford-like "Misener shales" (Fig. 6). Local correlations within the Misener correspond to minor sequence boundaries, which allow a generalized subdivision that includes (in ascending order): (a) sandstone and phosphatic sandstone, (b) shale, (c) mixed sandstone and dolomite, (d) dolomite, and (e) shale (Fig. 5). There are several intra-Misener stratigraphic deviations from this typical Misener lithologic sequence described above. Most common is the position of the "Misener shale," which most often is located between the sandstone and dolomite, but can be found throughout the Misener section (most commonly at the top). In some areas the "Misener shale" is confused with the overlying Woodford shale.

Correlation of the Misener from well to well is usually difficult outside of local field areas, due to a lack of continuity of the various lithofacies (which is a result of the depositional progradation and intra-Woodford/Misener unconformities) or of difficulty in recognition of subcropping formations of similar lithology.

DEPOSITIONAL ENVIRONMENT AND PETROGRAPHY

The Misener, as discussed, is part of two major transgressive/regressive episodes that occurred during the Devonian. Deposition occurred in a shallow-marine environment that ranged from tidal ridge to estuarine, to tidal flat. Many Misener cores show an overall shallowing-upward sequence and a change from a terrigenous to a carbonate regime—from phosphatic sands upward to sandy dolomites. This sequence, compared with the regional configuration of the Woodford Shale, suggests that the Woodford developed in two cycles, with the Misener section genetically equivalent to the lower Woodford transgressive/regressive cycle.

The composition of the Misener is rather simple, but unusual. Changes in the proportions of various constituents have resulted in noticeable lithologic heterogeneity, which is very significant in the interpretation of depositional environments. In addition, this variability has been a dominant factor in diagenesis and the development of reservoirs. Although quartz and dolomite are the major constituents, the depositional setting of the Misener was closely related to the Woodford shale, as indicated by stratigraphic relationships and composition (shales, phosphatic components, glauconite, and pyrite).

The Amcana Anderson No. 1-2 (NE¼NW¼NW¼ sec.

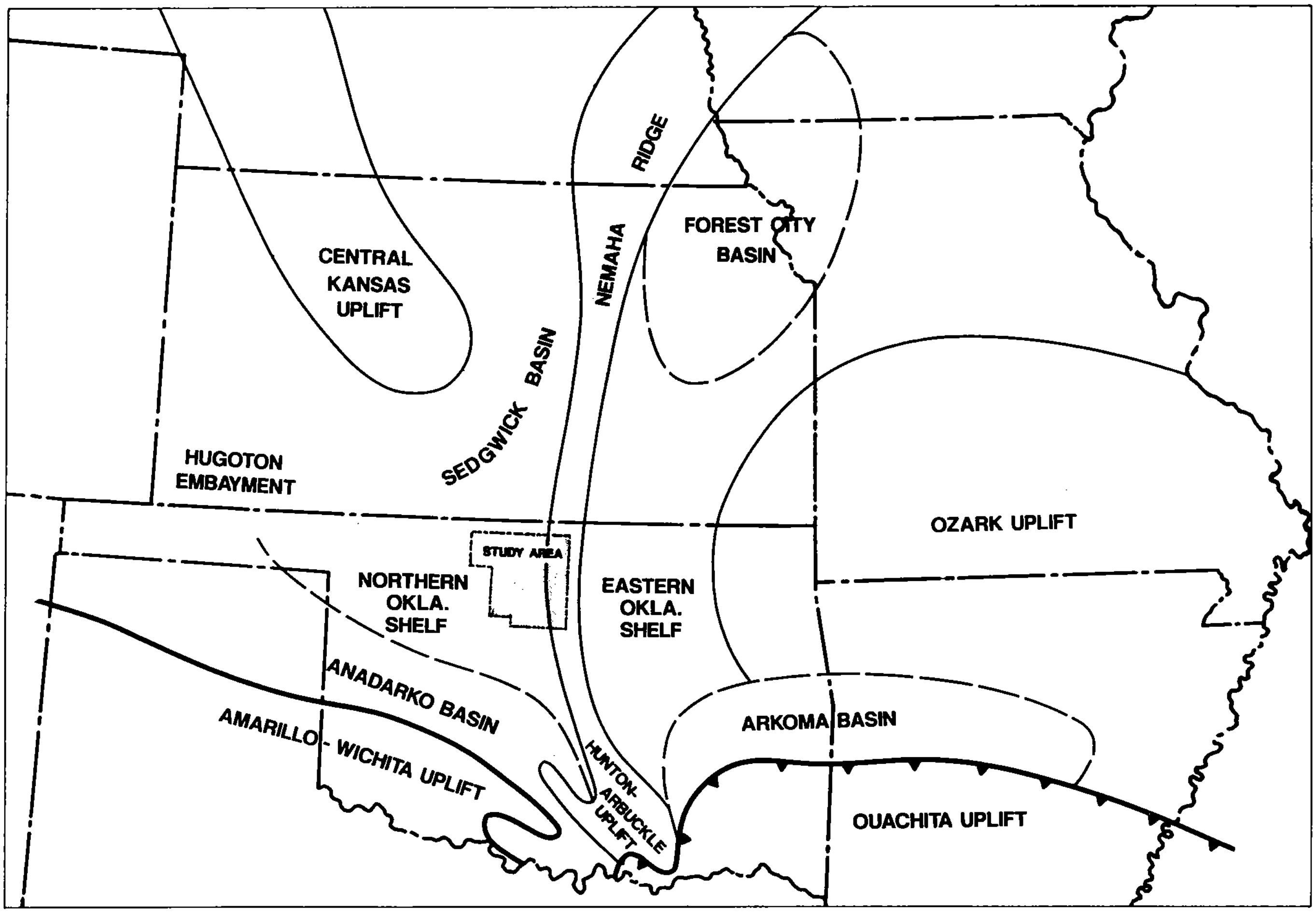

Figure 1. Major geological features of the Midcontinent region and location of the detailed petrostratigraphic study of the Misener sandstone (MASERA, 1990).

2, T. 14 N., R. 3 W., Oklahoma County) serves as an example of the lithologic and stratigraphic interrelationships among the Woodford, Misener, and Hunton (Fig. 7). The cored interval contains the Woodford Shale, Misener sandstone, and Hunton limestone. The Woodford Shale is black and organic-rich, and has a gradational contact with the Misener sandstone.

The Misener is divided into an upper unit (6,069.5–6,080.0 ft) and a lower unit (6,080.0–6,089.5 ft). The upper unit is a black, fine-grained, poorly sorted sandstone (quartzwacke) containing abundant clay matrix. White calcite and dolomite grains and pebbles occur throughout. Inclined shale interbeds (Fig. 8A) are present, as well as relict bedding and flowage features.

The lower unit consists of sandstone with interbeds of shale and dolomitic zones. The sandstone ranges from fine- to medium-grained and poorly to moderately sorted sands with a clay-matrix content that increases upward. The dolomitic zones contain shale laminae, chert nodules, and quartz grains. Flowage features correspond to zones containing diagenetic chlorite. Burrowing and bioturbation are common in the sandstone beds (Fig. 8B). The contact between the Misener and Hunton is sharp and marked with a zone (clast) of chert (Fig. 8C). The Misener interval above this contact contains abundant chert nodules and fossil fragments.

The underlying Hunton limestone is a mudstone to wackestone containing abundant crinoid fragments, and it corresponds stratigraphically to a portion of the lower Chimneyhill Subgroup of the Hunton Group.

Petrographically, the Misener is composed primarily of detrital quartz grains, ranging from 45 to 80% of the rock. Collophane and apatite grains average 3%. Scattered fossils (2–3%), mainly crinoids, are commonly recrystallized. Carbonate clasts, composed primarily of calcite, are common (2%) in the upper unit (Fig. 9A).

Illitic clay is the dominant detrital matrix and constitutes as much as 46% of the rock in portions of the upper unit; chlorite is present in lesser amounts. Calcite, dolomite, and silica are present as cements (Fig. 9B). Late-stage calcite cement and baroque dolomite occur in the lower unit. Dolomite, as very small subhedral rhombs, occurs mainly in the illitic matrix. Silica cement occurs as chert and quartz overgrowths and generally is restricted to the lower unit. Silica replacement of anhydrite, dolomite, and detrital matrix is common in the dolomitic zones.

Porosity (6–12%) is both primary and secondary (due to dissolution of clayey detrital matrix) in origin and consists of intergranular and oversized/moldic pores, respectively.

Based on presence of fossils, phosphate, carbonate zones, bioturbation, detrital-clay content, flowage, and inclined shale interbeds, the Misener interval represents

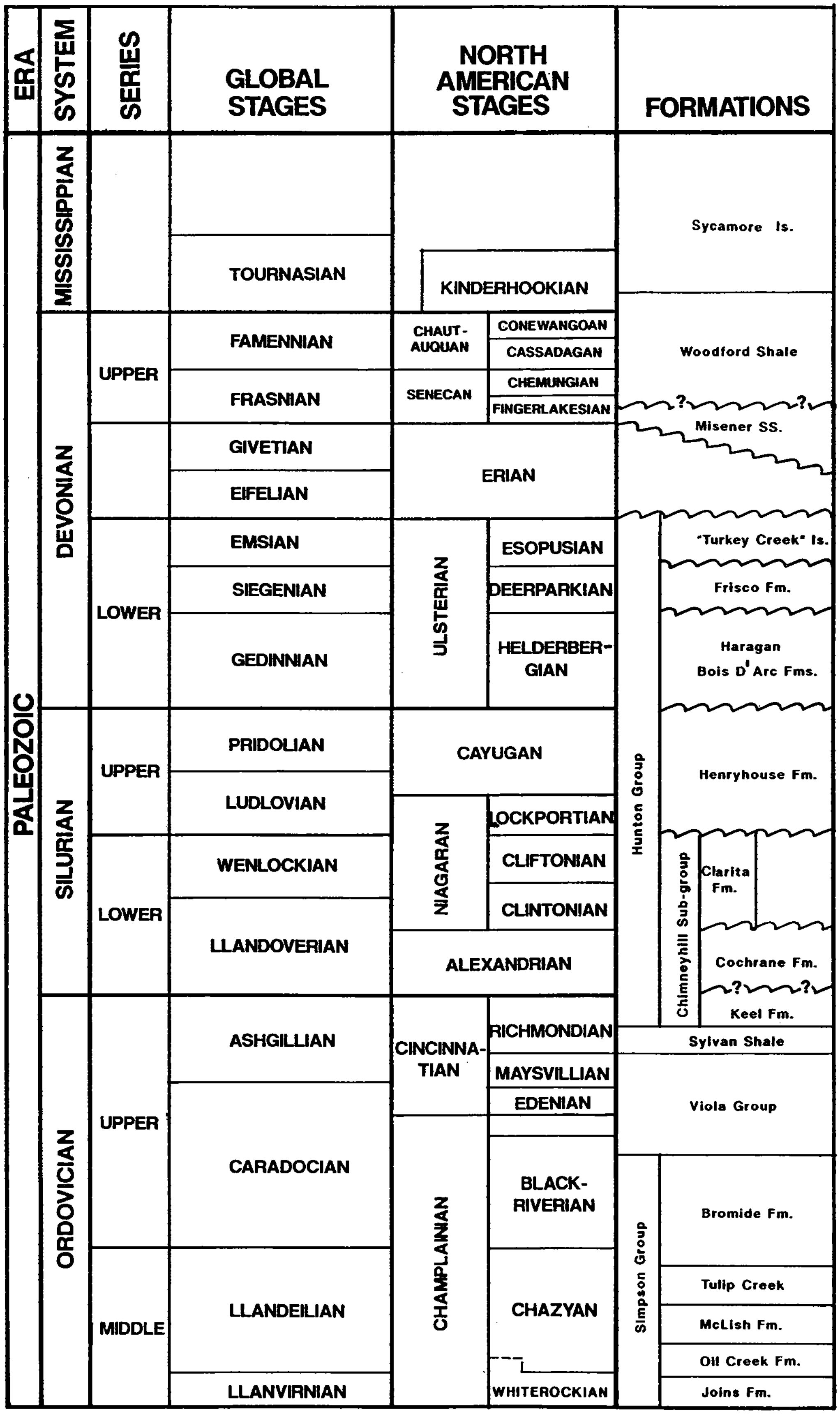

Figure 2. Generalized stratigraphic correlation chart of the Middle Ordovician through Devonian.

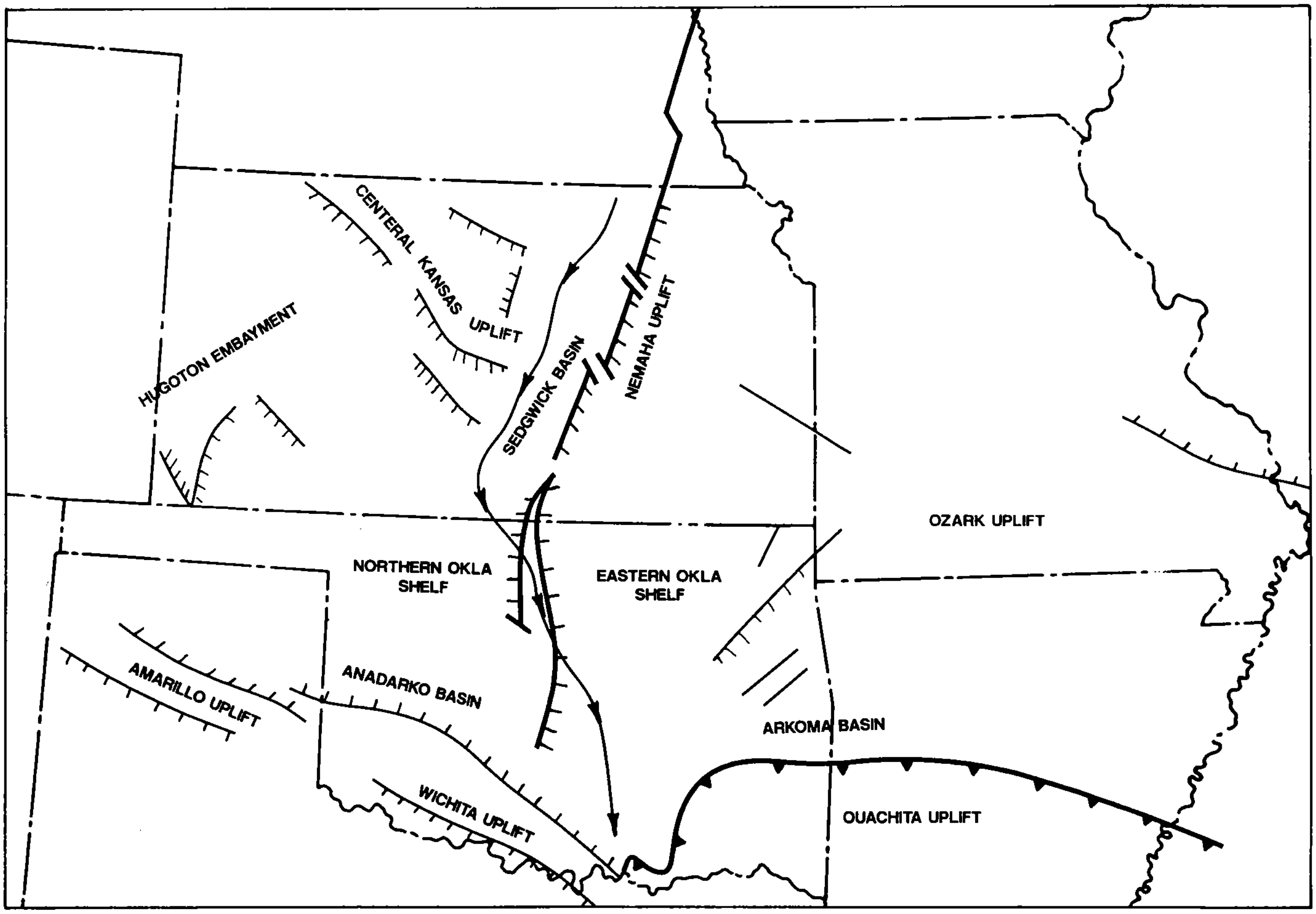

Figure 3. Major tectonic features of the Midcontinent and generalized pre-Woodford/Misener paleodrainage pattern (arrows from north to south).

shallow-marine (tidal-flat) deposition in an embayment/estuary. Figure 10 shows whole-core photographs of the cored interval and Figure 11 is a petrologic log of the cored interval.

In the area along the northern shelf of the Anadarko basin, the Misener contains sandstones, shales, and dolomites. Thick sandstones (as much as 100 ft thick) are composed of two or more sandstone sequences separated by scour surfaces, and/or "Woodford-like" shales. The composition of the sandstones is primarily quartz, with traces of K-feldspar, plagioclase feldspar, and chert. Small, but petrographically significant, amounts of phosphate, glauconite, collaphane, apatite, detrital matrix, and fossil fragments also occur. Illite is the most common clay mineral, and both siliceous and carbonate cements are present. The upper part of the Misener tends to be dolomitic, and in places is a sandy dolomite. This relationship indicates that the Misener formed in a mixed terrigenous/carbonate regime, and it also suggests that the upper Misener, which was deposited when terrigenous input was sharply limited, may represent a shallowing-upward sequence. Sedimentary features in the Misener include: cross-bedding and massive bedding, ripple laminae, scour surfaces, bioturbation and burrowing, and sharp basal contacts.

Porosity (0 to >20%; average 10–15%) in the Misener is both primary and secondary in origin and generally consists of enlarged intergranular and/or oversized pores. Permeabilities can be quite high (up to several darcies) in the more porous sandstone reservoir lithofacies, with initial production rates exceeding 2,000 BOPD and 1 MMcfgpd.

Wells at the edge of some of the Misener fields in the study area contain thin Misener sandstones that are relatively dolomitic and pyritic. The Kremlin area (T. 24 N., R. 6 W.) shows the best development of upper dolomitic/shaly Misener and lower sandy/porous Misener.

UNCONFORMITY RELATIONSHIPS

One of the most important aspects to consider in determining the Misener's erratic distribution and ultimate reservoir geometry is its relationship to Middle Devonian unconformities and resultant paleogeomorphology.

Two major unconformities are indicated in the Late Devonian. The post-Frisco (pre-Woodford/Misener) unconformity is commonly thought to be the most significant, because it is considered the cause of most Hunton erosional features and because of the obvious change in lithologies from shallow-marine carbonates to deeper marine anoxic shales. However, findings from MASERA Corp. Hunton study of the Anadarko basin indicate that the dominant unconformity is the pre-Frisco/post-Bois d'Arc unconformity (MASERA, 1990). It is probable that the strong drainage pattern on the Hunton surface, as well as that on the Sylvan, Viola, and Simpson subcrops, occurred during the pre-Frisco unconformity and was then rejuvenated by the pre-Woodford erosion. In areas where no Frisco is found (i.e., northern shelf of the Anadarko basin), the pre-Frisco and pre-Woodford/Misener unconformity surfaces are merged into one unconformity; however, it is believed that little Misener was deposited until after the post-Frisco unconformity.

Although seldom discussed in the literature, the pre-Frisco unconformity, based on regional stratigraphy, appears to represent a period of significant erosion. The top of the pre-Frisco Hunton section is defined by an angular unconformity which is best developed basinward of the Hunton hingeline. Over most of the area it is impossible to differentiate between the two unconformities; however, the relationship of the Frisco to the older Hunton units indicates that it was deposited on a marked angular unconformity and that it overlies many of the pre-Frisco units—from the Chimneyhill to the Bois d'Arc. Thus, this relationship strongly suggests that most of the Hunton was eroded prior to Frisco deposition. Additional evidence for the magnitude of the pre-Frisco unconformity is that relative sea level apparently was lowest in the Devonian prior to Frisco deposition (Fig. 4).

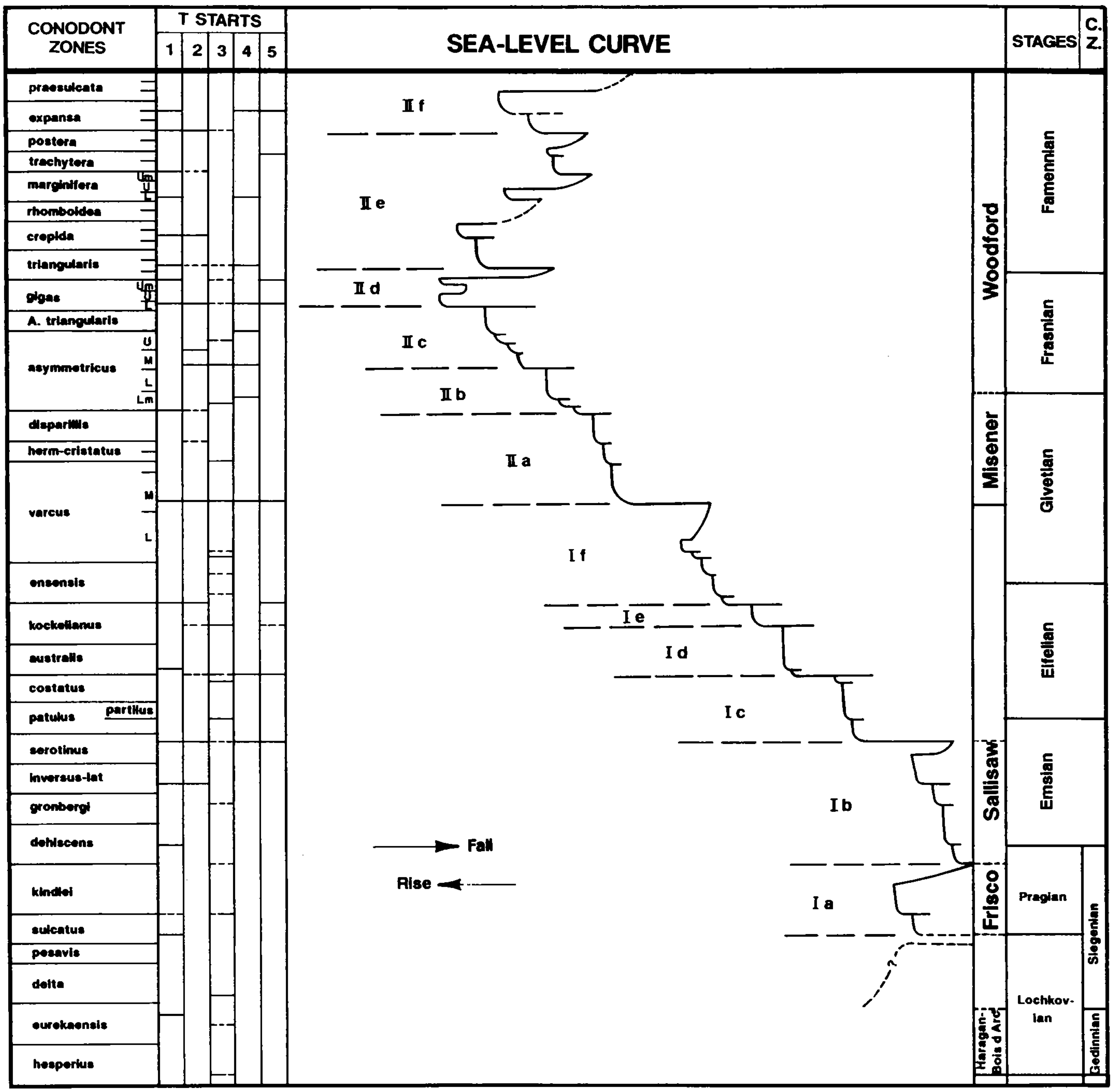

Figure 4. Qualitative eustatic sea-level curve for the Devonian transgressive/regressive cycles and their relationship to Devonian conodont zones and to Oklahoma stratigraphic units (after Johnson and others, 1985, and Mankin, 1987).

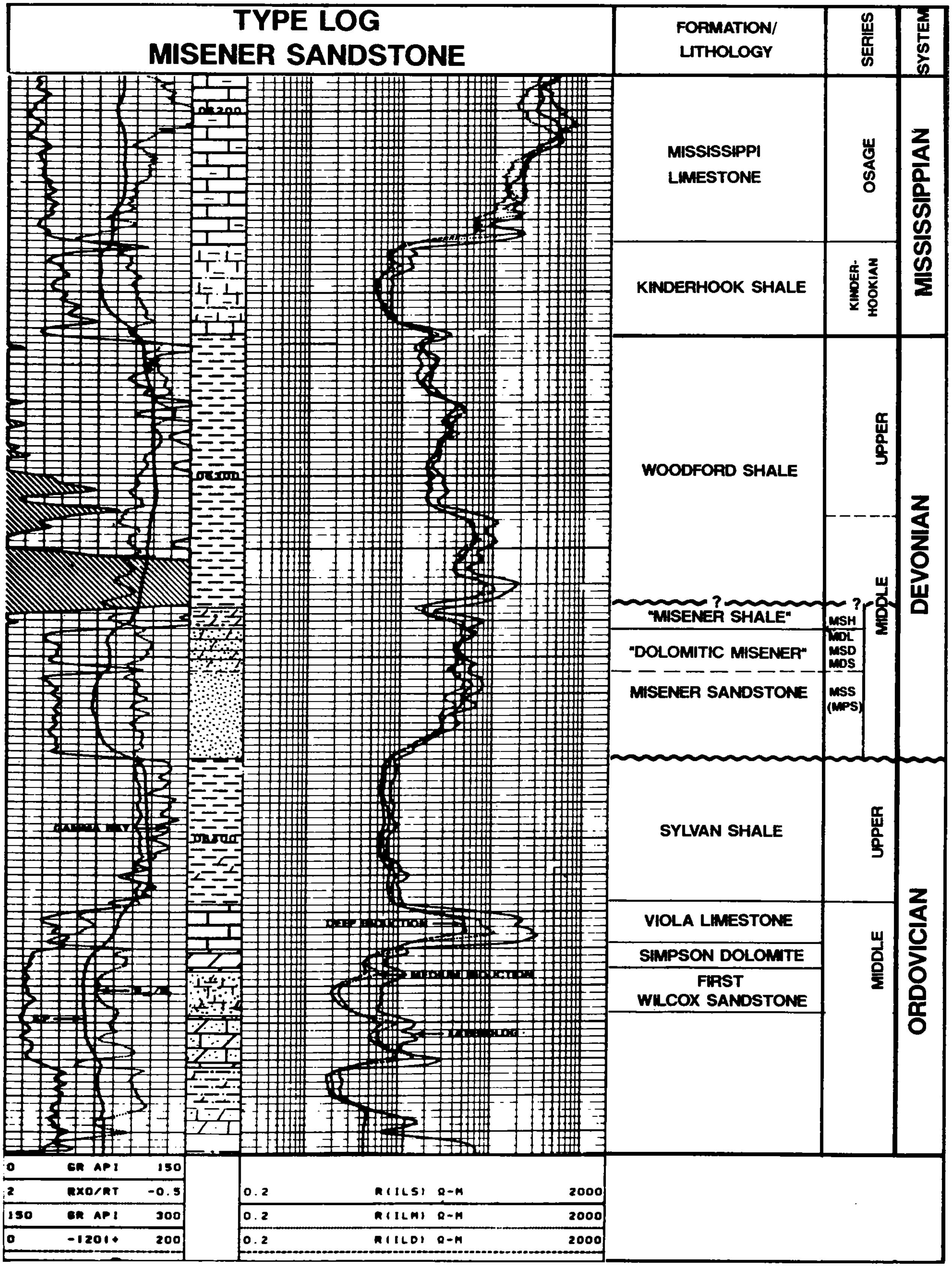

Figure 5. Type log (dual-induction) of a typical Misener section showing Misener lithofacies and bounding strata.

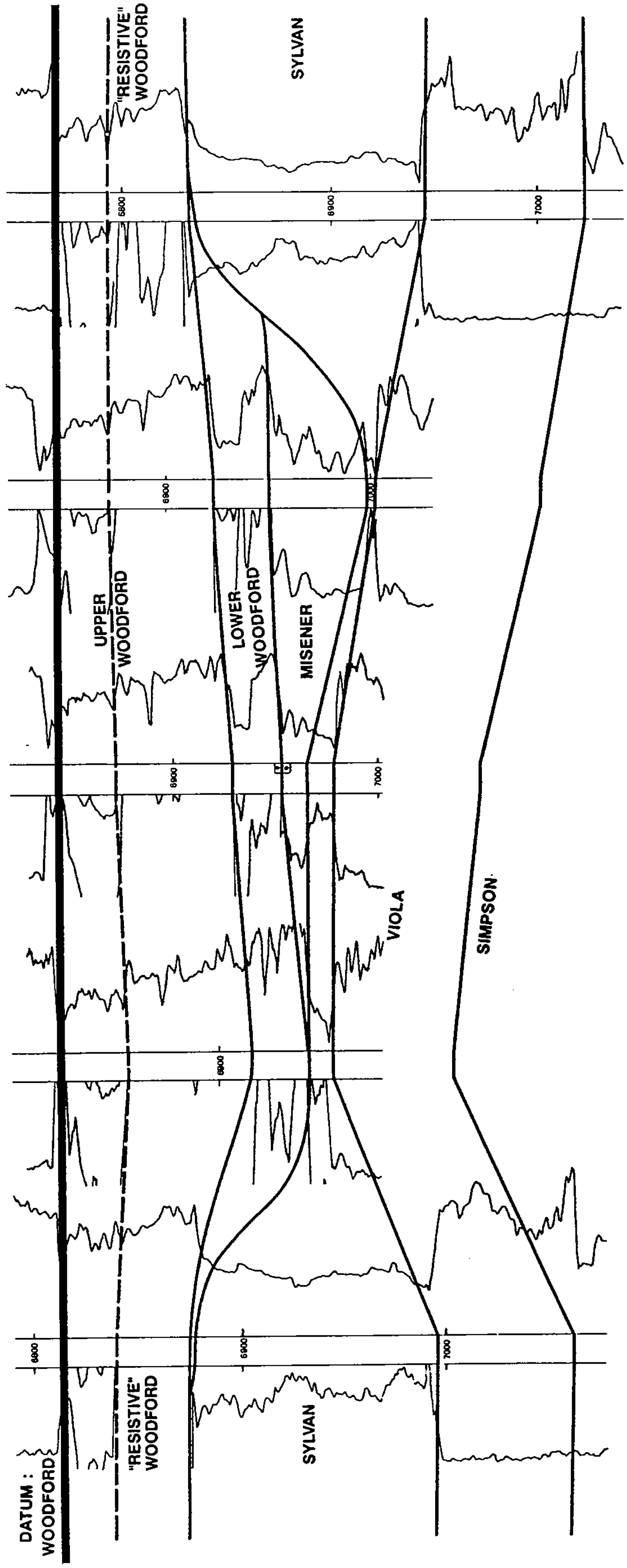

Figure 6. Stratigraphic cross section in the study area showing the relationship between Misener and upper ("resistive") and lower Woodford.

After deposition of the Frisco, another low-stand developed and the Hunton was eroded to near its present-day subcrop configuration. Although the actual duration of the low-stand is unclear, it may have been a relatively short time, based on the absence of dolomite and karst features in the Frisco. The subsequent "Woodford transgression" actually consisted of two transgressive/regressive cycles; the first of which is represented by the Misener ("lower Woodford cycle"), and the second is represented by the Woodford. Intra-Woodford and intra-Misener unconformities, or disconformities, are indicated by both petrography and correlations. Based on conodonts, Amsden and Klapper (1972) concluded that Woodford deposition in southern Oklahoma, at least in part, was contemporaneous with Misener deposition in north-central Oklahoma.

DISTRIBUTION AND SOURCE

The Misener occurs primarily in north-central and northeastern Oklahoma and south-central Kansas (Fig. 12). Examination of several well-exposed outcrops in north-central Arkansas and eastern Oklahoma, along the margin of the Ozark dome, clearly show the stratigraphic relationships associated with the pre-Woodford/Misener unconformity. The Misener (Sylamore) sandstone on outcrop is relatively thin and extensive, and generally is composed of

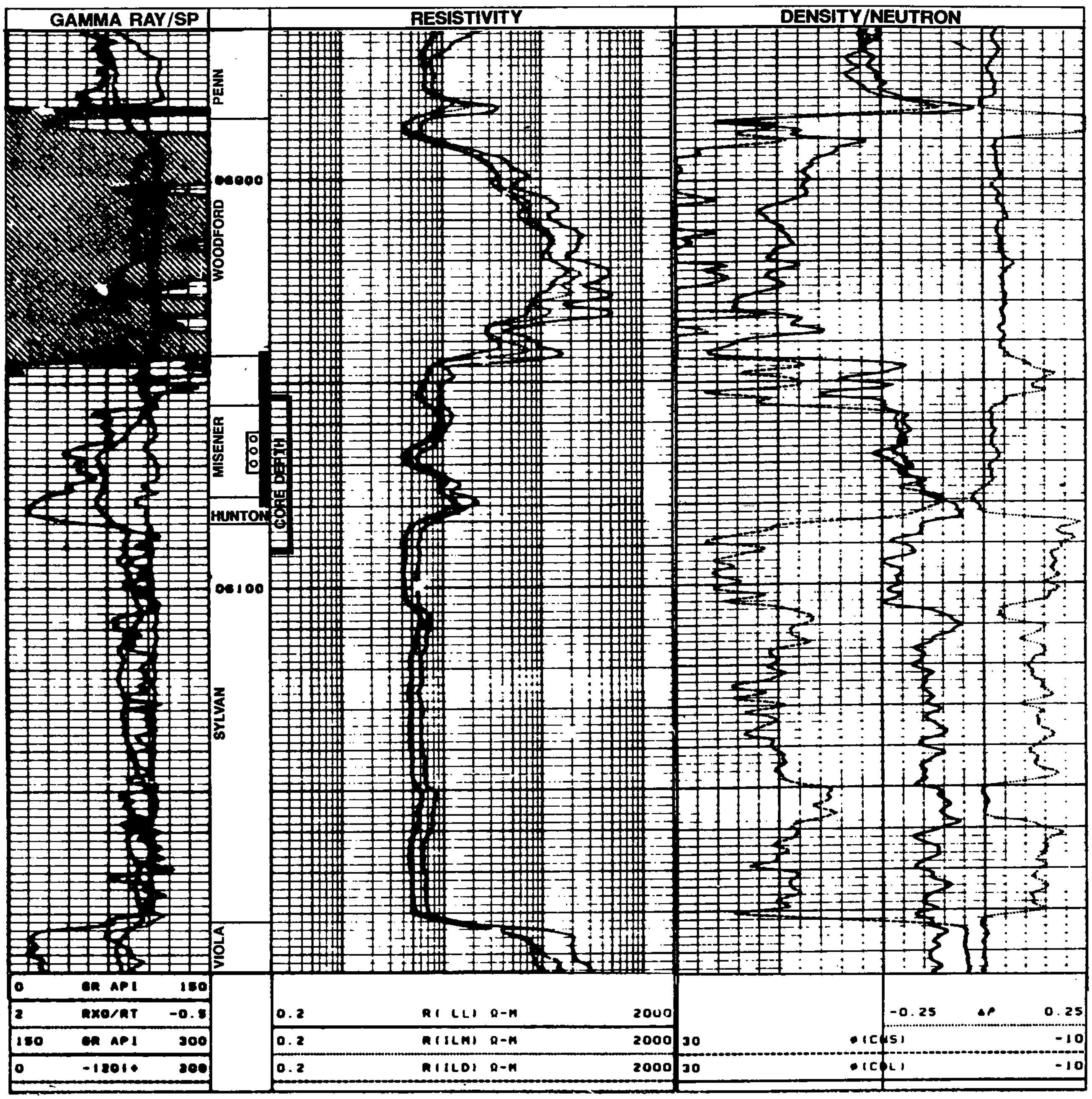

Figure 7. Composite log (dual-induction and compensated density/neutron) of the Amcana Anderson No. 1-2 (NE¼ NW¼ NW¼ sec. 2, T. 14 N., R. 3 W., Oklahoma County, Oklahoma).

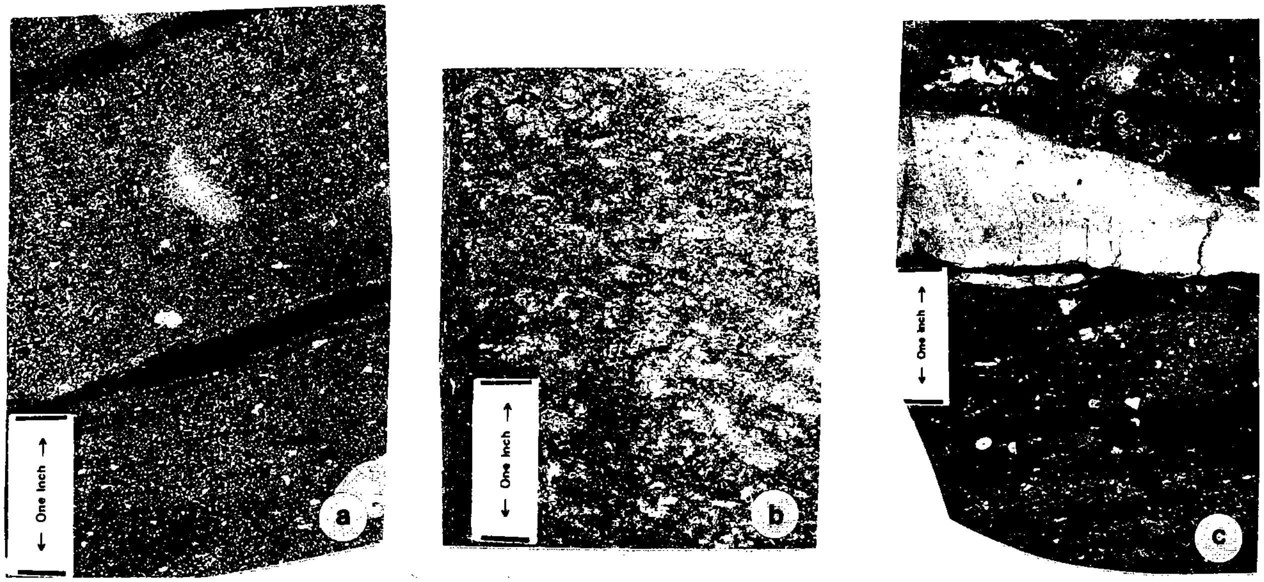

Figure 8. Core photographs of the Misener sandstone from the Amcana Anderson No. 1-2 core: (*A*) quartzwacke with carbonate clasts and low-angle crossbedding (6,070 ft); (*B*) sandstone with abundant matrix, flowage features, and bioturbation (6,080 ft); (*C*) sharp contact between Misener and Hunton, marked by a chert zone (6,088.6 ft).

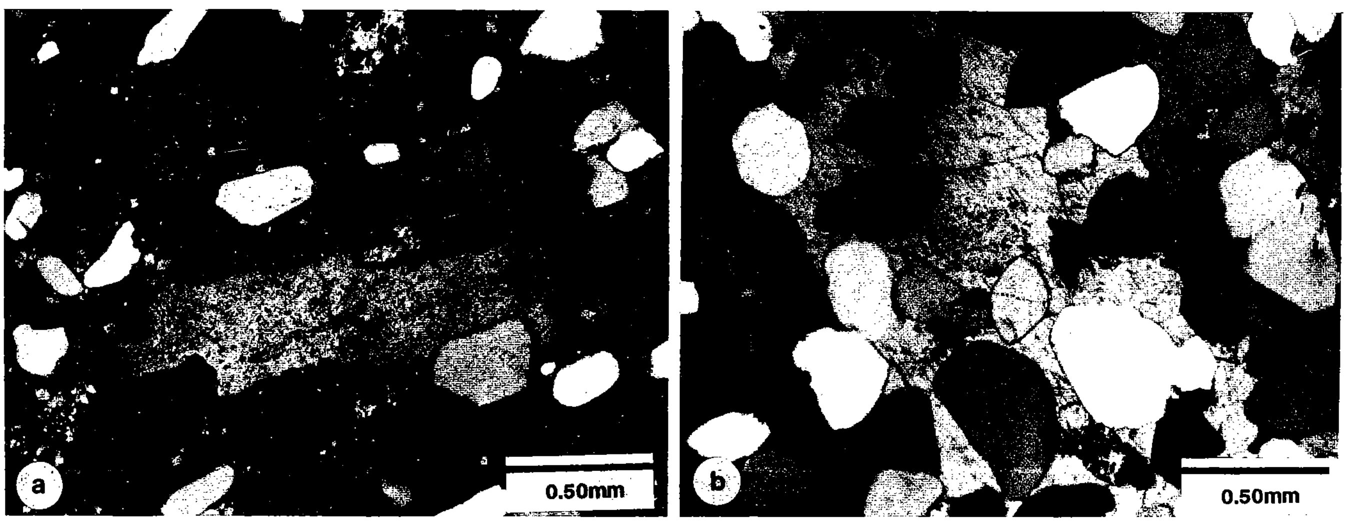

Figure 9. Thin-section photomicrographs (crossed-nicols light) of the Misener sandstone from the Amcana Anderson No. 1-2 core: (*A*) sandstone with abundant clay matrix, small dolomite rhombs, and a carbonate clast (6,074.5 ft); (*B*) poikilotopic calcite occurring as a pore-filling and replacement cement.

a fine- to medium-grained, light-gray to brown, quartzose sandstone with dolomite and traces of glauconite, phosphate, chert, and pyrite. Features associated with probable karstic profiles occur where the Misener (Sylamore) sandstone is in contact, or associated, with portions of the Devonian Penters Chert in the Arkoma basin, and as karstic-fill in fissures/cavities within the Hunton across Arkansas and Oklahoma.

Distribution of the Misener is quite erratic across northern Oklahoma. In the study area it is generally present as pods to irregularly shaped bodies in areas commonly corresponding, but not limited, to the Sylvan subcrop. The Misener also occurs as local sheet sands and as isolated pods related to Hunton paleokarst. It is thought that some of the isolated bodies of Misener may be interconnected, even though in some areas that type of configuration is not presently substantiated by current well control. East of the Nemaha uplift, Misener distribution appears to be more interconnected, as compared to its distribution west of the Nemaha. This relationship perhaps is due in part to the greater density of well control to the east, but may also reflect differences in distinct Misener facies, related to depositional controls such as paleodrainage patterns, paleotopography, eustatic sea-level fluctuations, and specific source areas.

The Misener sand probably was derived from erosion of Simpson Group sandstones and/or equivalents. Because of the alluvial/fluvial-formed surface on which the

Figure 10. Whole-core photographs of the cored interval (6,053–6,091 ft, core; 6,042–6,080 ft, wireline) from the Amcana Anderson No. 1-2. Photos here show the interval 6,069.9–6,090.3 ft.

Misener was deposited, it seems logical to assume that the Misener sand was derived from Simpson subcrops to the north. However, due to the marine nature of the Misener sand, and due to the lack of Misener sand along the "Sylvan subcrop valley" north of the Kremlin area, it is possible that the sand was derived from eastern Oklahoma and the Ozark dome area, where there is an abundance of both Simpson and Misener (Sylamore) sand (Fig. 12). This hypothesis is supported by the poorly developed Viola–Simpson subcrop paleohigh that allowed sand movement from east to west by marine currents. An alternative explanation is that fluvial sands, derived from adjacent Simpson subcrops to the north and in Kansas, were eroded and redeposited during the "lower Woodford cycle."

Also, there is evidence that local paleostructures may have controlled the Misener source, distribution, and even the depositional environment, as in portions of the Cushing field, in parts of Lincoln and Creek Counties, Oklahoma (Fig. 13). At the time of Misener deposition, the area of the Cushing field was undergoing uplift and erosion. Erosion cut down into the Ordovician during this uplift, exposing Simpson sands along the axis of the Cushing ridge. Bauernfeind (1980) showed that the Misener distribution, petrology, and underlying paleogeography in-

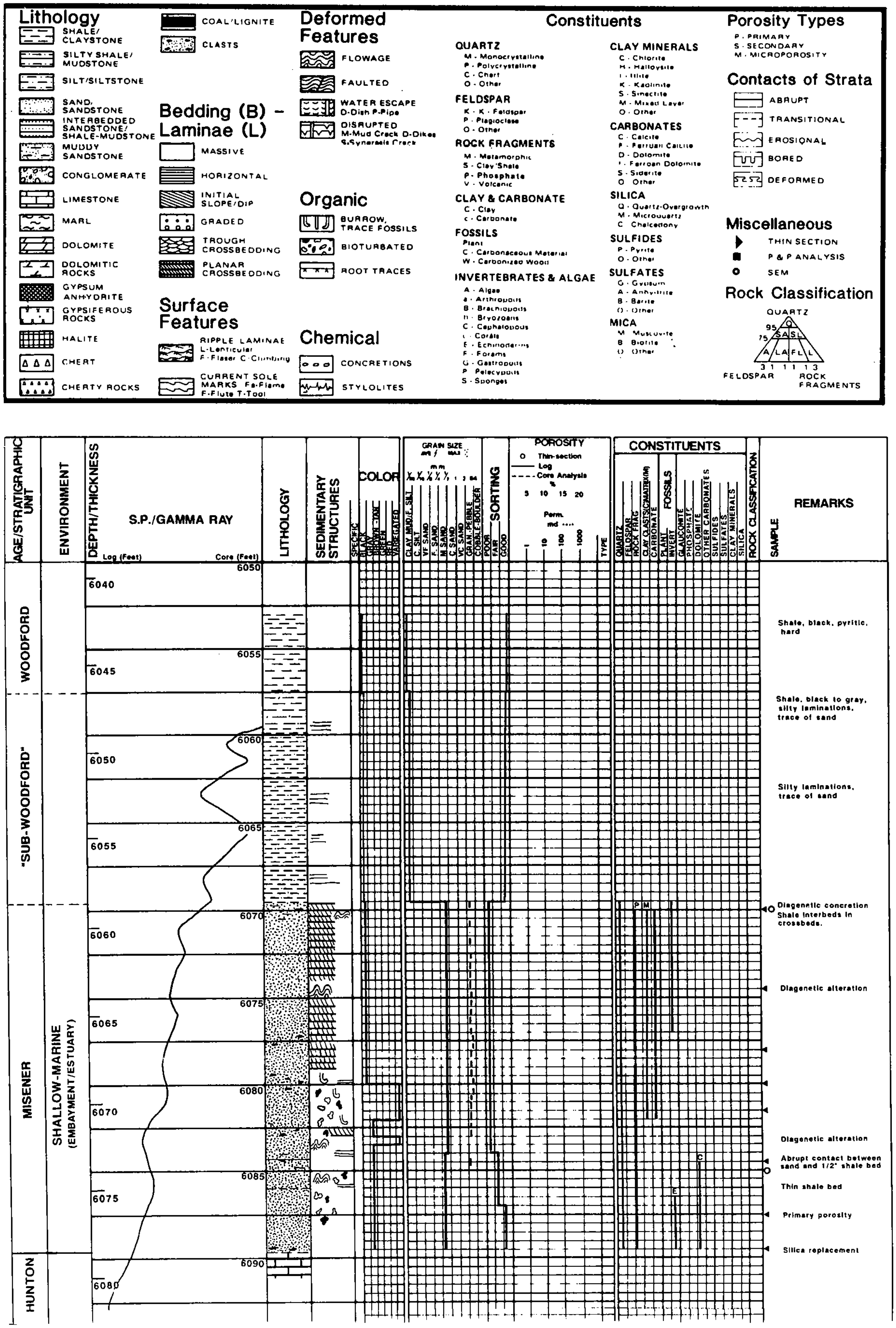

Figure 11. Petrologic log of the Amcana Anderson No. 1-2 core showing the lithologic, mineralogic, and sedimentary characteristics of the Woodford, Misener, and Hunton.

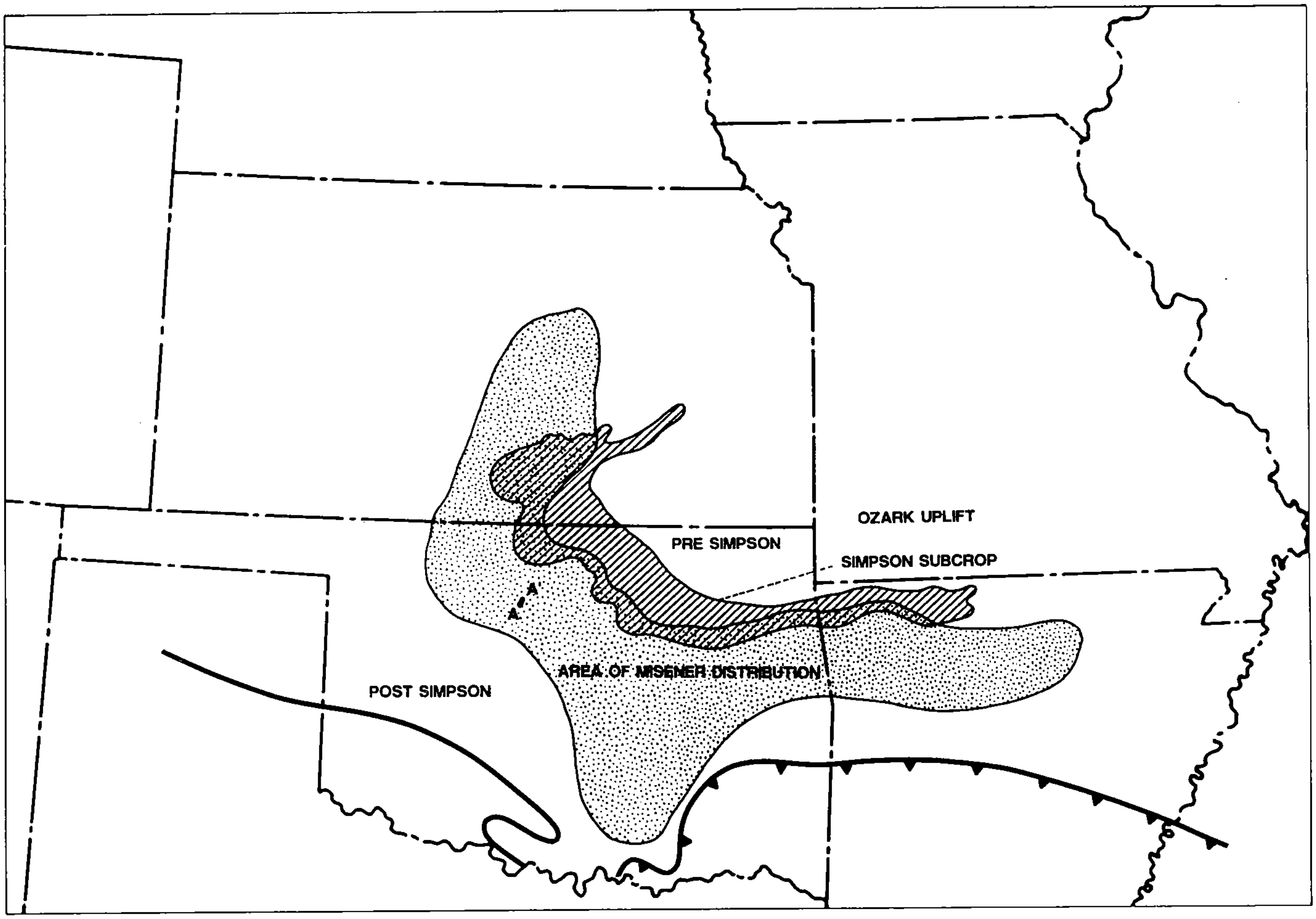

Figure 12. Approximate distribution of the Misener sandstone and its relationship to Simpson and pre-Simpson subcrop (MASERA, 1990).

dicates Misener-age stream/river channels drained the southern flanks of the Cushing ridge (Fig. 13). Southeast, along and near the updip limit of the Hunton subcrop, the Misener contains dolomite and glauconite, in contrast to clean, well-sorted sands indicative of the channel-like deposits to the north. The Misener distribution, to the south in this area, also tends to lose its channel-like geometry. This, along with the petrographic evidence, suggests reworking of Misener sands in a marine-influenced environment, probably a tidal-flat and/or tidal-channel setting. Thus, local Misener distribution resulted from the presence of two distinct, but interrelated, depositional environments, and apparently one local source. Similar examples of a local Misener source exist in proximity to eroded Ordovician paleostructures along the Nemaha uplift (i.e., Lucien and Polo field areas, western Noble County, Oklahoma).

PALEOGEOGRAPHY

Since the distribution and source of the Misener is closely associated with pre-Woodford paleotopography, a close examination of the pre-Woodford subcrop patterns is important.

The main controls on pre-Woodford paleotopography were structure (folding, faulting, fracture patterns, etc.), outcrop pattern, and lithologies of the exposed formations and component units. To a large extent, the structure of the area controlled the drainage and outcrop patterns, as well as the overall topographic relief, whereas the lithologies influence local topographic expression and preferential drainage patterns. Regional tilting of pre-Woodford strata to the southwest, and subsequent truncation and erosion by streams, resulted in a parallel/subparallel outcrop pattern of the Simpson, Viola, Sylvan, and Hunton (Fig. 14). An examination of the overall trend of the pre-Woodford subcrop patterns shows a distinct northwest to southeast trend, except where the "Sylvan valley" crosses the Nemaha uplift and turns south to follow the uplift for more than 20 mi before resuming its southeastward trend (Fig. 14).

The irregular pattern of each pre-Woodford subcrop boundary/contact is due primarily to etching and erosion by subaerial drainage systems during the pre-Frisco unconformity, which was later enhanced by pre-Woodford erosion. Local V-shaped patterns represent the intersection of stream valleys with the inclined formation contacts. These V-shaped subcrop patterns typically point in the direction of dip; the lower the angle of dip, the larger the V-shaped outcrop pattern. In areas of gentle dip or near-horizontal strata, such as along the Hunton subcrop, stream erosion and incision have resulted in a dendritic subcrop pattern. Areas with broader and/or nonparallel subcrop

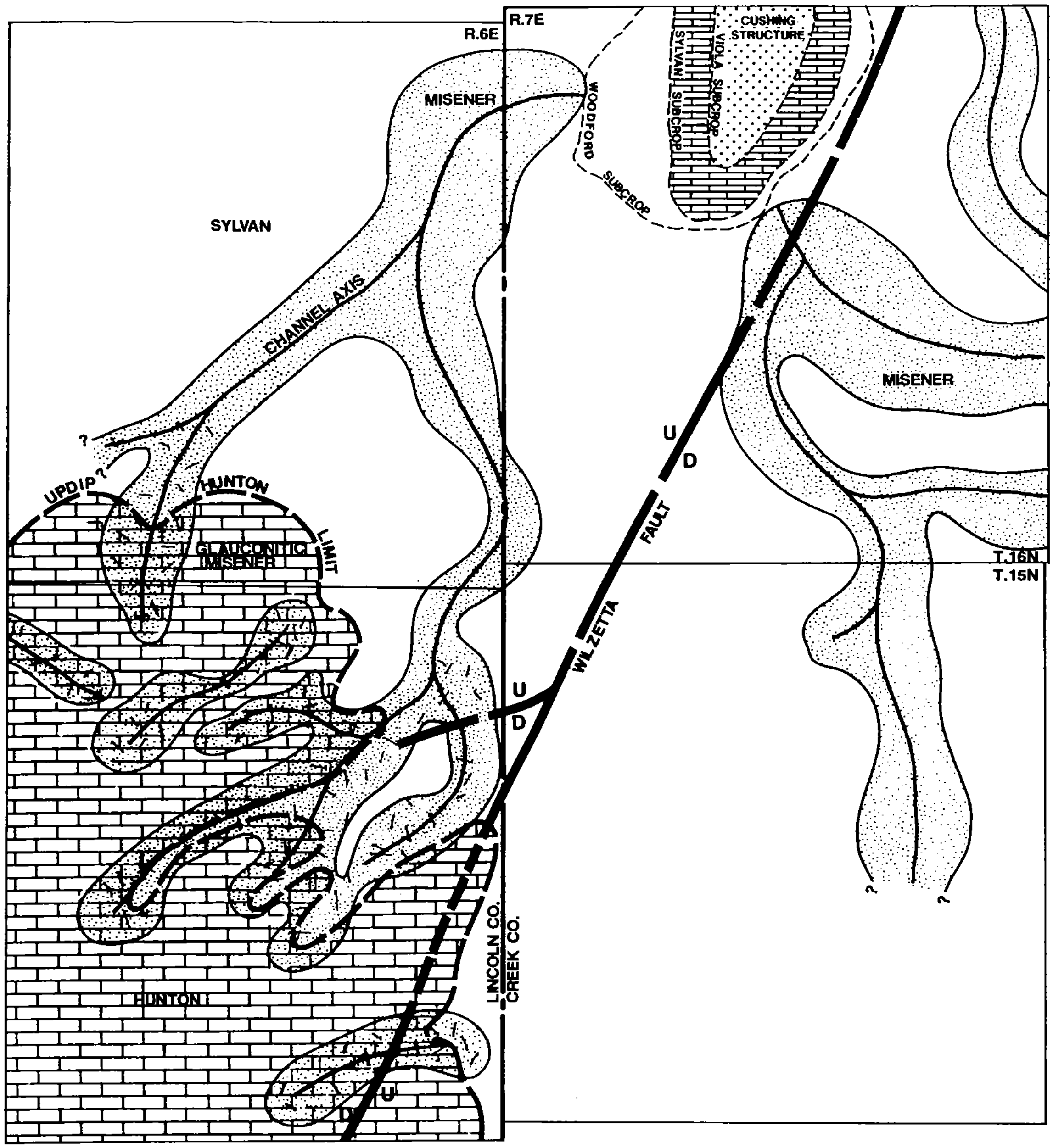

Figure 13. Paleogeology of the Misener showing its distribution and relationship to the Hunton subcrop, and inferred local source along the southern flank of the Cushing structure, Creek County, Oklahoma (after Baurenfiend, 1980).

patterns may represent regional structural features not necessarily associated with stream valleys. However, for exploration purposes, the proximity to local V-shaped subcrop patterns should be thoroughly evaluated for Misener distribution.

Thickness patterns of the Woodford/Misener sequence enables inferences to be made about paleodrainage patterns, position of cuesta slopes, and, in some local cases, recognition of paleosinkholes related to Hunton karstification or paleostructures. The Hunter field and West Kremlin field, in portions of T. 24 N., Rs. 4–6 W., Garfield County, are good examples of distinctly different structural influences on the control of pre-Woodford/Misener paleodrainage (Fig. 14). A distinct change in the direction of paleodrainage around the Hunter paleostructure is reflected by the thickness pattern of the Woodford. Paleostructure has also inversely influenced Misener distribution in many areas, as in the West Kremlin field where the paleodrainage system apparently breached an underlying paleoanticline and the resulting paleovalley was filled with Misener sediments.

In general, the Hunton paleotopography consisted of low-lying, somewhat resistive limestone hills and cuestas drained by low-gradient streams, which were influenced by a mature karstic terrain, resulting in a dendritic drainage pattern as shown by subcrop patterns (Fig. 14). Average thickness of the Hunton increases at about 5–15 ft/mi southward from its truncated edge. In the area where isopach values are 100–150 ft, however, there is a marked break in paleoslope. Southward of the break in paleoslope

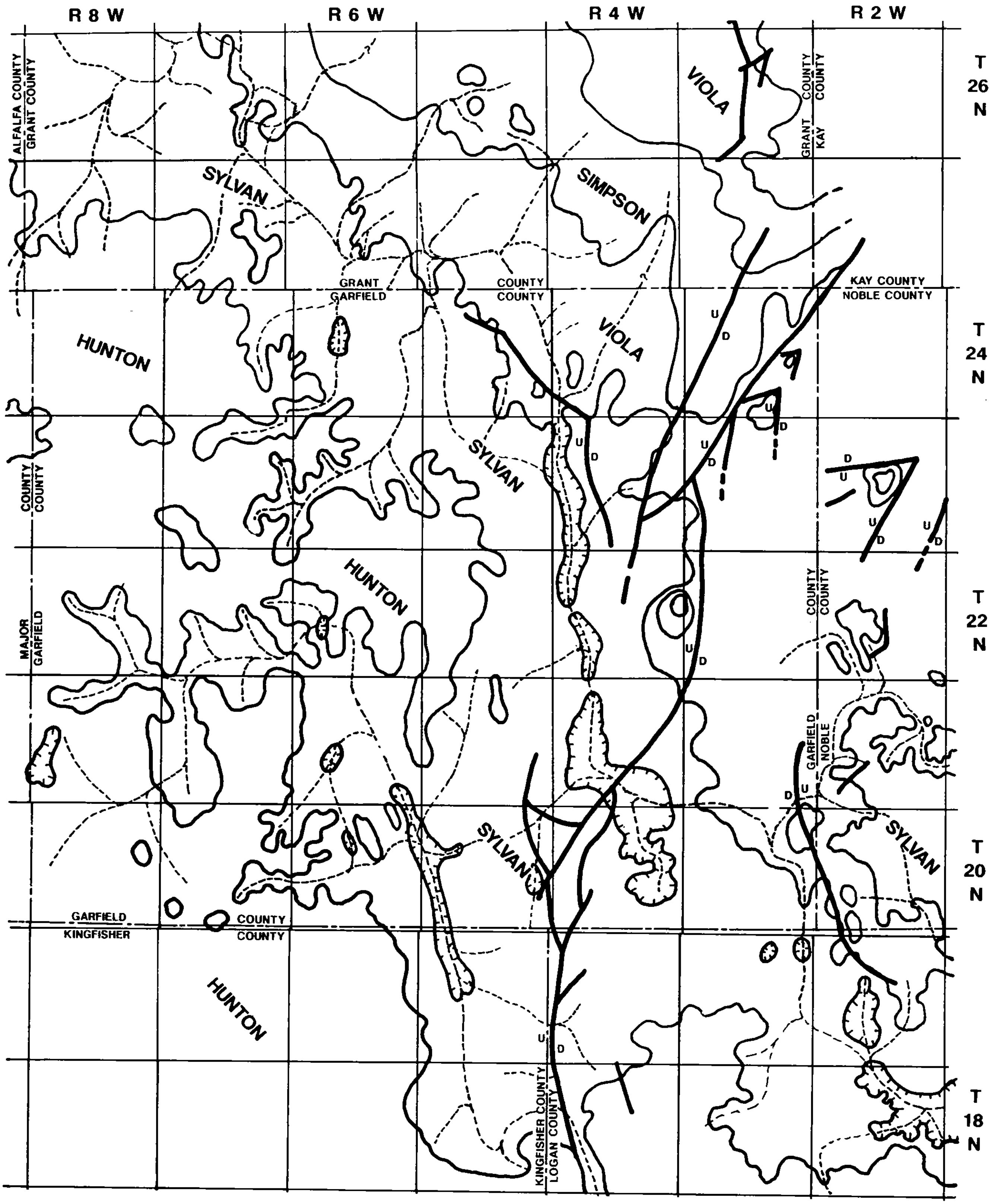

Figure 14. Pre-Woodford/Misener subcrop map in portions of Grant, Garfield, Kingfisher, Kay, and Noble Counties, Oklahoma, showing the inferred paleodrainage pattern (dashed lines) and major faults (bold lines) associated with the Nemaha uplift (MASERA, 1990).

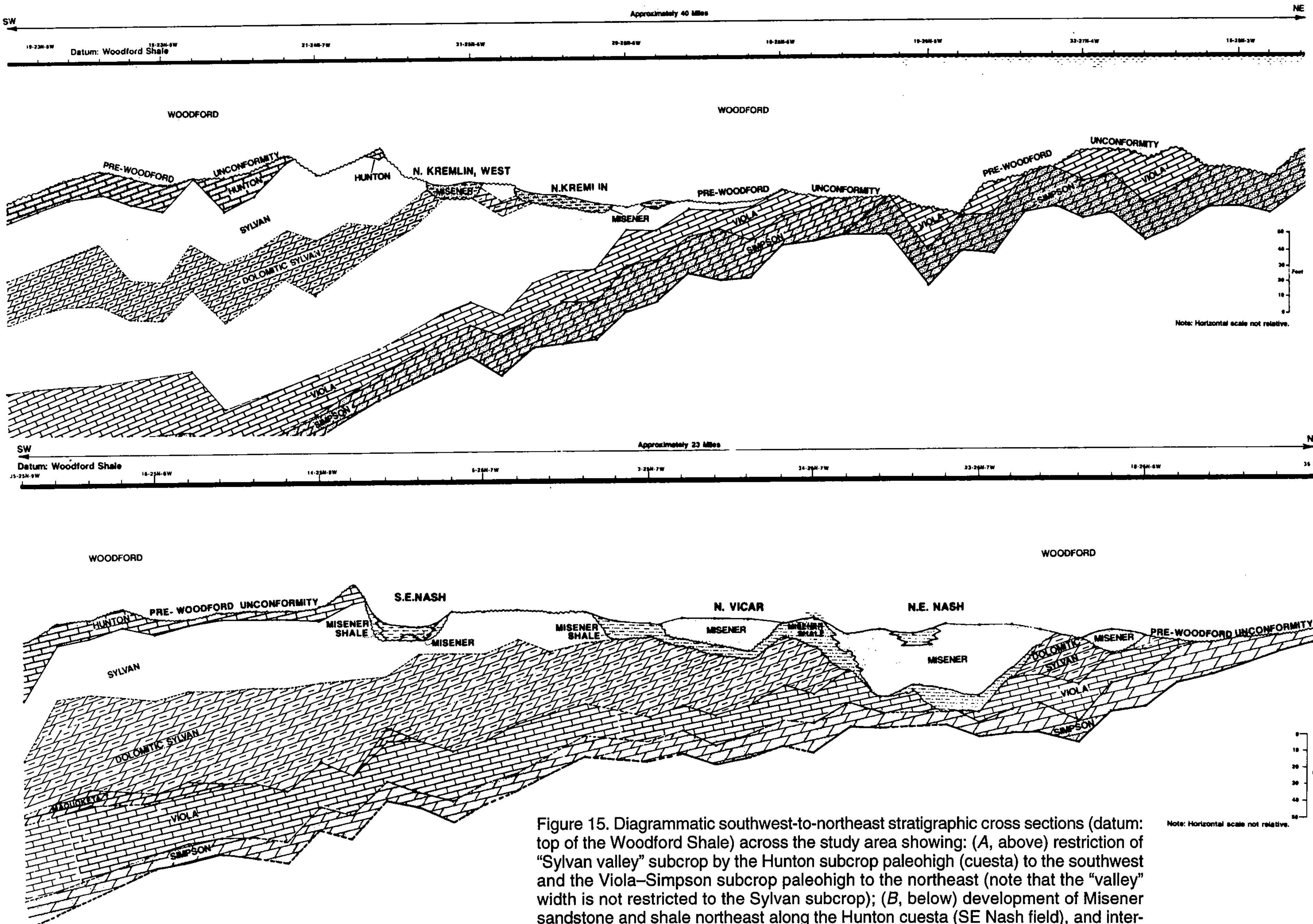

Figure 15. Diagrammatic southwest-to-northeast stratigraphic cross sections (datum: top of the Woodford Shale) across the study area showing: (*A*, above) restriction of "Sylvan valley" subcrop by the Hunton subcrop paleohigh (cuesta) to the southwest and the Viola–Simpson subcrop paleohigh to the northeast (note that the "valley" width is not restricted to the Sylvan subcrop); (*B*, below) development of Misener sandstone and shale northeast along the Hunton cuesta (SE Nash field), and interfingering of the "Misener shale" and Misener sandstone in several other Misener fields within the dolomitic Sylvan subcrop to the northeast (MASERA, 1990).

(the hingeline), the Hunton thickens at rates of 20–50 ft/mi, generally increasing in thickness from 150 ft to 350 ft in distances of <6 mi. From this hingeline, the thickness increases basinward gradually at about 15–20 ft/mi, with local variations to the margin of the deep basin (about –15,000 ft), where Hunton rocks are thicker than 1,500 ft.

The Sylvan surface primarily constituted a low-lying, nonresistive (shale) area or "valley" which was situated between the Hunton, to the southwest and west, and the Viola and Simpson, to the northeast and east (Fig. 15). Slightly dolomitic, and more resistant, zones within the Sylvan may have influenced local paleotopography and distribution of Misener sediments. A Sylvan Shale isopach map may be used as a paleotopographic indicator. Major paleodrainage systems are recognized where Sylvan isopach values tend to be thin (or absent), indicating erosion by streams or truncation related to local or regional structural features. Linear/elongated trends of thin and/or absent Sylvan generally correspond with thick Woodford, and in some areas with the presence of Misener strata.

The major paleostream system in the study area is located along the Sylvan Shale subcrop (Fig. 13). Most of the well-developed Misener production is located along this system (or "Sylvan subcrop valley"). Regionally, it represents a relatively short section of a major drainage system that extended northward into Kansas, and perhaps as far as Canada, and which appears to have been controlled by faulting and fracturing along the Midcontinent rift system (Fig. 3). The "Sylvan subcrop valley" was restricted on the southwest by the Hunton subcrop paleohigh and, to a lesser extent, on the northwest by the Viola–Simpson subcrop paleohigh. The latter paleohigh was not as prominent as the former, as indicated by an increase in Woodford thickness from the Hunton to the Simpson subcrop, with only a slight additional increase in thickness across the valley (Fig. 15).

SUMMARY

In summary, the Misener sandstone represents a complex depositional sequence whose distribution and geometry is influence by paleogeomorphology. It is very similar in many respects to the lower Morrow and Spiro sandstones of the Anadarko and Arkoma basins, respectively, in that all are associated with regional unconformities and have "channel-like" and/or "sheet-like" distributions, but are petrographically shallow marine to estuarine.

The Misener was deposited on a substrate that was channelized by mostly suspended-load streams during development of the pre-Frisco and pre-Woodford/Misener unconformities. The resulting paleomorphology significantly controlled the distribution and geometry of the Misener lithofacies. Earliest sand deposition appears to have occurred during the initial flooding stage, or transgression, with the drowned channels/valleys becoming shallow-marine to estuarine environments. The early sand source probably was the Simpson subcrop along paleohighs to the northeast and isolated paleostructures associated with local uplift/faulting. As flooding of the area continued, marine currents dominated, thus bringing in sand from the east, from as far away as the exposed Ozark dome. The influx of sand in the study area was intermittent, with marine reworking into the lower areas of pre-existing channels. Examination of the internal geometry of the sands indicates deposition in several episodes, which was probably the result of several minor changes in sea level. Because of the relative lack of sand and the sporadic nature of deposition, carbonate environments apparently existed concurrently with the clastic environment and developed best in areas where there was little clastic input, or during times when clastic influx was low.

Reliable determination of Misener distribution and reservoir geometry can only be made by understanding the depositional environment and interrelated critical controlling factors, such as stratigraphy, petrography, source, and paleogeomorphologic relationships resulting from the pre-Frisco and pre-Woodford/Misener unconformities. Future exploration of the Misener may ultimately depend on integration of geologic parameters with geophysical data and tools such as 3-D seismic.

ACKNOWLEDGMENTS

Appreciation and gratitude go to MASERA Corp. for allowing presentation and publication of portions of the *Misener Sandstone Study of Oklahoma*. Thanks go to several geologists who worked on various aspects of the study; they include: Ellen Hooker, David Hooker, John Shelton, and Dennis Presbindowski. Thanks also go to Rick Elliot for drafting many of the maps, cross sections, and diagrams. Special thanks to Eldon Cox and Ken Johnson, of the Oklahoma Geological Survey, for providing (respectively) many cores and a forum to present this work.

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Stratigraphy of the Cason Shale, the Brassfield, St. Clair, and Lafferty Limestones, and the Penters Chert (Upper Ordovician–Lower Devonian) in the Arkansas Ozarks

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ABSTRACT.—Equivalents to Oklahoma Hunton Group strata are present in the Upper Ordovician to Middle Silurian Cason Shale and the Middle to Upper Silurian St. Clair and Lafferty Limestones of Arkansas. The Arkansas units agree closely, both in lithology and fauna, with their Oklahoma correlatives, indicating a parallel depositional/erosional history. Strata of the Arkansas Cason Shale include units equivalent to the Ideal Quarry Member of the Keel Formation through the Prices Falls Member of the Clarita Formation. The St. Clair and Lafferty Limestones in Arkansas are correlative to the Fitzhugh Member of the Clarita Formation, and probably to the lowermost part of the Henryhouse Formation. Equivalents to the middle and upper Henryhouse, and the Haragan, Bois d'Arc, and Frisco Formations of the Oklahoma Hunton Group are not present in the Arkansas section. The stratigraphic interval of the Oklahoma Sallisaw Formation, generally included in the Hunton Group by subsurface workers, is occupied in Arkansas by the Penters Chert, a close lithic equivalent to the Sallisaw.

These Arkansas rocks and their Hunton equivalents record accumulation on a shallow-water carbonate shelf that was relatively isolated from terrigenous sources and repeatedly was exposed to subaerial weathering and erosion. The resultant unconformities are well documented, both physically and paleontologically.

GEOLOGIC SETTING

Hunton Group equivalents in northern Arkansas crop out discontinuously in a narrow east–west belt across the north-central part of the State. The major regions of outcrop, from east to west, are the Batesville district in Independence, Izard, and eastern Stone Counties, the Allison–Blanchard Spring area in central Stone County, and the Gilbert area in Searcy County (Fig. 1). The sediments represented by these rocks accumulated on a widespread, shallow-water carbonate shelf relatively isolated from terrigenous sources. Geologically, the region is on the southwestern flank of the Ozark dome. Regional dip of <1° to the south is interrupted by minor folds and down-to-the-south normal faults with displacements of a few meters to a few tens of meters. Physiographically, the outcrop belt lies along the southern margin of the Springfield Plateau, a rolling upland capped by cherty limestone of the Mississippian Boone Formation. Dissection of the plateau surface by the White and Buffalo Rivers and their tributaries has provided valleys 120–150 m deep, affording excellent natural exposures of lower Paleozoic strata along the valley walls.

The total thickness of Hunton-correlative rocks in Arkansas is slightly more than 65 m, but nowhere is this maximum preserved at a single locality. Hunton equivalents are present in the Upper Ordovician to Middle Silurian Cason Shale and the overlying Middle to Upper Silurian St. Clair and Lafferty Limestones. The depositional history of these Upper Ordovician through Silurian-age strata is nearly parallel to that of Oklahoma surface exposures of coeval Hunton Group rock reported by Amsden (1962). The uppermost Silurian and lowermost Devonian portions of the Oklahoma Hunton Group appear to be absent in Arkansas.

Because geologists working in the subsurface have traditionally included in the Hunton all strata between the Sylvan and Woodford/Chattanooga Shales, a brief discussion of the Penters Chert, which lies within the upper part of this interval, is provided in this summary. Comparison of the stratigraphic succession in Arkansas with correlative units in Oklahoma is shown in Figure 2.

Craig, W. W., 1993, Stratigraphy of the Cason Shale, the Brassfield, St. Clair, and Lafferty Limestones, and the Penters Chert (Upper Ordovician–Lower Devonian) in the Arkansas Ozarks, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 135–147.

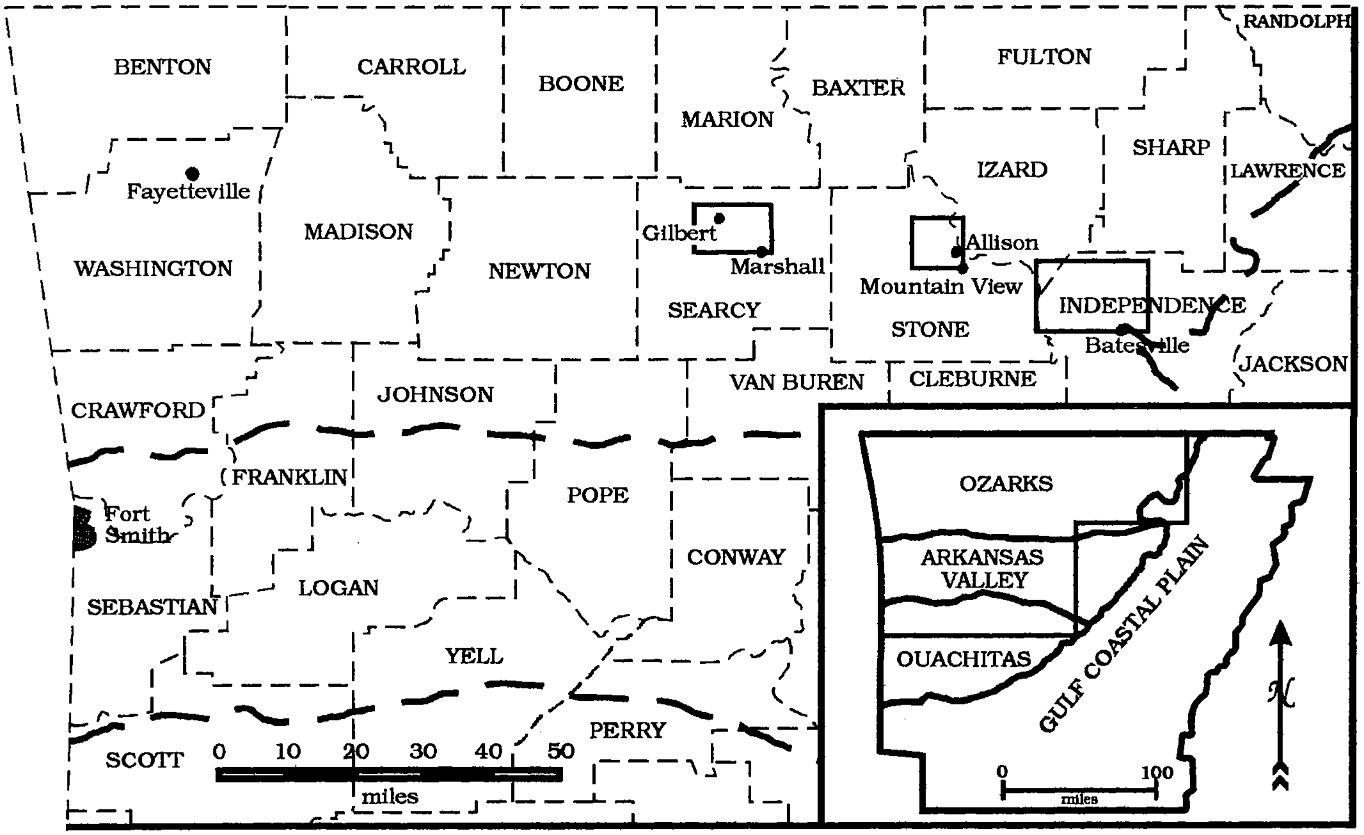


Figure 1. Major areas of outcrop of Arkansas Hunton Group equivalents.

STRATIGRAPHY

Cason Shale

Early geologists applied the name Cason Shale to a thin (<6 m thick) sequence of phosphatic sandstone and shale, in places oncolite bearing, that occurs between the Upper Ordovician Fernvale Limestone and the Middle Silurian St. Clair Limestone (Fig. 2). The oncolites, which characterize the so-called "button" shale of various authors, were identified as *Girvenella richmondensis* by Ulrich (in Miser, 1922). Accordingly, the Cason was assigned a Late Ordovician age. During the 1960s, several reports (Wise and Caplan, 1967; Amsden, 1968; Craig, 1969) showed the Cason interval to be lithologically more heterogeneous than earlier workers suspected. Geologists who have examined the interval in recent years agree, in general, that it contains the following succession of lithic types: (1) basal phosphatic, glauconitic sandstone and shale; (2) crinozoan packstone/grainstone, grading up into an intraclastic/oolitic grainstone and fenestral lime mudstone; (3) reddish skeletal packstone/grainstone, dominated by crinozoan fragments; and (4) an uppermost oncolite-bearing silty shale, with moderately common phosphate grains. The lower phosphate beds (unit 1) and "button" shale (unit 4) are the most persistent lithic types of this interval in the Batesville district, and it is with them that the concept of the Cason is most closely associated.

Earlier workers in northern Arkansas had observed the carbonates that make up units 2 and 3 (see especially Miser, 1922,1941), but in none of the exposures available to them was it clear that these units occurred between the

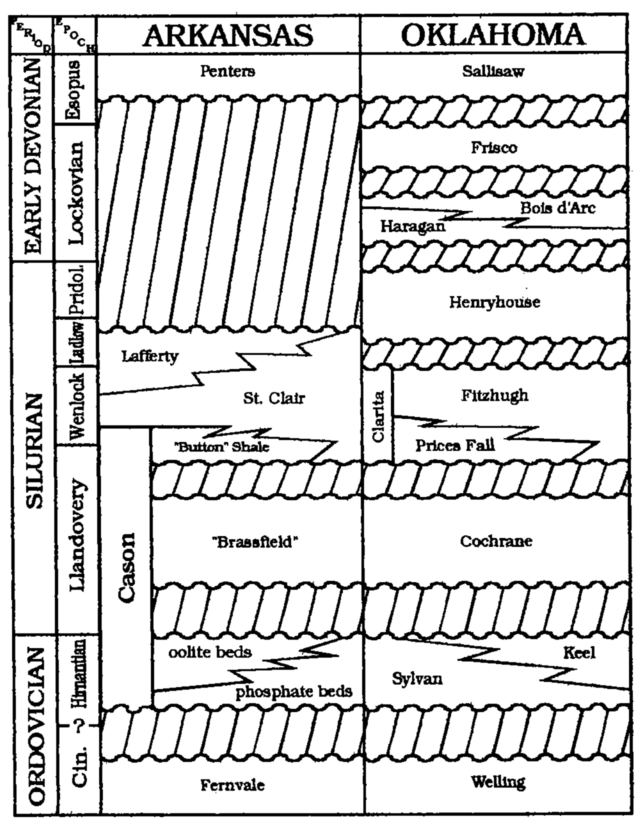


Figure 2. Correlation of Oklahoma Hunton Group rocks with equivalent formations in north-central Arkansas.

Figure 3. Photomicrograph of phosphatic conglomeratic sandstone of Cason (unit 1), Love Hollow quarry; unpolarized light. Round grains are crinozoan parts. Large grain truncated by bottom of photograph is fine-grained limestone. Bar is 0.28 mm.

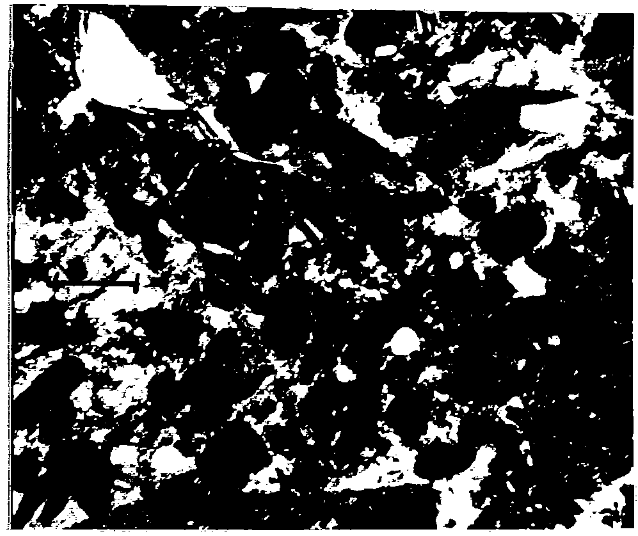


Figure 4. Same view as Figure 3; polarized light. Dark grains are phosphate. Bar is 0.28 mm.

two detrital units of the Cason. Unit 3 is the Lower Silurian Brassfield Limestone, long known in northern Arkansas but mistakenly assigned by early geologists to a stratigraphic position between the Cason and St. Clair. The complicated erosional/depositional history of the Cason has resulted in a random distribution of its individual units. This distribution, combined with a heavy manganese and iron mineralization that obscures the relationship between lithic types of this interval, has made difficult a resolution of the geology of the Cason. I have detailed elsewhere (Craig, 1969;1975a,b;1984) the history of the unraveling of Cason stratigraphy. Interested readers are referred to these sources.

Lower Phosphatic Sandstone (Unit 1) and Oolitic Limestone (Unit 2)

The conglomeratic sandstone and shale (unit 1) that comprise the basal unit of the Cason are best developed in the western part of the Batesville district (Fig. 1). The thickness of these beds ranges from 0 to 3 m, the variation appearing to result from deposition on an irregular erosional surface of the Fernvale Limestone. Characteristic grains of the conglomeratic sandstone are quartz, phosphate, and glauconite, which vary considerably in their relative abundance (Figs. 3,4). Phosphate can make up as much as 70% of total grains, and glauconite as much as 20% (Lemastus, 1979, p. 57). Thin sections reveal that the phosphate is a replacement of calcium carbonate grains. Particularly abundant among the replaced grains are crinozoan parts; but clams, snails, trilobites, brachiopods, and fragments of fine-grained limestone also are common. Fragments of phosphatic brachiopods constitute as much as 8% of the grains. The rock is very poorly sorted, with grains ranging from silt size (generally quartz) to medium-pebble size (limestone clasts). Glauconite occurs mostly as very fine to fine sand, probably of fecal origin, but it also is present as a replacement of calcium carbonate and as cement.

Residues, derived by boiling the shale in water, contain an abundance of phosphatic and glauconitic grains of the same types observed in sandstone thin sections. The most striking features of these grains are their high degree of polish and rounding. There is little doubt that they underwent abrasion prior to deposition. It also seems likely that replacement took place prior to transport; otherwise the grains would not possess the high luster they now have. The most likely source for these grains is the underlying Fernvale Limestone, and, indeed, thin sections of the upper Fernvale reveal phosphatized grains. Furthermore, heavy-liquid separations of acetic-acid residues from the Fernvale contain phosphate grains similar to those in the overlying Cason, except for their lack of polish and rounding.

Occurring stratigraphically above the phosphate beds, and apparently conformable with them, is a carbonate unit (unit 2) that has, in its base, ~30 cm of crinozoan grainstone with oolitic coatings (Fig. 5). This crinozoan rock grades upward into an oolitic grainstone (Figs. 6,7) overlain by an intraclastic/oolitic grainstone and fenestral lime mudstone (Fig. 8). One of the most remarkable features of the oolitic rock is the size of its ooids, some of which are 6 mm in diameter. This limestone is known from only two places in the Batesville district, Love Hollow quarry and St. Clair Spring (see Craig and others, 1984, for locations). In the early- and mid-1960s, ~1.5 m of oolitic packstone/grainstone were present at the quarry between unit 1, below, and the Brassfield Limestone (unit 3), above. Quarrying since that time has resulted in the disappearance of both it and the Brassfield, leaving a Cason section of "button" shale (unit 4) resting directly on the phosphatic sand and shale of unit 1. The only presently known locality to view the Cason oolitic rock in the Batesville district is St. Clair Spring, just north of Batesville. Here the phosphate beds of unit 1 are missing, and ~2.7 m of unit 2, exhibiting all of its textural variations, occur between the Fernvale and Brassfield Limestones.

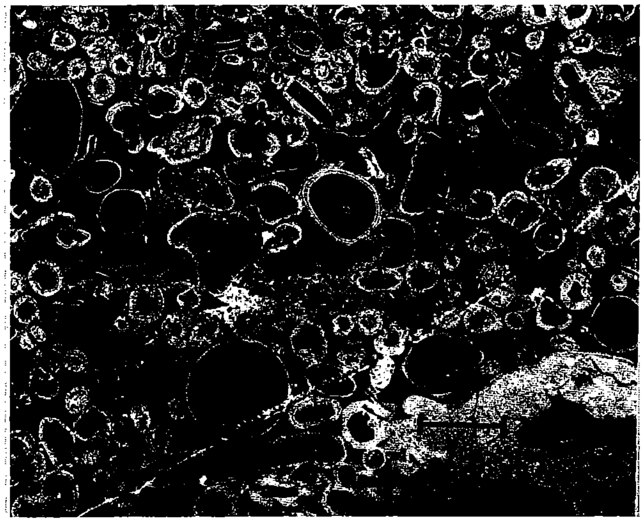
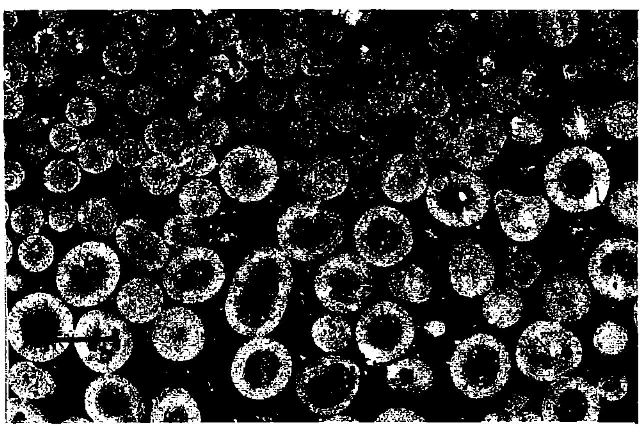


Figure 5. Negative print of thin section showing base of Cason oolitic limestone (unit 2), Love Hollow quarry. Coarse- to very coarse-grained crinozoan grainstone with superficial oolitic coatings. Grades up into an oolitic grainstone (Fig. 6). Bar is 2 mm.



Figures 6. Negative print of acetate peel showing Cason oolitic limestone (unit 3), Love Hollow quarry. Bar is 2 mm.

To the west of the Batesville district, the phosphate content of unit 1 decreases markedly. In the Allison area (Craig and others, 1984) the rock is a green, silty shale and clayey siltstone with thin phosphatic sandstone layers and disseminated phosphate grains. Wise and Caplan (1967, 1979) describe the oolitic grainstone from above the shale of unit 1 in the Blanchard Spring Recreation Area, a short distance from Allison. Farther west, in Searcy and Newton Counties (Fig. 1), the unit generally has a sandy, phosphatic base that grades up into green, silty dolomitic shale resembling the Sylvan Shale of Oklahoma (Fig. 9). Macrofossils are absent, but rare conodonts indicate a fauna similar to the oolitic limestone of unit 2 (Craig, 1986). Stringers of phosphatic and siliceous ooids in this shale are judged to be equivalent to the oolitic packstone/grainstone of the Batesville district. A notable exposure in the Gilbert area, showing the dolomitic shale of the Cason, is just downstream from the U.S. Highway 65 bridge over the Buffalo River, Searcy County (Craig, 1984,1988). Here nodules at the top of the unit are composed of silicified fossil hash with ooid stringers.

Brassfield Limestone (Unit 3)

The Brassfield presently is known from only one locality in the Batesville district, the above-cited St. Clair Spring site, where 5.8 m of the formation occurs between the oolitic rock of Cason unit 2 and the St. Clair Limestone. In the early- and mid-1960s, the Brassfield was present in the Love Hollow quarry section, resting in some places on the oolitic limestone and in others on the phosphatic sandstone and shale of unit 1. It was everywhere overlain by the "button" shale (unit 4). A number of visits to the quarry during this period provided convincing evidence that the Brassfield was preserved at this locality as an unconformity-bounded erosional remnant between the overlying "button" shale and the lower Cason units. This evidence is thoroughly discussed and illustrated in a number of previous articles (Craig, 1969,1975a,1984; Craig and others, 1984). For a somewhat different interpretation of the Cason interval at Love Hollow quarry, the reader is referred to Amsden (1968,1986). It is probable that other remnants of the Brassfield are present along the bluffs of the White River in the vicinity of Love Hollow quarry. Reports (Miser, 1941) of a manganese carbonate, ≤1 m thick, within the Cason Shale of this general region probably are records of the Brassfield between the phosphate beds and the "button" shale.

The best and thickest exposures of the Brassfield are in the vicinity of Gilbert, Searcy County (Fig. 1), where it ranges in thickness from 0 to nearly 12 m. These occurrences have been mapped by McKnight (1935), Maher and Lantz (1953), and Glick and Frezon (1965). In the absence of the "button" shale in these western exposures, the Brassfield is overlain unconformably by the St. Clair Limestone (Fig. 10).

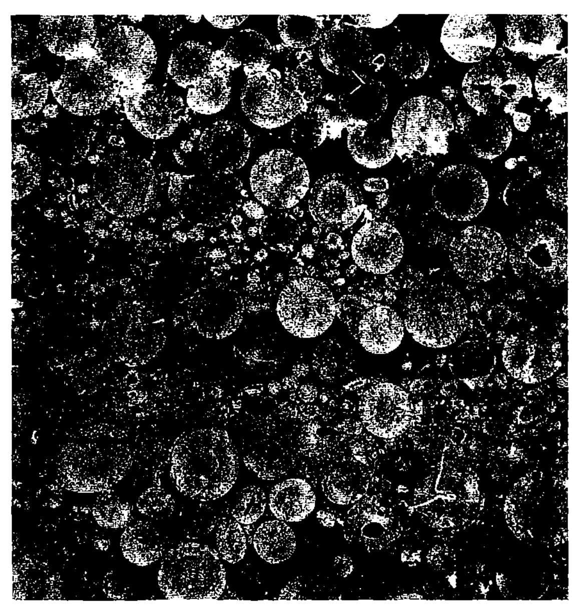


Figure 7. Negative print of acetate peel, Cason oolitic limestone (unit 3), St. Clair Spring. Bar is 2 mm.

Figure 8. Negative print of acetate peel showing Cason intraclastic-oolitic grainstone and fenestral lime mudstone (unit 2), St. Clair Spring. Note peloidal material at bottom of interstices and fenestral lime mudstone at top of photograph. Bar is 2 mm.

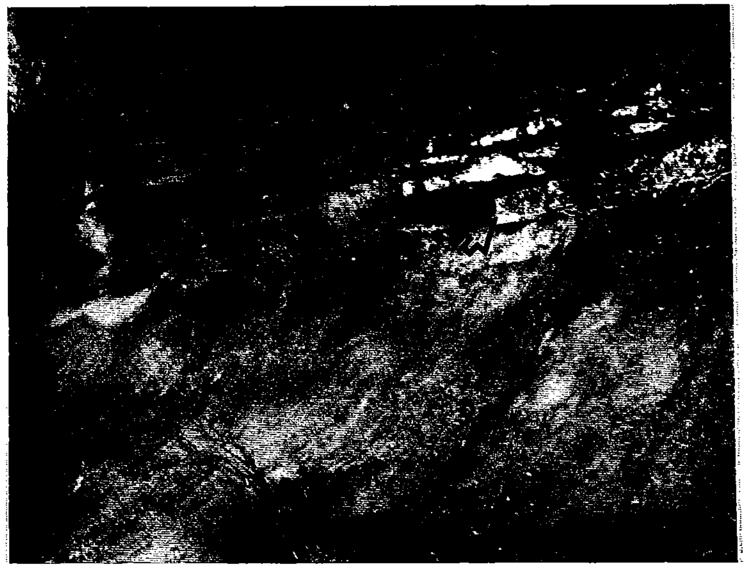


Figure 9. Boone Formation (Lower Mississippian) overlying Cason dolomitic shale of Gilbert area, in road cut along Arkansas 333 connecting Gilbert to U.S. 65. Arrow points to remnant of St. Clair, preserved between the two formations to the west (right part of outcrop) but removed by erosion to the east. Dolomitic shale of the Cason here correlates with the phosphate beds (unit 1) in the Batesville district.

The Brassfield is dominantly a bioclastic limestone of variable texture. Its chief constituents are fragmented crinozoans, but it also contains abundant debris of other common Paleozoic fossil groups. The limestone has an important constituent of lime mud, as discontinuous stringers or layers, as mud resting on the floors of interstices, and as irregular patches that resulted from burrowing. These patches, which are dark red from the admixture of detrital mud, give the Brassfield a characteristic mottled aspect. A detailed, illustrated account of the Brassfield in northern Arkansas is given by Craig (1984).

"Button" Shale (Unit 4)

The major occurrence of the oncolite-bearing "button" shale of the Cason is in the Batesville district, where it ranges in thickness from 0 to ~6 m. West of the district, it is known from only one locality within the Blanchard Spring Recreation Area (Lemastus, 1979). The most notable exposure of the unit is its type section at the old Cason (manganese) mine, where it is the only unit of the Cason present, and in the adjacent Midwest Lime Co. quarry (see Craig and others, 1984, for locations). About 4.5 m of the shale, which contains abundant oncolites flattened in the plane of bedding, is exposed in the mine face and floor. The unit thins rapidly eastward into the quarry. In the western face of the quarry, the "button" shale is in distinct erosional contact with the Fernvale, its material filling fractures and irregularities in the upper portion of limestone (Fig. 11). The shale, which is only ~60 cm thick here, grades upward into the base of the St. Clair Limestone (Fig. 12). Abundant spherical oncolites occur in the basal St. Clair.

Figure 10. Photomicrograph showing welded contact of St. Clair on Brassfield, west bank of Bear Creek along Crane Bottom, 0.5 mi southeast of Gilbert, Arkansas. Note truncation of Brassfield fossils. Scale (0.28 mm) on photograph.

Figure 11. Cason "button" shale (unit 4) filling in fractures in top of Fernvale Limestone, west wall of Midwest Lime Co. quarry, 0.5 mi north of Batesville, Arkansas. Also pictured is O. A. Wise, Arkansas Geological Commission (retired).

Figure 12. View of Cason "button" shale grading up into base of St. Clair Limestone, west wall of Midwest Lime Co. quarry. Approximate contact shown by arrow in lower part of photo. Note "buttons" (oncolites) near base of St. Clair. Camera lens cap (50 mm) in upper right for scale.

North along the quarry face the "button" shale and overlying St. Clair climb onto an expanded Fernvale section, the St. Clair eventually onlapping the shale to come to rest directly on the Fernvale (Fig. 13). This relationship shows that the "button" shale is a basal detrital phase of the St. Clair transgression. Some topography apparently existed, at least locally, on the Fernvale erosional surface. Deposition of detrital material, which was in short supply, was restricted to topographic lows. This depositional pattern explains the inconsistent occurrence of the "button" shale at the base of the St. Clair throughout north-central Arkansas. An indication of the topography present at the Midwest Lime Co. quarry is seen in the variation in the thickness of the Fernvale, which is ~23 m in the northern face of the quarry, but only 4.5–6 m thick in the western face. The topography here probably developed on the upthrown side of a fault system directly to the south. The fault was active during this time as evidenced by an angularity between the Fernvale and St. Clair.

The shale at the Cason mine is difficult to decipher petrographically, because of heavy manganese and iron mineralization. It is red to red-and-green mottled, silty to finely sandy calcareous clay shale. Dolomite rhombs occur scattered throughout. In addition to algal buttons, other fossils include fragmented ostracodes, crinozoans, trilobites, and calcareous and phosphatic brachiopods. Calcareous cement increases upward toward the St. Clair, with the upper meter being calcareous enough to dissolve in acetic acid. Acid residues produce *Ammodiscus*- and *Psammosphaera*-type agglutinated foraminifers. Conodonts characteristic of the basal St. Clair are also present and corroborate the brachiopod-based conclusion of H. S. Williams (1900), who, at the time he named the formation, concluded that the Cason at the mine contains the beginning of the St. Clair fauna.

Phosphate pebbles occur in the shale, but not in the abundance that they appear in the phosphate beds of unit 1. However, the presence of phosphate in the "button" shale provides a certain similarity between it and the basal Cason sandstone and shale that no doubt has helped obscure the distinction between the two units. The lower phosphatic beds of the Cason in the Batesville district differ from the "button" shale in that they are distinctly interbedded hard sandstone and shale, in which all fossil debris is phosphatic. They do not contain oncolites or calcareous shells, nor are they characteristically calcareous or dolomitic.

A clean exposure of the "button" shale occurs at the Love Hollow quarry. The unit at the quarry averages ~1 m thick and has a scoured contact with the rocks below (Figs. 14–16). The contact is sharp and distinct where the Brassfield is the underlying unit; it is more subtle where the Ordovician phosphate beds are the underlying unit. The basal 15 cm of the "button" shale at the quarry is dark-red, quartz-sandy siltstone with pebbles of phosphate and chert and fragmented crinozoan parts. This basal conglomeratic unit grades up into 0.6 m of red to gray-green pyritic calcareous silty shale, with scattered dolomite rhombs and abundant flattened oncolite "buttons." Also present are sparse, fragmented ostracodes, brachiopods, trilobites, crinozoans, and abundant agglutinated foraminifers.

Figure 13. Onlap of Cason "button" shale (thin dark band at base of Ssc) by St. Clair Limestone (Ssc) onto Ordovician Fernvale Limestone (Of), west wall of Midwest Lime Co. quarry. North (right) along wall the St. Clair is truncated by the Penters Chert (Dp, on dark unit), which lies directly on the Fernvale.

Figure 14. Oolitic limestone (unit 2) of lower Cason Shale, Cason "button" shale (unit 4), and St. Clair Limestone, Love Hollow quarry. Hammer rests on light-colored oolitic limestone. To the right of hammer, the dark-colored "button" shale is channeled into the limestone. Lower arrows point to contact. Gradational contact between "button" shale and St. Clair occurs at about the level of the upper arrow.

St. Clair and Lafferty Limestones

The St. Clair Limestone is the most widespread and continuous of the Hunton-equivalent units in Arkansas. The limestone is best developed in the Batesville district, where it ranges in thickness from 0 to ~30 m. It is thickest in the eastern part of the district, but occurs at only isolated localities. Its distribution is apparently controlled by preservation in structural downwarps that were protected from post-St. Clair erosion. In the western part of the district, the unit is more continuous, but it averages only ~4.5 m thick.

The St. Clair is discontinuous in the Allison–Blanchard Spring area. The best exposure of the formation is along the north bank of South Sylamore Creek, just downstream from Gaylor Crossing (see Craig and others, 1984, for location). To the west, in Searcy County, the St. Clair ranges from 0 to 10.6 m thick.

The spectrum of St. Clair textures parallels that of the Brassfield. Both are dominantly bioclastic limestones characterized by fragmented crinozoans and with abundant lime-mud matrix. The St. Clair is somewhat coarser grained than the Brassfield and lacks the Brassfield's distinctive dark-red mottling.

Vertical trends, not identified in the Brassfield, are present in the St. Clair. These trends are significant in the interpretation of the depositional history of the formation, particularly when considered in conjunction with the conformably overlying Lafferty Limestone. In the Batesville district, the St. Clair is a pinkish- to light-gray, coarse-grained crinozoan packstone/grainstone. Some layers, especially at the base and top of the unit, are poorly washed and contain a matrix of gray-green lime mud. In all sections where the base of the St. Clair is exposed, the lower 0.6–1.5 m is a pyritic, gray-green or red ostracode wackestone that contains detrital clay, angular to subrounded quartz silt, conodonts, abundant agglutinated foraminifers, large well-formed dolomite rhombs (Fig. 17), and minor amounts of fragmented crinozoans, corals, brachiopods and trilobites.

The overlying main body of crinozoan-rich St. Clair is dominated by wackestone and packstone in upper and lower layers, and by grainstone and poorly washed grainstone in middle layers. The wackestone and packstone contain abundant whole skeletons of brachiopods, ostracodes, cephalopods, and well-preserved pygidia and cephala of trilobites (Fig. 18). Holloway (1980,1981) has reported on the St. Clair trilobites and Amsden (1968) has described the brachiopods. The middle grainstone is dominated by crinozoan calcarenite and calcirudite containing fragmented skeletal material of other fossil groups (Fig. 19). This superposition is well displayed at the now-abandoned Batesville Stone quarry and the type section near St. Clair Spring (Craig and others, 1984). It is less well developed in the thinner (3.6 m) section at Love Hollow quarry.

In the western part of its outcrop, in the Gilbert area, the St. Clair contains a greater percentage of lime mud. Grainstone, although present in some layers, is not as characteristic of the unit as in the Batesville district, and layers of skeletal wackestone are common throughout the limestone.

The Lafferty Limestone was named by Miser (1922) for an exposure at Tom Tate Spring (see Craig and others, 1984, for location), in the western part of the Batesville district. Miser gave the name Lafferty to ~30 m of slabby, earthy, gray-green to red lime mudstone and sparsely fossiliferous skeletal wackestone overlying the St. Clair. Miser recognized the Lafferty only in the immediate vicinity of

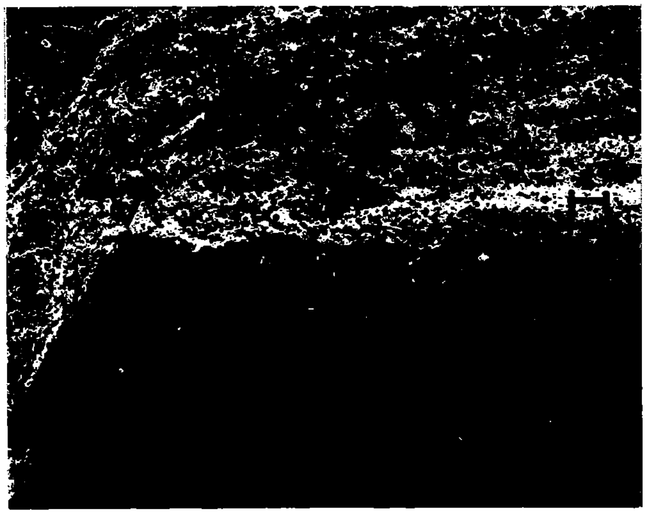


Figure 15. Negative print of thin section showing Cason "button" shale (unit 4) on Brassfield Limestone (unit 3), Love Hollow quarry. Note irregularity of contact. Arrow shows position of enlargement illustrated in Figure 16. Bar is 1 mm.

Figure 16. Photomicrograph showing enlargement of contact illustrated in Figure 15. Note truncation of crinozoan parts in Brassfield. Grains in "button" shale are quartz and phosphate in clay matrix. Bar is 0.2 mm.

its type locality. Beneath the type Lafferty, the upper 3.8 m of St. Clair Limestone, as understood by Miser, is a slightly dolomitic, pyritic, gray-green ostracode wackestone with scattered pink crinozoan fragments and other calcareous skeletal material, subrounded-quartz sand, clay, and agglutinated foraminifers. This rock type, which occurs above the coarse-grained crinozoan rock of the St. Clair at most localities, was included in the St. Clair by Miser. There is little to distinguish it from the 25.9 m of dominantly red limestone that Miser called Lafferty. Miser's Lafferty differs from the lime mudstone he included in the St. Clair only in that the Lafferty is red from the inclusion of detrital clay and contains less micrite and fewer fossils in its upper portion. In 1950, Straczek and Kinney remapped the Batesville district and included in the Lafferty all of the gray-green lime mudstone and ostracode wackestone above the coarse-grained crinozoan limestone of the St. Clair. Geologists working in northern Arkansas since that time have followed this redefinition.

The lithic description given above for the Lafferty fits the unit across northern Arkansas. Its contact with the St. Clair is placed above the highest appearance of abundant coarse-grained bioclastic allochems. The change from St. Clair to Lafferty results from a decrease in the percentage of allochems and an increase in micrite and detrital grains. The basal beds of the Lafferty contain pinkish crinozoan skeletal material as laminae and scattered grains. There is a gradual increase in detrital mud and decrease in carbonate, including fossil debris, upward in the unit. These trends are best seen in thicker sections, where the basal ostracode wackestone grades upward through a lime mudstone into a rock that approaches a calcareous claystone. With the addition of clay, the unit becomes characteristically slabby bedded.

The Lafferty everywhere rests conformably on the St. Clair, as evidenced by the transitional passage between the two units, and is overlain unconformably by younger units. There is a remarkable similarity between the Lafferty and the lime mudstone occurring within the St. Clair, especially the basal ostracode wackestone. At its type section the Lafferty is 29.7 m thick, which includes the 25.9 m that Miser named Lafferty and 3.8 m of ostracode biomicrite he originally included in the St. Clair Limestone. The formation is considerably thinner in other parts of the Batesville district. Along the bluffs of the White River, in the western part of the district, the Lafferty averages 6 m thick. In the eastern part of the district its average thickness is 1.5 m. The unusual preservation of the limestone at its type locality is probably structurally controlled.

Geologists mapping in the Gilbert area (McKnight, 1935; Maher and Lantz, 1953), although recognizing the distinction between the St. Clair and Lafferty lithologies, have dealt with the two units as a single, undifferentiated formation. The presence, in the St. Clair, of micrite beds that are lithologically similar to those in the Lafferty, and the absence of the Lafferty over most of the area, have no doubt contributed much to this way of handling the two units. It is probable that the Lafferty lithology was once present above the St. Clair throughout the Gilbert area, just as it is in outcrop areas to the east, but has been removed by pre-Boone erosion. Present evidence from some of the thicker sections in the region (Maher and Lantz, 1953; Craig, 1984) indicates that the Lafferty lithology becomes dominant about 10.5–12 m above the base of the St. Clair.

Figure 17. Negative print of thin section showing basal ostracode wackestone of St. Clair. White areas are pyrite. Bar is 1 mm.

Figure 18. Negative print of acetate peel showing poorly washed grainstone of St. Clair Limestone, Batesville Stone quarry. Note unfragmented trilobites and sparry calcite (dark grains) in protected areas. Bar is 2 mm.

Penters Chert

Miser (1922) named the Penters Chert for exposures in the vicinity of Penters Bluff, which borders the White River in the western part of the Batesville district. Based on its stratigraphic position, he assigned the unit to the Lower Devonian. Kinney (1946) reported, from the base of the Penters, a small brachiopod fauna that he interpreted as Early or Middle Devonian in age. The presence in the Penters of fragmented Pa elements of the conodont genus *Latericriodus* Müller suggests that the formation is no younger than Early Devonian. The upper part of the unit is well exposed along the railroad right-of-way just south of Penters Bluff. Most of Miser's understanding of the lithic character of the Penters came from these exposures, which are structureless beds of dense light- to bluish-gray chert with some unsilicified patches of gray fine-grained limestones. The only other places the Penters is preserved within the district is near Cushman, in the central part of the district, and just north of Batesville. During the time of Miser's work, the latter occurrence was mostly covered by float from the overlying Boone Formation. The excellent exposures at the Batesville Stone quarry and the Midwest Lime Co. quarry presently afford the best available view of the entire formation (Fig. 20).

Based on his observations along the bluffs of the White River, Miser reported the unit as 6–7.5 m of bedded chert with only a small amount of limestone. North of Batesville the unit is ~8.5 m thick, but only the upper 2 m are pervasively silicified. The lower 6–6.5 m are gray, fine-grained quartz-silty limestone and dolomitic limestone, with some chert nodules and zones of partial silicification (Fig. 21). Cursory examination of a few thin sections from the unit shows that the relative percentages of its lithic constituents vary considerably at different levels, but, except for a general upward increase in chert and detrital clay, no trends are noted. In places the unit is abundantly fossiliferous and

Figure 19. Negative print of acetate peel showing crinozoan grainstone of middle St. Clair Limestone, Batesville Stone quarry. Note absence of lime mud. Bar is 2 mm.

almost pure limestone. Dominant fossils are tentaculitids, brachiopods, trilobites, and ostracodes (Fig. 22). Shells are disarticulated and fragmented, indicating post-mortem transportation. Dolomite, as well-formed finely crystalline rhombs (Fig. 23), comprises as much as 12% of the rock, and angular quartz silt comprises as much as 5%. Glauconite is particularly common in the lower few centimeters of the unit, occurring in patches and stringers that might represent feces trails of infauna. Fine grains of pyrite occur throughout the unit in patches and stringers, as well as in disseminated grains.

The Penters exhibits an irregular contact with the underlying Lafferty at the Batesville Stone quarry and in the west wall of the Midwest Lime Co. quarry. In the northern and eastern walls of the Midwest Lime Co. quarry, the Penters rests directly on the Fernvale because of pre-Penters erosion of the St. Clair–Lafferty. The base of the Penters is marked by a 5-cm-thick glauconitic, dolomitic sandy shale. A thin chert breccia (2.5 cm thick) occurs at the top of unit and apparently represents residuum on the post-Penters erosional surface. Miser reported 1.8–2 m of chert breccia in the upper part of the Penters along the White River. The breccia, which is not sharply separated from bedded chert below, contains large blocks and even, unbroken beds tilted at angles as high as 40°. Miser interpreted this breccia to result from subsidence of the Penters into sinks formed in the underlying St. Clair during subaerial exposure. Inasmuch as the overlying Chattanooga and Boone are not affected, this karst development must have occurred prior to their deposition.

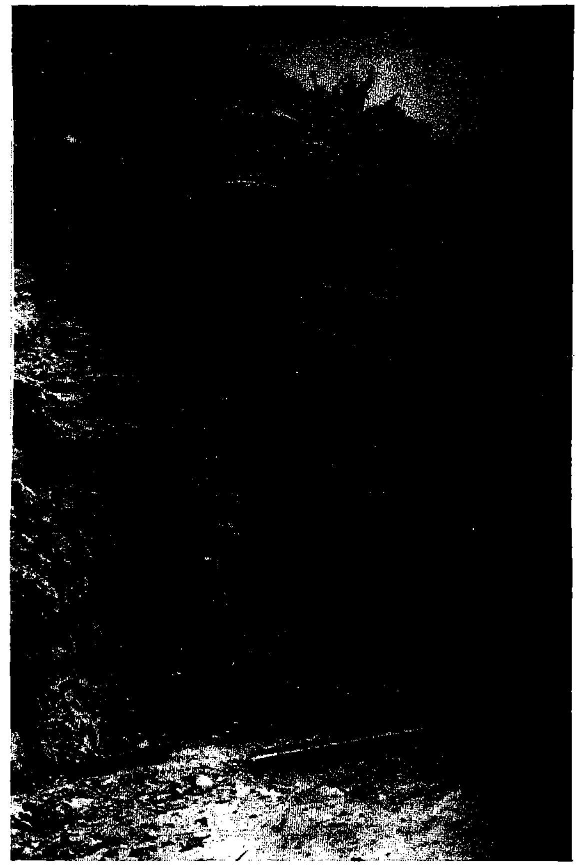


Figure 20. Penters Chert at Batesville Stone quarry. Arrow (on right) on contact between Penters and underlying Lafferty Limestone.

INTERPRETATION

Comparison to Hunton Group of Oklahoma

The stratigraphy and paleontology of the Hunton Group in Oklahoma have been studied extensively by T. W. Amsden, of the Oklahoma Geological Survey, and reported in a series of publications too numerous to mention here. Citations to the most important of these reports are in Amsden (1968,1986), two articles that also include important observations on Hunton equivalents in Arkansas. With minor exceptions, Amsden's synthesis of Upper Ordovician and Silurian stratigraphy of north Arkansas and its comparison to that of Oklahoma agrees with interpretations presented here (Fig. 2).

The lithic succession of the Cason oolitic limestone at St. Clair Spring in the Batesville district (Lemastus, 1979; Craig and others, 1984)—basal oolitic/skeletal grainstone; oolitic grainstone; intraclastic/oolitic grainstone and fenestral lime mudstone; and upper oolitic grainstone—bears a remarkable resemblance to the best-developed sections of the Keel Formation illustrated by Amsden (1986, p. 15–16). The basal oolitic/skeletal grainstone of the St. Clair Spring sequence is equivalent to the Ideal Quarry Member of the Keel. The Oklahoma equivalent of the phosphate beds, which conformably underlie the oolite in the Batesville district, is problematic. If it is true, as suggested earlier, that the ooid-bearing, greenish-gray dolomitic shale in the stratigraphic position of the Cason in the Gilbert area is coeval with the Batesville phosphate beds; and if it is further true that this shale is an eastward extension of the Oklahoma Sylvan Shale, which it resembles so closely; then it is probable that the basal Cason rocks in Arkansas have their Oklahoma correlatives in the Sylvan. Amsden (1986, p. 6) finds no evidence of an unconformity separating the Keel and Sylvan and hints (p. 9) that the two units could be lateral equivalents. This relationship would suggest that the Sylvan–Keel couplet and the phosphate beds–oolite couplet of the lower Cason are approximately the same age. It is not clear just what age.

The Sylvan traditionally has been considered as Late Ordovician (Maysvillian–Richmondian) in age, based on the occurrence of the graptolite species *Dicellograptus complanatus* (Decker, 1935). Working with what appears to be a correlative unit in Missouri, the Noix Limestone, McCracken and Barnes (1982) concluded that its conodont fauna, similar to that produced by the Keel and Cason oolite (as well as the greenish-gray shale in the Gilbert area) is no younger than Richmondian (Late, but not latest, Ordovician). On the other hand, recent paleontologic studies by Amsden (1986), Barrick (1986), and McAuley and Elias (1990) agree that the greater thickness of the Keel and Cason oolite is latest Ordovician (post-Richmondian) in age. In the construction of Figure 2, I have accepted the McCracken and Barnes (1982) analysis for convenience. The major point made by Figure 2 is the physical relationship of the oolite to the underlying shale in Oklahoma and Arkansas, regardless of what age these units might ultimately prove to be.

Amsden (1968, p. 6) was the first to report that the correct stratigraphic position of the Brassfield Limestone of the Batesville district is within the Cason Shale interval, and not between the Cason and St. Clair. The Brassfield is clearly correlative with the Cochrane Limestone; the two formations are similar in lithology and fauna (Amsden, 1968, p. 6; Barrick, 1986, p. 67). Like the Cochrane, the Brassfield has unconformable relationships with subjacent and superjacent units (see Amsden, 1986, p. 6, 17; 1968, p. 15, for Cochrane relationships). Both conodonts (Barrick,

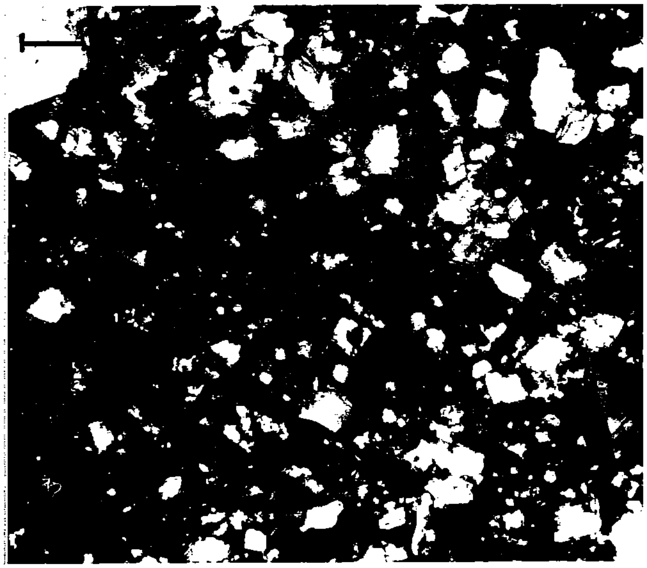


Figure 21. Photomicrograph of silicified limestone in Penters Chert, Batesville Stone quarry. Light grains are dolomite in a dark matrix of finely divided silica (chert). Bar is 0.2 mm.

Figure 22. Photomicrograph of fossiliferous Penters Chert, Batesville Stone quarry. Note transverse and longitudinal sections of tentaculitids. Bar is 0.2 mm.

1986; Martinez, 1986; Craig, 1986) and brachiopods (Amsden, 1968,1986) date the Brassfield as Early, but probably not earliest, Silurian.

The "button" shale is a basal detrital phase of the St. Clair Limestone, and it rests unconformably on all underlying units down to, and including, the Fernvale. Its contact with the overlying St. Clair is gradational. The equivalent in the Oklahoma section, the Prices Falls Member of the Clarita Formation, has similar stratigraphic relationships. Conodonts date the "button" shale (Craig, 1986) and Prices Falls (Barrick and Klapper, 1976) as late Llandoverian to early Wenlockian (late Early to early Middle Silurian) in age.

The St. Clair–Lafferty sequence in the Batesville district is, for the most part, equivalent to the Fitzhugh Member of the Clarita Formation, and possibly a few meters of the lower part of the Henryhouse Formation. Within the sequence there is no evidence of an unconformity comparable to that which exists between the Fitzhugh and Henryhouse (Amsden, 1968, p. 16). Lithologically, the interbedded lime wackestone and skeletal packstone/grainstone of the St. Clair–Lafferty agree well with the Fitzhugh. A similarity between these Oklahoma and Arkansas rocks is an upward increase in detrital mud, a feature exhibited strikingly by the Lafferty. It is puzzling that, unlike the Lafferty (the upper marlstone of which is essentially devoid of fossils), both microfauna and macrofauna are fairly common in the muddy Fitzhugh. The faunas of the two limestones place them in the Wenlockian Stage (Middle Silurian) (Barrick and Klapper, 1976; Amsden, 1968), with the possibility that the St. Clair–Lafferty may range as high as Ludlovian (Late, but not latest, Silurian).

The Upper Silurian–Lower Devonian stratigraphic succession in north-central Arkansas appears to have nothing correlative with the greater part of the outcropping Henryhouse Formation and the Haragan, Bois d'Arc, and Frisco Formations of Oklahoma. If strata correlative with these Oklahoma units were ever present in the current outcrop areas of Arkansas, it has been removed by pre-Penters and/or pre-Boone erosion. Figure 2 shows the Penters to be approximately the same age as the Sallisaw Formation, although there is no conclusive paleontologic evidence to support this. This matching follows the most recent comprehensive attempt at the regional correlation of these strata (Haley and Stone, 1986). The silty, siliceous dolomitic limestone of the Penters agrees closely with the lithic character of the Sallisaw, as described by Amsden (1961). Both the Penters and Sallisaw are lithologically distinct from underlying units, which for the most part are repetitions of compositions and textures similar to the St. Clair and Lafferty of Arkansas and the Clarita of Oklahoma.

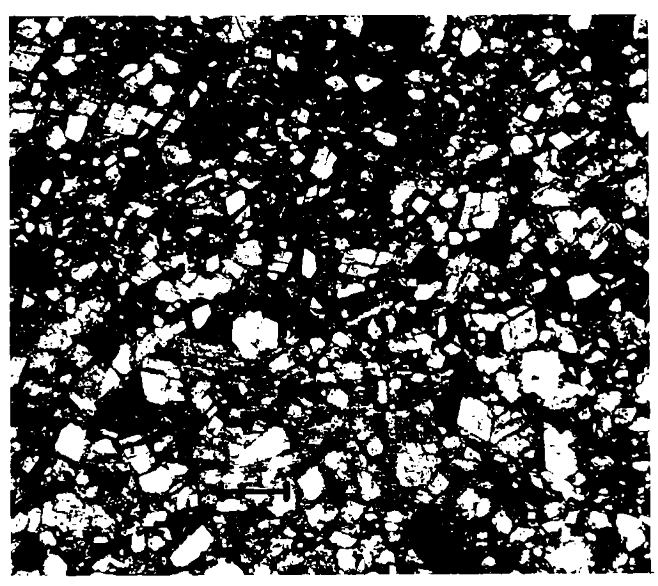


Figure 23. Photomicrograph of dolomitic limestone in Penters Chert, Batesville Stone quarry. Bar is 2 mm.

Depositional History

The Cason oolitic grainstone and its associated carbonate facies represent development of shoal-water conditions, in places possibly intertidal, following a transgression recorded by the underlying lag concentrate of the Cason phosphate beds. As Amsden (1986, p. 9) suggests, the irregular distribution of the oolitic rock probably results both from the localized development of oolitic shoals in these Late Ordovician seas and from removal during post-oolite erosion. The extent of the area of oolite formation is not clear from its poor preservation in the surface exposures of Arkansas, but Amsden (1986, p. 43) reports disjunct occurrences of the Keel in Oklahoma over an area of ~45,000 km^2. Amsden (1986, p. 47) suggests that this shoaling might have resulted from a eustatic lowering of sea level, caused by the well-documented Late Ordovician–Early Silurian glaciation in North Africa.

Present evidence suggests that the Brassfield Limestone is a transgressive unit that records a subtidal complex of wave-exposed shoal and wave-protected quiet subenvironments, all generally shallow water abounding in life. In the Gilbert area, which is the major region of Brassfield preservation, the limestone rests with irregular contact on the Cason. The upper layers of the Cason in this area are dolomitic mudstones that possess few grains that could contribute to a detrital lag deposit, such as that which occurs in the base of the overlying St. Clair Limestone.

Prior to deposition of the overlying Cason "button shale" and St. Clair Limestone, the region underwent broad epeirogenic uplift that exposed the surface to subaerial erosion. In the Batesville district, where the "button shale" is the basal unit of the succeeding transgression, the unconformity on top of the Brassfield is apparent. In the Gilbert area, lack of a basal detrital phase has created a subtle contact, with St. Clair directly on Brassfield. Truncation of Brassfield fossils at "welded" unconformities (Fig. 10) seems to be an expression of a planar erosional surface developed in the absence of much terrestrial vegetation. The irregular distribution of the Brassfield probably represents preservation of the unit in structural lows, and complete removal of it on the highs created by the gentle epeirogenic movement.

The St. Clair Limestone, with the Cason "button" shale as a basal detrital phase, transgressed over an erosional surface developed on the Brassfield and older units. In the Batesville district, particularly the eastern part, some relief existed on the surface, as evidenced in the previously discussed relationship exhibited at the Midwest Lime Co. quarry. Irregular topography does not seem to be characteristic of the surface over most of northern Arkansas. At the Midwest Lime Co. quarry, topographic relief is probably related to the Batesville graben directly to the south. Fractures in the Fernvale that are filled with "button" shale and St. Clair material, and fractures in the St. Clair that are filled with Devonian sediment, indicate that the structure was active during the Early Paleozoic. The St. Clair and "button" shale appear to have collected on the flanks of paleohighs, pinching out altogether on the crests, thus explaining the rapid disappearance of the two units. Throughout the district, detrital material making up the "button shale" was limited in quantity, no doubt derived locally from the underlying phosphate beds of the lower Cason, and thus it has a discontinuous distribution. In the Blanchard Spring and Gilbert areas, where the lower Cason has a much smaller percentage of sand and granule-size grains, no detrital phase collected at the base of the St. Clair.

The "button" shale is the basal unit of a transgressive/regressive cycle of sedimentation that includes the overlying St. Clair and Lafferty Limestones. The shale represents terrigenous sediment that accumulated in a nearshore environment containing spherical algal growths, agglutinated foraminifers, and ostracodes. Seaward, lime mud was collecting in a lagoonal environment that abounded with agglutinated foraminifers and ostracodes. This lagoonal environment produced the ostracode wackestone present at the base of most St. Clair sections. The lagoonal environment was restricted from the open ocean by a crinozoan sand bar, now represented by the coarse-grained St. Clair. On the lagoon side of this bar, marine invertebrate life abounded on a substrate composed of intermixed lime mud and skeletal debris. Currents in this area were strong enough to oxygenate the water and remove some of the finer material, but not strong enough to shift the substrate. Thus, finer particles tended to collect beneath shells that were concave downward, or in the protected hollow of shells that were concave upward. The fossiliferous, poorly washed crinozoan grainstone in the lower part of the St. Clair was deposited in this environment. Above this, the St. Clair is composed almost entirely of crinozoan debris with only minor lime-mud matrix. This lithology is interpreted as representing the ocean-facing, high-energy portion of the bar. The poorly washed upper layers of the St. Clair represent recurrence, during the regressive phase, of the conditions on the lagoon side of the bar. The ostracode biomicrite of the Lafferty marks the regression of the lagoonal environment.

The vertical succession of lithic types illustrating the depositional environments now represented by the "button" shale, St. Clair, and Lafferty is best seen in the Batesville district. The thick buildup of coarse-grained St. Clair in the eastern part of the district indicates close proximity to the major buildup of crinozoan sand that formed the protective barrier. The western part of the district was farther removed from the barrier, as demonstrated by the thin interval of coarse-grained crinozoan rock followed by a relatively thick Lafferty section.

In the Gilbert area, most of the accumulation of the St. Clair seems to have taken place on the lagoon side of the barrier. Occasional incursions of the barrier and lagoonal facies are represented by layers of crinozoan grainstone and skeletal wackestone, respectively, that are interbedded with the main body of poorly washed skeletal grainstone.

ACKNOWLEDGMENTS

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Penters Formation Paleokarst in the Arkoma Basin and the Black Warrior Basin

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ABSTRACT.—The Penters Chert is the uppermost formation in the Hunton Group of the Arkoma basin, and it is equivalent to an unnamed Devonian chert in the Black Warrior basin. These units are part of a regional Early- and Middle-Devonian chert accumulation. In the Arkoma basin, the Penters is a significant gas producer; in the Black Warrior basin, sporadic production occurs from the Devonian chert.

Cores, outcrops, and cuttings were analyzed to determine the depositional nature, stratigraphy, and diagenetic history of the Devonian cherts. Although diagenesis has greatly affected the chert, and in some areas has obliterated primary textures, petrographic examination indicates an abundance of normal-marine fossils (including sponge spicules). Subsurface mapping of the Devonian indicates a bank-like geometry. In both basins the chert was deposited in a shelf environment, with the thicker accumulations indicating shelf-edge deposition.

After deposition of the Devonian cherts, the pre-Woodford unconformity was formed. During this time the Hunton was subaerially exposed, and, in the Arkoma and Black Warrior basins, dissolution and remobilization of the biogenic silica led to karstification of the Devonian units. The karstic conditions are indicated by collapse breccias, internal sediments, and fractures, all of which are common in cores and outcrops.

INTRODUCTION

The Devonian chert of the Arkoma and Black Warrior basins is part of a regional chert accumulation across the southern North American Continent. In the Arkoma basin these strata are called the Penters Chert in the subsurface and the Sallisaw Formation on outcrop (Fig. 1). In the Black Warrior basin the Devonian chert is an unnamed formation. The Arkansas Novaculite and the Caballos Formation are siliceous equivalents found in the Ouachita and Marathon region, respectively. Other shelfward equivalents of the novaculites are the Thirty-one Formation of the Permian basin, Camden Chert of Tennessee, and Clear Creek Chert of Illinois. The Devonian chert of the Black Warrior and Arkoma basins has a greater affinity towards its shelf equivalents to the west, east, and north than to Ouachita equivalents to the south.

From 1985 to 1992, a series of regional studies were conducted by MASERA Corp. to evaluate the oil and gas potential of Devonian cherts in the Arkoma and Black Warrior basins (Figs. 2,3). One of the primary purposes of this project was to evaluate the source and diagenesis of the cherts, and their relationship to porosity and fracture development. Whereas there has been little published on the shelf equivalents, a large amount of data exists on the Arkansas and Caballos novaculites. The source of silica and the depositional environment for the novaculites and cherts in general have been debated for years. A biogenic source for the silica has been accepted by most workers; other theories for the silica source advocate alteration of volcanic ash and volcanism, promoting growth of siliceous organisms and the production of siliceous sediments; however, the evidence of volcanism in the Ouachitas is scarce (limited to several tuff beds in the Mississippian). General agreement on the depositional environment of the novaculites has not been reached, with Folk and McBride (1976), McBride and Folk (1977), Sholes (1978), Thomas (1988), McBride (1989), interpreting the novaculites to represent deep-water depositional environments. Conversely, a shallow-water depositional model has been proposed by Lowe (1975,1989), Folk and McBride (1976), and McBride and Folk (1977).

By analogy to the underlying succession of shallow-water marine limestones of the Arkoma and Black Warrior basins, the Devonian Chert is also interpreted to be a normal-marine deposit. A normal-marine fauna typically is present in the Devonian chert, and it includes sponges, pelmatozoans, brachiopods, and bryozoans. Thomas (1988) suggested that there may be an intermediate shaly facies separating shelf-like Devonian chert in the Black Warrior basin from the more basinward chert/novaculite facies to the southwest.

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Age			East of Trans-Continental Arch			
			Arbuckle Mts/ Anadarko Basin/ Eastern Oklahoma	Black Warrior Basin	West Texas Basin	Illinois Basin
Devonian	Middle	Pre-Onondagan Onesquethawan (Emsian)	Sallisaw, Camden Sdy; Penters Remnant Dol-Lst	Devonian Chert	Lt- Colored Cherty Lst	Clear Creek
			Frisco			
	Lower/Early	Oriskannian (Siegenian) Helderbergian (Gedinnian)	Bois d' Arc	------	Foss Lst	Backbone/ Grassy Knob
			Haragan		Dark Cherty Lst	
						Bailey

Figure 1. Stratigraphic correlation chart for Lower and Middle Devonian.

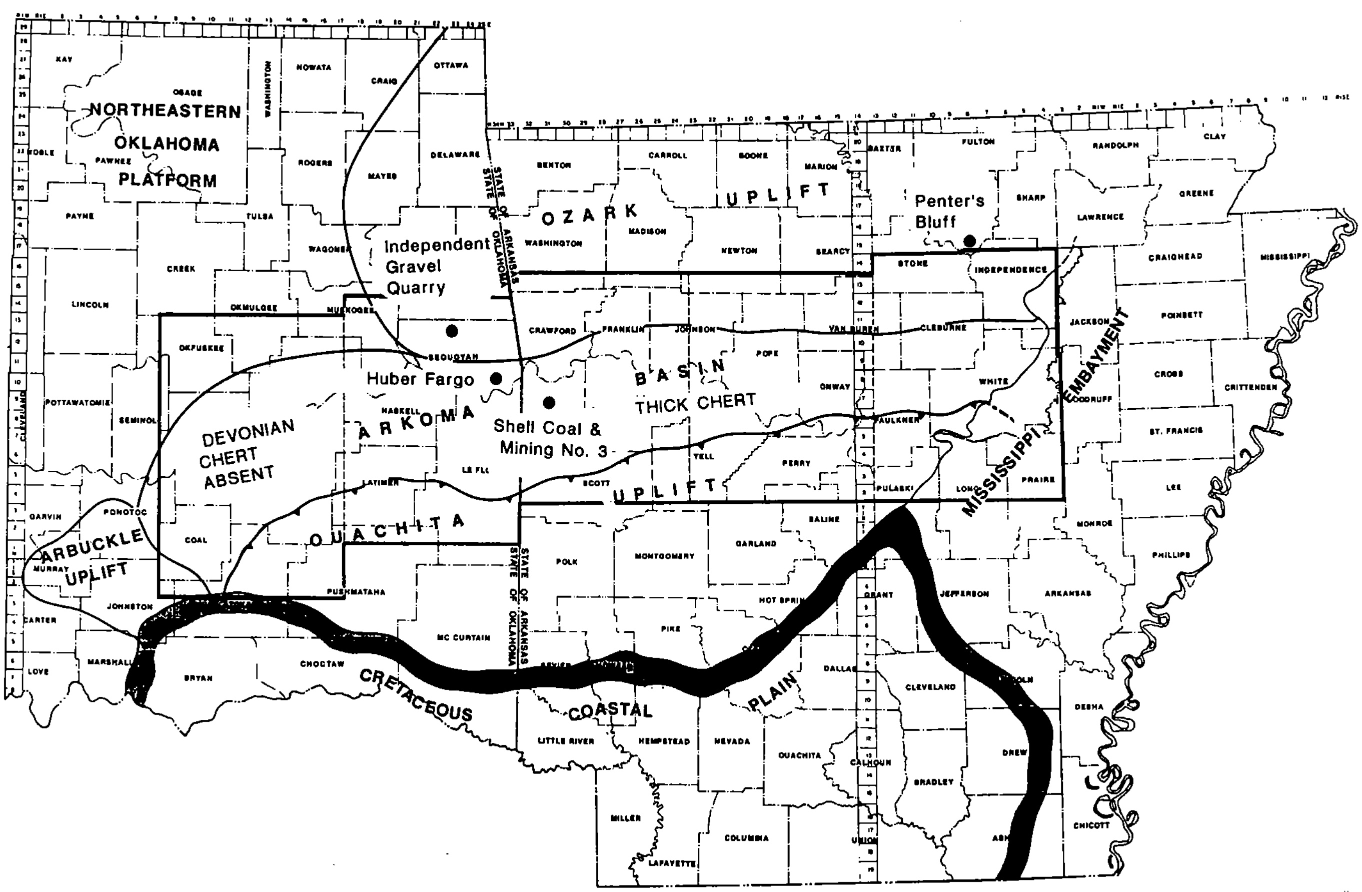


Figure 2. Map showing Arkoma basin and nearby structural features; also shown are location of cores and outcrops of the Penters Formation.

STRATIGRAPHIC DISTRIBUTION AND GEOMETRY OF THE DEVONIAN CHERT

The Devonian chert (Penters Formation) is thin or absent in the western Arkoma basin (Oklahoma portion); however, it thickens to >250 ft in central Arkansas, where the thick chert parallels the Ouachita structural front (Fig. 2). Subcrop limits in eastern Oklahoma and northern Arkansas are erosional, but there is the possibility of nondeposition. Southern limits are inferred in western Arkansas, but in central Arkansas there is no indication of a southern subcrop limit.

Portions of the Devonian, especially in the Black Warrior basin, exhibit some stratification of chert and carbonate. In the northeastern portion of the Black Warrior basin, the Devonian can be subdivided into an upper unit composed of limestone with minor amounts of chert, and a lower unit composed mostly of silica (Fig. 4). However, where the Devonian is thickest in the Black Warrior basin and throughout the Arkoma basin, the chert and carbonate occur at several stratigraphic intervals (Fig. 5). It is common for all the carbonates present in the Devonian to contain varying amounts of silica, ranging from highly siliceous to slightly siliceous.

In the Black Warrior basin, the subsurface geometry of the Devonian chert resembles a marine bank. A thick interval of the Devonian parallels the northwest–southeast Ouachita structural trend, in a manner similar to the Penters in the Arkoma basin (Fig. 3). This trend begins approximately in northern Greene County, Alabama, and continues to the northwest into central Mississippi. The northwestern limit is not well constrained, due to the lack of subsurface information. To the southwest in the Black Warrior basin, the Devonian chert thins toward the basin; to the northeast, it thins and is missing (possibly pre-Chattanooga/Woodford erosion has removed the Devonian from this area).

Some of the variation in thickness of the Devonian chert is related to pre-Chattanooga/Woodford erosion and the development of karst conditions. Typically, near the northeastern subcrop limit in the Black Warrior basin, the percent of silica is 100%, indicating that the uppermost carbonate unit has been removed, presumably by erosion.

CORRELATION

The Devonian chert has a distinctive log profile and generally is quite easy to correlate in the subsurface (Fig. 6). The log response is directly tied to the lithologic characteristics of the Devonian chert, which, in turn, are related to diagenesis (in particular to karstification and silicification). Large amounts of replacive chert and detrital infill can give a distinctive "dirty carbonate" gamma-ray response, which contrasts sharply with the underlying carbonates. In

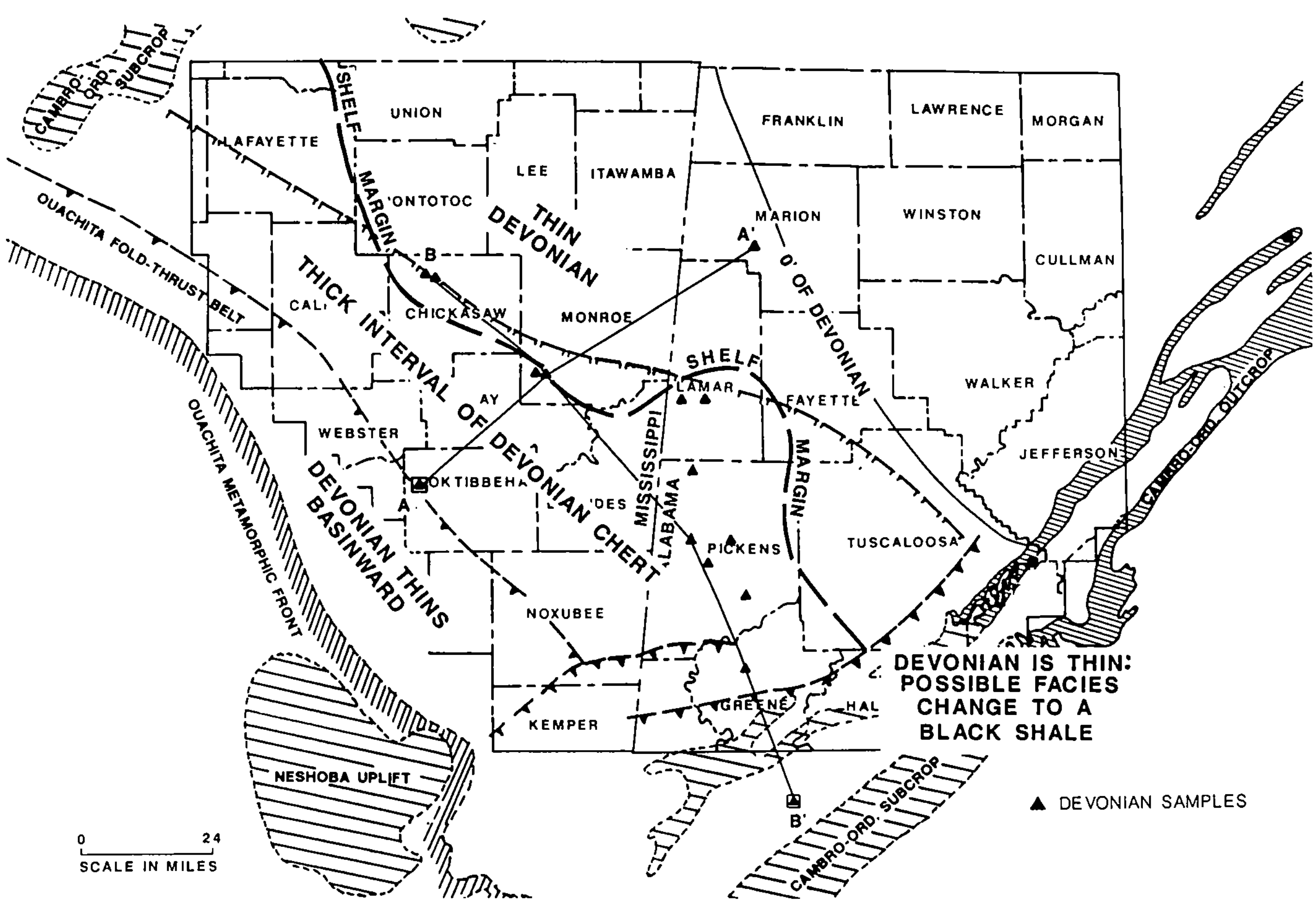


Figure 3. Index map for the Black Warrior basin showing structural features, Devonian sample locations, and lines of cross sections A–A′ and B–B′, shown in Figures 4 and 5.

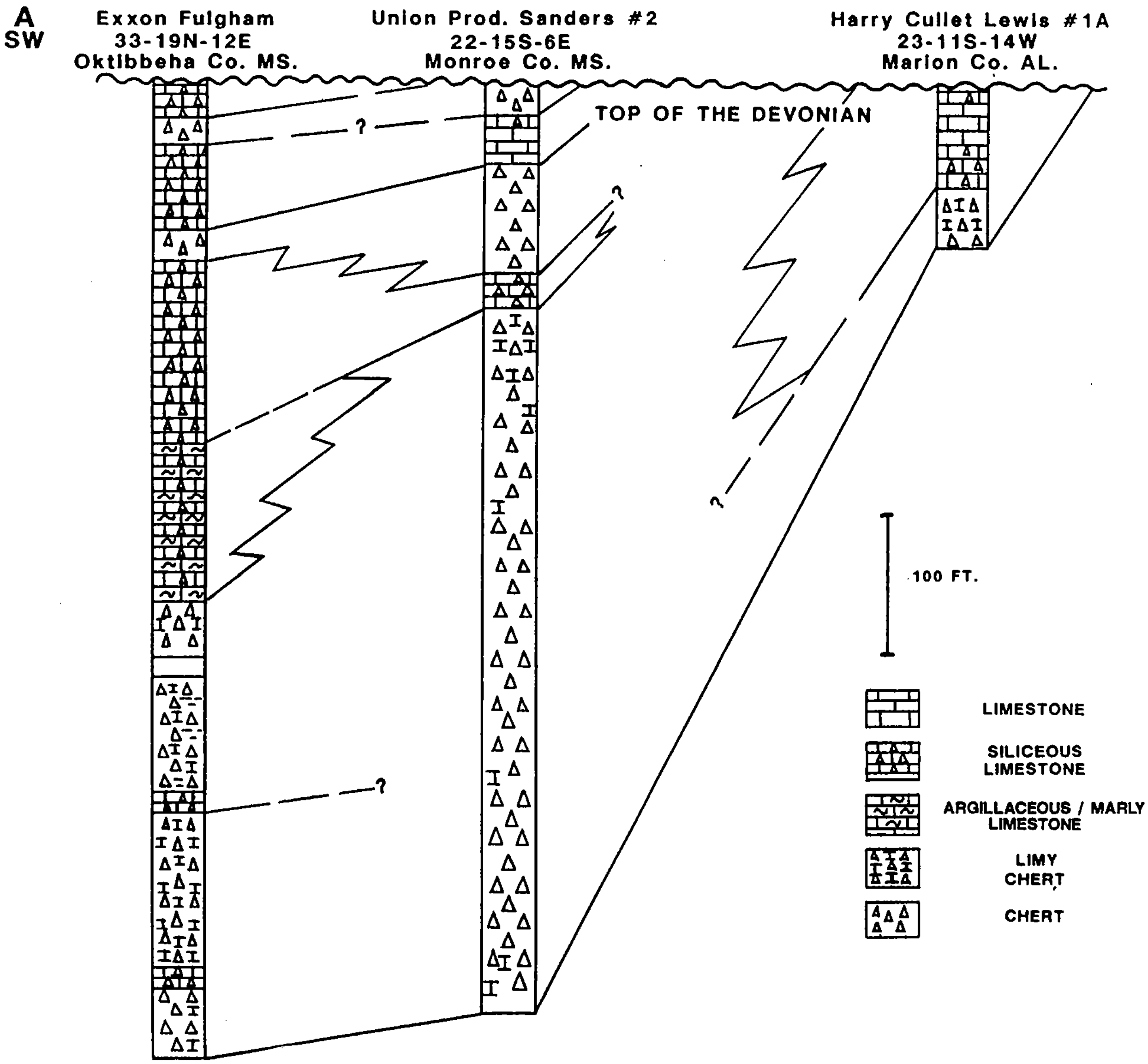


Figure 4. Dip cross section A–A′ through the Devonian chert of the Black Warrior basin showing an upper limestone section and a lower chert section where the Devonian is thin (on right). Where the Devonian is thick there are several intervals of chert and carbonate.

the Arkoma basin, the Penters Chert can be further subdivided into several zones, based on shale markers.

PETROGRAPHY

The Devonian is composed of carbonate and chert, with some intervals represented by a chert breccia that consists of angular to subrounded fragments and blocks of chert in a fine-grained matrix of clay-, silt-, and sand-sized carbonate and chert (microcrystalline quartz).

Commonly, fossil fragments have been completely silicified or dolomitized, and are difficult to recognize. In the more limy portions of the Devonian, fossil fragments are readily recognizable and appear to represent a normal-marine fauna that consists of pelmatozoans, brachiopods, mollusks, and sponge spicules. Radiolarian fragments were identified, but are rare. An important aspect of the matrix is that it has been extensively dolomitized. Dolomite commonly occurs as euhedral rhombs encased in chert, indicating formation during early diagenesis. Intermediate- and later-stage dolomites are also present, and generally they act as cements for the breccias. Figure 7 represents the typical sequence of diagenesis for the Devonian chert; however, variations do occur.

Although chert is the major constituent of the silicified breccia, chalcedony occurs as clast coatings, as infillings or linings of intraparticle porosity, and as sporadic patches within chert fragments. Some fractures are open in the extensively fractured breccia, whereas others are filled with chalcedony or dolomite.

Selected Core Descriptions

Huber No. 1 Fargo

This well is located in sec. 4, T. 10 N., R. 24 E., in Sequoyah County, Oklahoma. Penters Chert and Frisco Limestone were cored from 3,092 to 3,121 ft—log (3,090–3,119 ft—core).

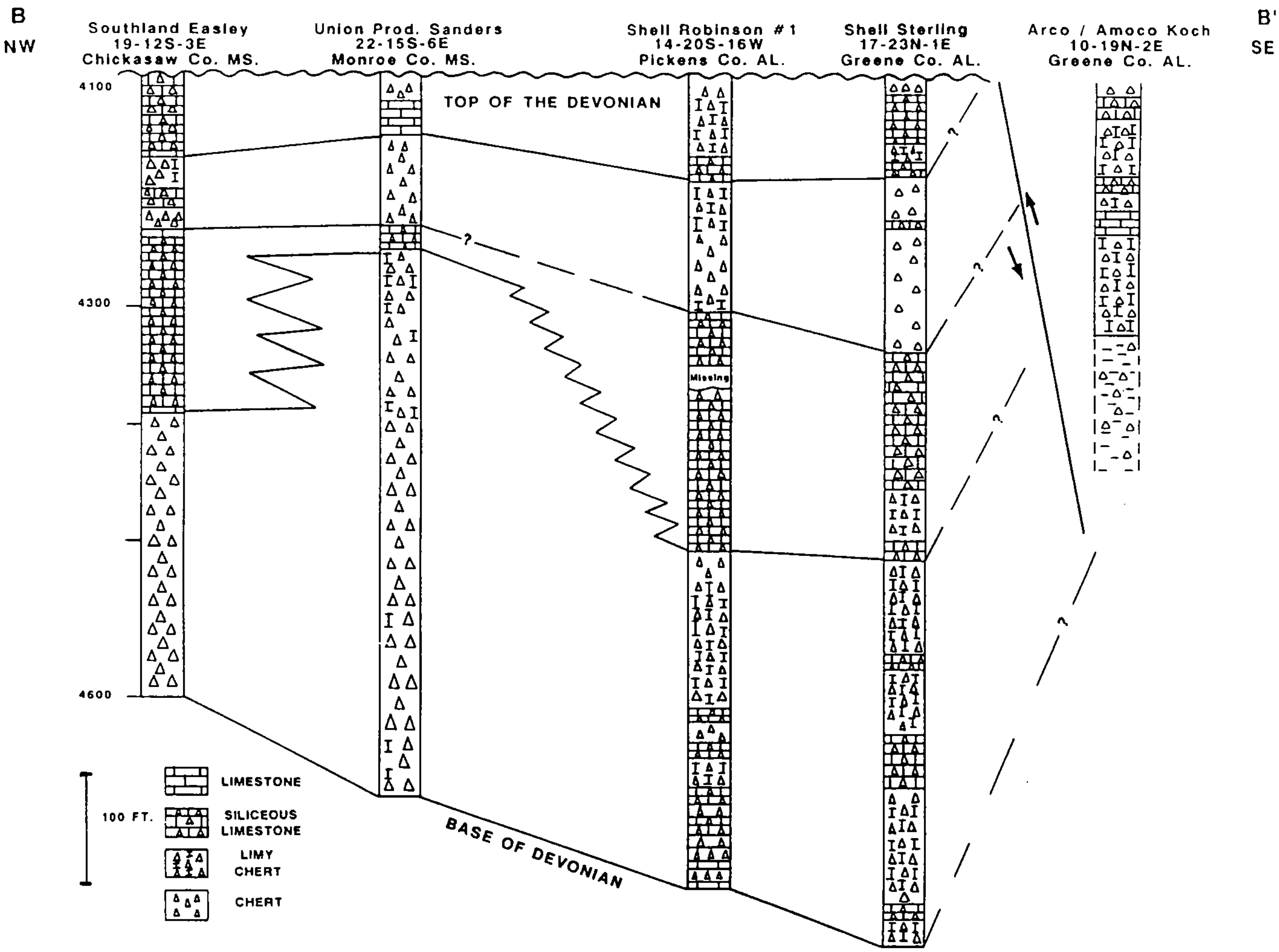


Figure 5. Strike cross-section B–B′ through the Devonian chert of the Black Warrior Basin showing that where the Devonian is thick it consists of several intervals of chert and carbonate.

The Penters Chert (3,092–3,104 ft) is a collapse-breccia complex, which is stylolitized and contains steeply inclined stratification (Fig. 8). Major constituents are chert, dolomite, and sand grains. Chert replaced the precursor wackestone or packstone during karstification, but only after a minor dolomitization episode, as is evident by small rhombs encased in chert. Larger (and occasionally zoned) dolomite occurs, along with sand grains, between chert clasts. A second dolomitization episode probably occurred during and after collapse. Inclined strata formed due to collapse and infill of sediments. Two stages of fracturing are indicated by cross-cutting relationships; one occurring before collapse, the other after collapse. Second-stage fractures are more likely to remain open. A late stage of silicification is seen as pore-lining megaquartz and chalcedony.

Western Coal and Mining Co. No. 3

This well is located in sec. 31, T. 7 N., R. 31 W., in Sebastian County, Arkansas. Penters Chert was cored at 7,748–7,751 ft.

At this site, the Penters Chert is a collapse breccia with steeply dipping strata (Fig. 9). Constituents include chert, dolomite, and sand grains, in order of decreasing abundance. The chert appears to have replaced pre-existing limestone. The majority of the dolomite and sand grains appear to have been formed, or introduced, during karstification and after silicification of the precusor rock. The rock is heavily fractured, with fractures tending to remain open in the chert portion of the rock. Chalcedony and megaquartz that line and fill pores are late-stage events.

Greg Drilling No.1 Gilmore Puckett

This well is located in sec. 18, T. 9 S., R. 9 E., Itawamba County, Mississippi. Tuscumbia/Ft. Payne, Chattanooga(?), and Devonian strata were cored at 1,586–1,596, and 1,625–1,645 ft.

At 1,594 ft, there is an undulatory contact between the Devonian chert and a black shale which is interpreted to be the Chattanooga Shale (Fig. 10A). The contact appears to be erosional, with the black shale having small clasts of chert incorporated in it. The chert is present in two cored intervals (Fig. 10B,C; 1,594–1,596 and 1,625–1,645 ft).

The upper interval of chert (1,594–1,596 ft) is gray and somewhat dense (non-tripolitic). Brecciation is present in this interval and is chaotic breccia to fracture breccia. Most of the fractures and the intergranular spaces have been

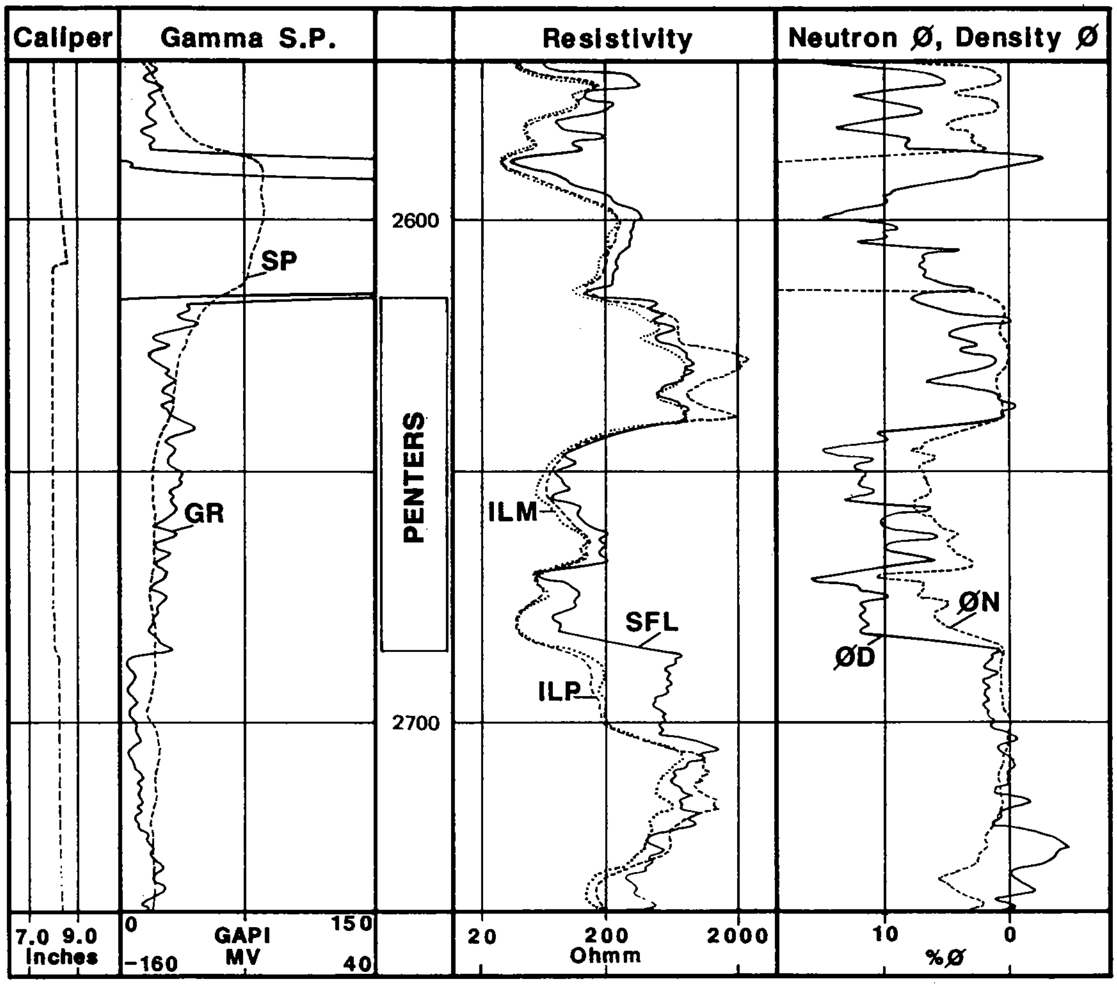


Figure 6. Composite well log of the Hunt Oil Co. Federal Land Bank of Wichita (sec. 28, T. 12 N., R. 25 E., Sequoyah County, Oklahoma) showing the typical log response of the Penters Chert.

filled with carbonate cement; however, minor amounts of fracture porosity have been preserved (Fig. 11).

The lower interval of this core consists of chert and tripolitic chert. Like the upper interval, the chert here is also brecciated, with most of the breccia resembling a fracture breccia. Some oil stain is present in this interval.

Pruet and Hughes No. 1 Vaughan

This well is located in sec. 10, T. 13 S., R. 19 W., Monroe County, Mississippi. Devonian strata were cored at 3,500–3,519 ft.

The interval from 3,500 to 3,519 ft is Devonian rock, based on log correlations. This interval is composed of chert and tripolitic chert, with minor amounts of dolomite and calcite. Dolomite generally is present as fine euhedral rhombs within the chert. Calcite typically is present cementing fractures.

Brecciation is common throughout this interval of the Devonian (Fig. 12A). Because of brecciation the original texture is hard to discern; fossil fragments and other primary features are not recognizable. Secondary features, such as fractures, internal sediments, and geopetals, are much more common. Internal sediments typically are present in some of the fractures (Fig. 12B). Where present, geopetals are located in fractures and consist of internal sediment that is overlain by calcite cement.

Fractures, small clasts of chert, and internal sediments are interpreted to have formed during karstification; however, most of the fracture porosity has been destroyed by cementation and sedimentation. There is some microporosity associated with the more tripolitic portions of the Devonian. Locally, stylolites segregate zones of microporosity.

DEPOSITIONAL ENVIRONMENT

Depositional Model

The presence of sponges, pelmatozoans, brachiopods, mollusks, bryozoans, and radiolarians indicate that the Devonian chert was deposited under normal marine conditions. The abundance of silica and siliceous organisms indicates that the Devonian chert may have been deposited in an area where currents supplied silica-rich waters.

Although the water depth for deposition of Devonian sediments is difficult to estimate, the geometry and constituents of the Devonian chert indicate a high-stand to shelf-margin systems tract. The fauna that is present precludes deposition in a peritidal/tidal-flat depositional en-

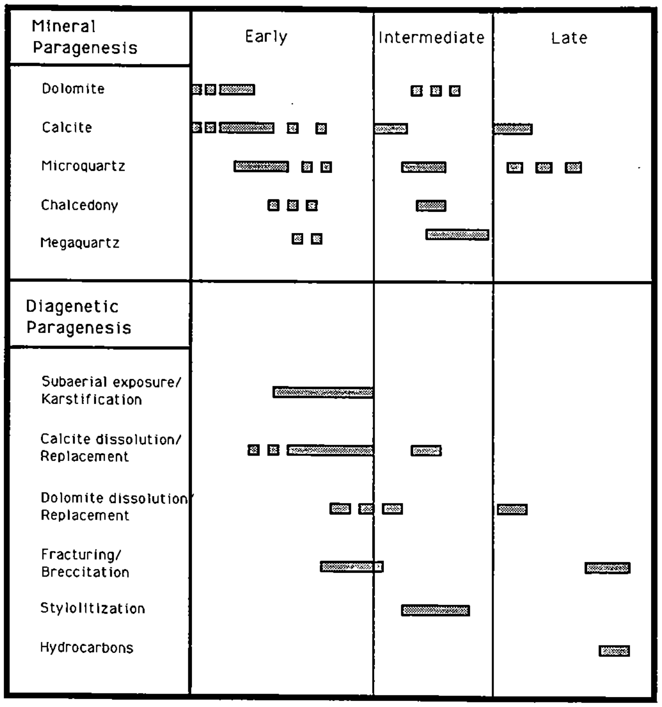


Figure 7. Typical diagenetic sequence of the Devonian chert in the Black Warrior and Arkoma basins.

Figure 8. Brecciated, dolomitic chert resulting from karst development in the Huber No. 1 Fargo well; depth is 3,092 ft.

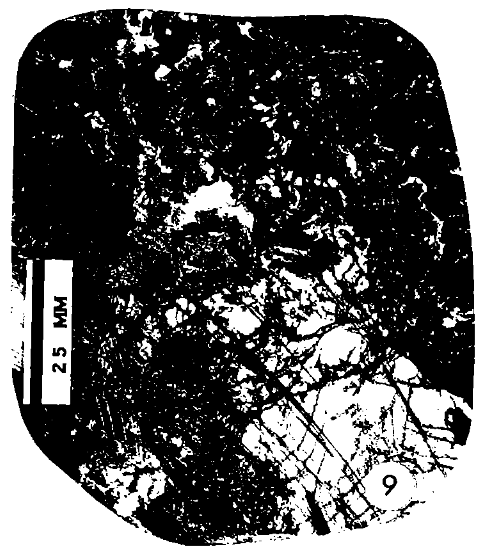


Figure 9. Penters Chert, showing fractures and dissolution features associated with karst development. Western Coal and Mining Co. No. 3 well; depth is 7,749 ft.

vironment. The absence of coated grains indicates that the formation probably was deposited below active wave base; calcareous algae and coral are not present, possibly indicating deposition below the photic zone. Fragmentation of the calcareous fossil fragments may have been accomplished by bioerosion and/or intermittent storm events. It is also unlikely that the fauna present, and the subsurface geometry of the Devonian, would represent an abyssal or basinal deep-water environment of deposition. It is probable that the Devonian was deposited in a shelf environment, under normal marine conditions, with water depths near or below the photic zone. The thicker areas of the Devonian chert, which parallel the Ouachitas, may indicate proximity to the shelf edge.

Silica Source

Important to the depositional model for the Devonian chert, or any other chert formation, is the source of the silica. This controversial subject has been debated for the novaculites in the Ouachita orogen. Several sources that have been hypothesized include: crystallization of siliceous oozes, alteration of volcanic ash, volcanic ash to increase the amount of silica available for siliceous organisms, silicification of a limestone precursor, and silica from a biogenic source (McBride, 1989). The probability of a volcanic source for silica is not very likely, in that ash beds have not been identified in the Devonian section; although, several tuff beds are present in the Mississippian of the Ouachita Mountains.

Silicification of a limestone precursor, by a mixture of marine and meteoric waters, has been proposed by Knauth (1979). In the mixing zone, water that is saturated with silica and undersaturated with calcium carbonate may cause silicification of the sediment; however, it is necessary for the sediments to be siliceous. Subaerial exposure and karstification during development of the pre-Chattanooga unconformity would have allowed the siliceous sediments to be flushed by meteoric waters, facilitating the transformation from biogenic silica to chert.

The primary source of silica in the Devonian chert in the Black Warrior and Arkoma basins is most likely biogenic. Siliceous fossils, specifically sponge spicules and rare radiolarians, have been identified petrographically. The occurrence of radiolarians is much less common than sponge spicules. The absence of siliceous fossils in some

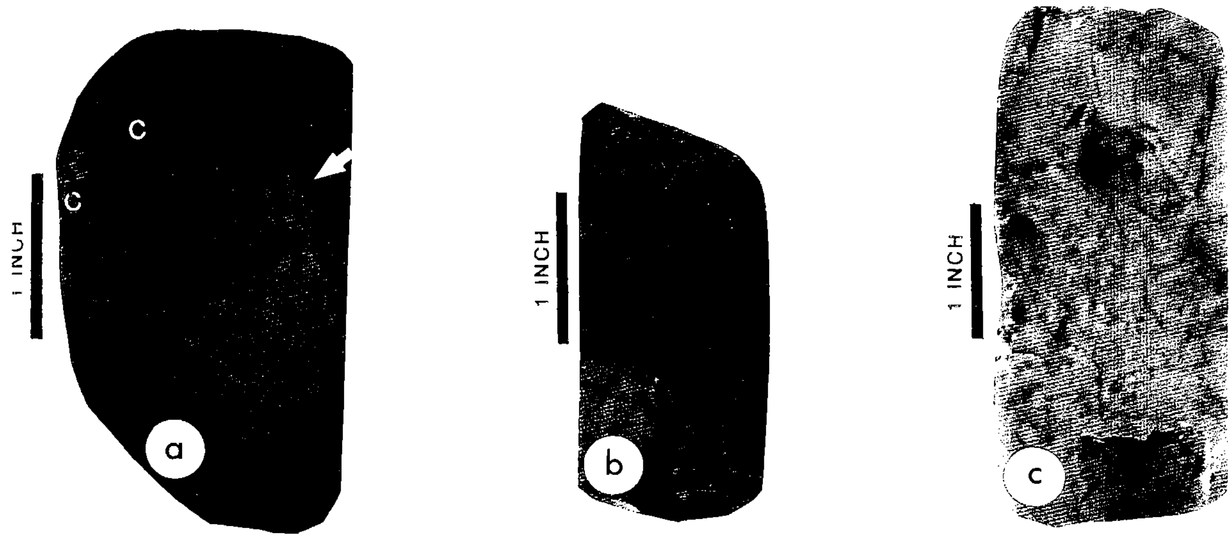
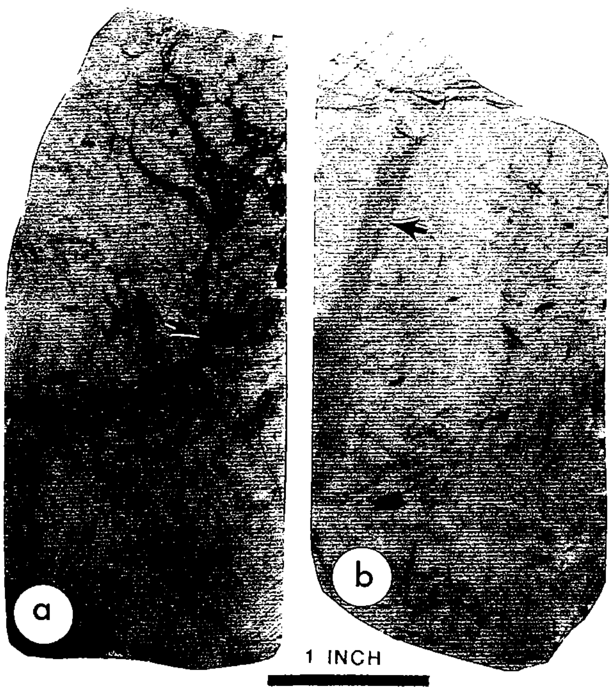


Figure 10. Selected photographs of core from the Greg Drilling No. 1 Gilmore Puckett well. (*A*) Contact (arrow) of the Devonian chert and Chattanooga black shale, at depth of 1,593 ft; several clasts ("c") of the chert are present in the shale. (*B*) Chert breccia in the Devonian, at a depth of 1,594 ft. (*C*) Minor amounts of oil stain in a tripolitic portion of the Devonian, at a depth of 1,626 ft.

Figure 11. Microphotographs showing same view of fractured chert with minor amounts of porosity in the Greg Drilling No. 1 Gilmore Puckett; depth is 1,594 ft. The material in the fracture consists of calcite (c), microquartz or chert (m), quartz grains/crystals (q), and minor amounts of pyrite (opaque). (*A*) Plane-polarized light; (*B*) cross-polarized light.



thin sections is to be expected, because biogenic silica (opal) is unstable to metastable, and with increasing diagenesis this mineral converts to more stable forms of quartz.

The question arises as to why there is an apparent proliferation of siliceous organisms in the Devonian? Greater concentrations of silica in sea water may have occurred in several different settings, including areas of upwelling, areas where surface currents diverge (equatorial regions), and along the west coasts of continents (Calvert, 1974). Lowe (1975) proposes that upwelling occurred along the Paleozoic southern margin of North America; i.e., the Ouachita basin and shelf area to the north (Fig. 13). Siliceous organisms flourished with the advent of silica-rich waters, resulting in an increase in siliceous sediments.

Figure 12 (left). Photographs of core from the Pruet and Hughes No. 1 Vaughan. (*A*) Chert breccia, where brecciation appears to crosscut stylolite (arrow) at a depth of 3,503 ft; fractures are filled with both calcite and internal sediment, which consists of very fine-grained carbonate and chert debris. (*B*) Fracture that is filled with very finely laminated internal sediment (arrow) at a depth of 3,505 ft; several chert clasts are present at the top of the fracture.

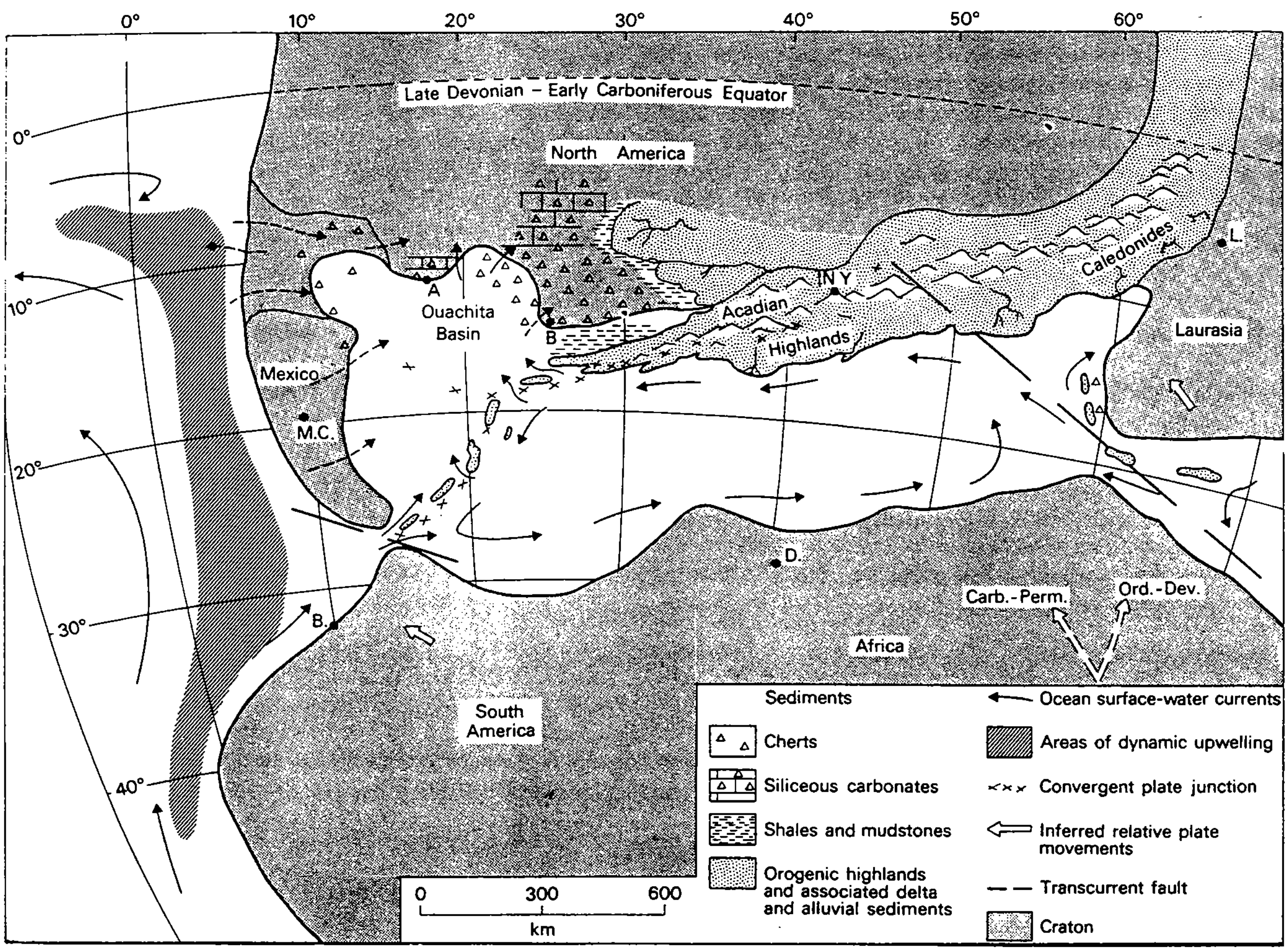


Figure 13. Late Devonian–Early Mississippian paleogeography, paleotectonics, and local sedimentary facies around the western Atlantic Ocean (Lowe, 1975). Letters refer to current geographical locations. In North America: MC–Mexico City; A–Austin, Texas; B–Birmingham, Alabama; NY–New York City. In Laurasia: L–London. In Africa: D–Dakar. In South America: B–Bogota, Columbia.

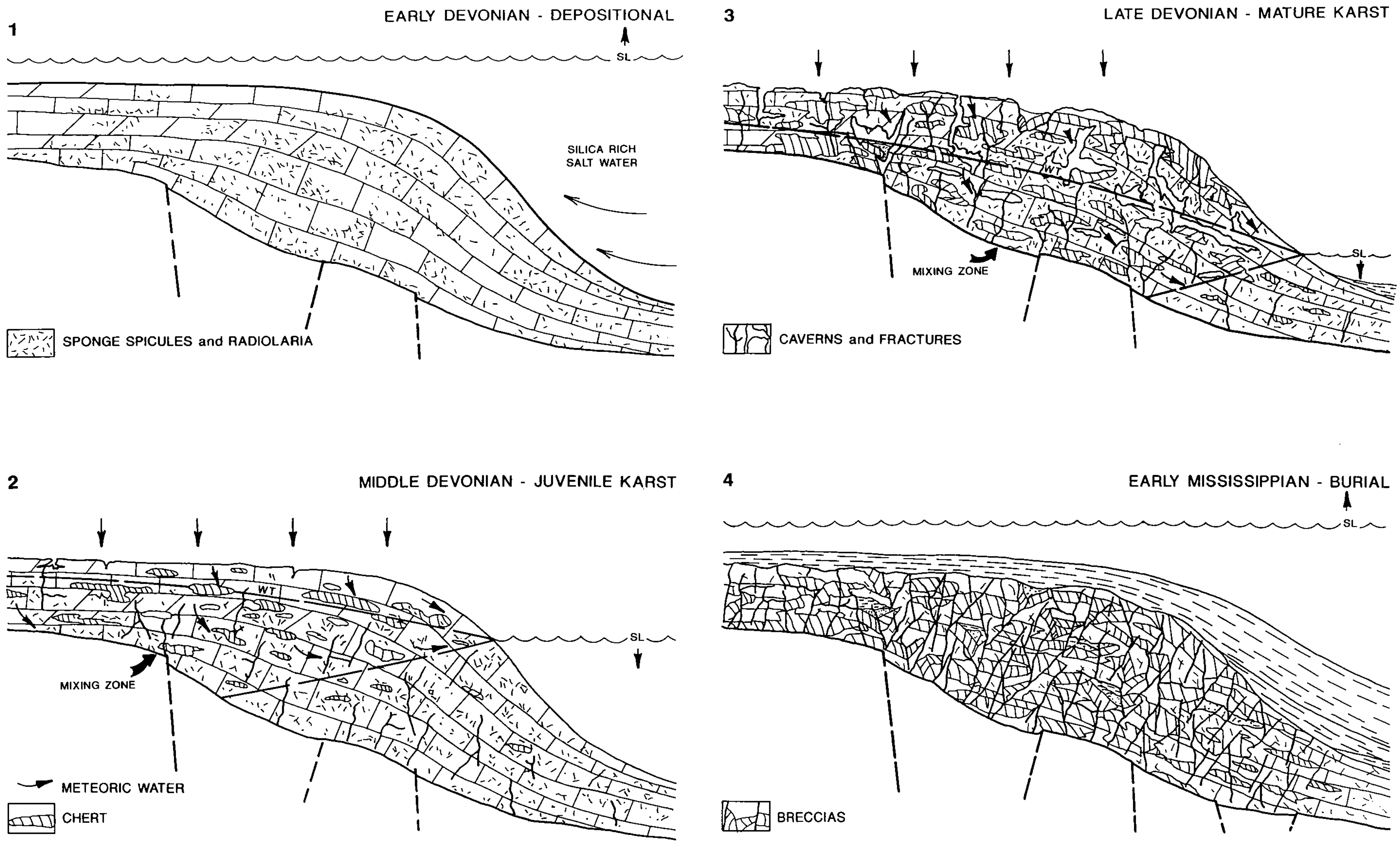


Figure 14. Devonian chert depositional sequence showing (1) deposition, (2 and 3) early karstification during development of the pre-Chattanooga/Woodford unconformity, and (4) deposition of the Chattanooga/Woodford Shale.

KARSTIFICATION AND DIAGENESIS

In many ways the Devonian chert is a product of karstification and diagenesis. Much of the silica, dolomite, and calcite were formed diagenetically. It is possible to subdivide the diagenesis into early, intermediate, and late events (Figs. 7,14). Early diagenesis, begun just after (and possibly during) deposition, consisted of dolomitization, silicification, and calcite cementation. Late diagenesis occurred after lithification, karstification, and some burial. Subaerial conditions, as a result of the pre-Chattanooga/Woodford unconformity, allowed karstification to occur.

Pre-Chattanooga/Woodford Unconformity and Karstification

The updip limit of the Devonian probably represents an erosional edge. Subaerial conditions, karstification, and brecciation have been advocated for zones within the Arkansas Novaculite and Caballos Formation (Folk and McBride, 1976; Lowe, 1989), the downdip equivalents to the Devonian chert.

Breccias have been identified at Devonian outcrops in the Ozarks and also in core and samples. Some of these breccias certainly represent subaerial exposure and karstification; however, later tectonic fractures also appear to be present and appear to crosscut earlier fractures.

During development of the pre-Chattanooga/Woodford unconformity, subaerial conditions, erosion, and karstification affected the Devonian sediments (Fig. 14). Meteoric waters, undersaturated with respect to silica, would have quickly mobilized the metastable biogenic silica. Silicification would have occurred once these waters became saturated with respect to quartz.

During subaerial exposure, meteoric waters infilitrated the Devonian sediment. A hydrogeologic profile formed with vadose and phreatic zones (Fig. 14). Dissolution of carbonates is favored in the vadose and upper phreatic zones. It is also likely that any remaining biogenic silica would either be converted to quartz or go into solution to later be precipitated as quartz. It is likely that some of the calcium carbonate (aragonite and calcite) was also dissolved during karstification. The clasts are predominately chert, indicating that dissolution of calcium carbonate promoted the loss of support and resulted in mechanical failure and brecciation of siliceous intervals. The matrix or cement between clasts includes diagenetic silica, sand, and carbonates. It is probable that the sand was introduced during subaerial exposure. Some of the internal sediment between clasts is laminated.

CONCLUSION

Many mature to senile karst breccias do not make good reservoirs, due to the lack of porosity and general heterogeneity; however, good-quality reservoirs are present in the Devonian of the Bonanza and Fort Chaffee fields in the Arkoma basin. The Penters Chert of the Arkoma basin and the Devonian chert of the Black Warrior basin are unusual in that their composition is predominantly chert, chalcedony, and dolomite. All of these lithologies are subject to fracture development, which provides both porosity and permeability pathways. Combined with intercrystalline and vuggy porosity of the dolomite, the Devonian chert can be excellent reservoirs and be good candidates for oil and gas exploration.

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Sequence Stratigraphy of the Hunton Group as Defined by Core, Outcrop, and Log Data

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ABSTRACT.—The uppermost Ordovician to lower Devonian Hunton Group of the Anadarko and Arkoma basins was deposited on a carbonate ramp in a shallow epicontinental sea. Environmental facies that can be recognized in cores and outcrops are supratidal, intertidal, and subtidal; each facies can be further subdivided into upper and lower parts. These facies occur in shallowing-upward cycles that are aggradational and progradational. Calibration of cores to electric logs allows qualitative facies identification on a large scale and permits mapping of these facies.

With the exception of the Frisco Formation, all Hunton reservoirs occur in predominantly dolomitized rock. Dolomite is, in part, facies related; the supratidal and intertidal facies are the most dolomitized, and the subtidal facies have less dolomite. Early, hypersaline dolomitization was followed by mixed-water dolomitization. Leaching of calcite fossils after hypersaline dolomitization, and preceding mixed-water dolomitization, formed excellent reservoir rock in the lower intertidal to upper subtidal facies.

The pre-Woodford and intra-Hunton (especially the pre-Frisco) unconformities removed and/or truncated much of the Hunton Group in Oklahoma. Updip pinchouts related to the unconformities, along with some structural overprints, have formed excellent reservoirs in the Hunton.

INTRODUCTION

The Hunton Group represents a prolific oil- and gas-producing horizon in both the Midcontinent and the Midland basin. Equivalent strata in the Black Warrior basin and other interior basins also show potential for production. Most of the Hunton accumulations are structural/stratigraphic traps, often produced by truncation of porous carbonate across structural noses. Some of the greatest potential for Hunton production is now found in the deepest parts of the Anadarko basin; some of the deepest gas fields in the world are located in the Hunton Group at depths below 20,000 ft along the Oklahoma/Texas border.

From 1982 to 1989, the Hunton Group was studied by employees and consultants of Masera Corp. as part of a commercial project. The project area included the Anadarko basin, central and southern Oklahoma, and the Arkoma basin (Fig. 1). More than 10,000 wells were correlated and evaluated for reservoir potential. Regional and detailed cross sections were built and used to construct detailed maps that show thickness and net porosity of individual zones within the Hunton. Although most of the results of this project are still proprietary, some of the basic findings are discussed in this paper.

REGIONAL SETTING

The Silurian–Devonian was a time of widespread marine-carbonate deposition. Marine waters covered the Trans-Continental arch and most of the Canadian shield. In fact, the transgression responsible for this expansive sea was fully as extensive as that of the Ordovician. Considering the great thickness of Cambrian–Ordovician carbonates, the Silurian–Devonian strata were deposited as a relatively thin veneer of limestones and dolomites, with locally significant deposits of sandstone and shale.

In the Midcontinent, the latest Ordovician–Silurian–Early Devonian is represented by the Hunton Group, which was deposited primarily as subtidal/intertidal facies in a ramp-type environment. Outside the Midcontinent, extensive reefs are present in equivalent strata along the cratonic edges in Nevada, Canada, Franklin trough of Baffin Island, and Greenland. Off-shelf, dark graptolitic shales are typical basinal sediments.

Silurian strata are more extensive than the Devonian and are preserved in diatremes in northern Colorado and southeastern Wyoming on the crest of the Trans-Continental arch (Fig. 2; Wilson, 1975). Silurian and lower Devonian carbonates are also found in karstic deposits within underlying Ordovician strata in the Llano uplift.

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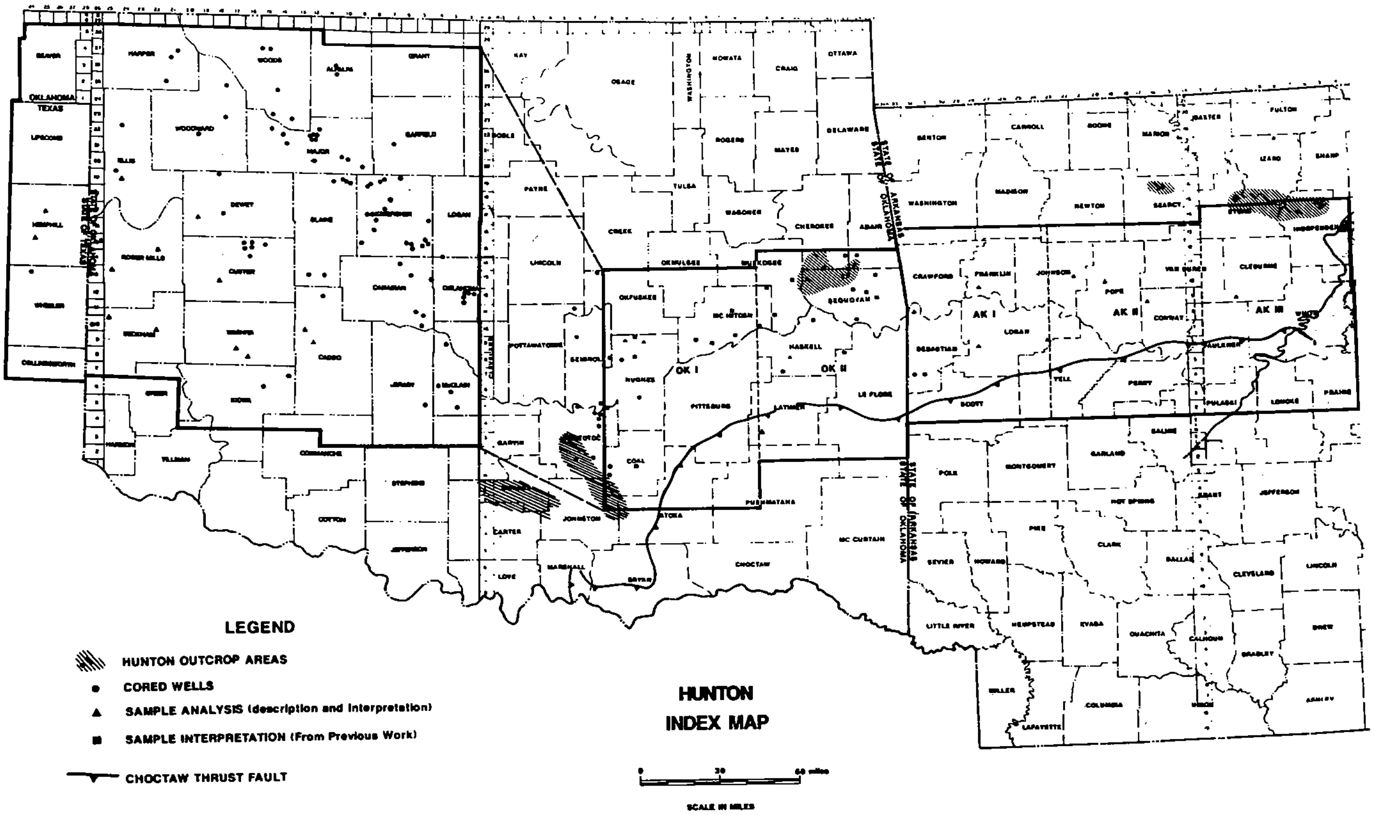

Figure 1. Index map showing area of the Hunton Group study.

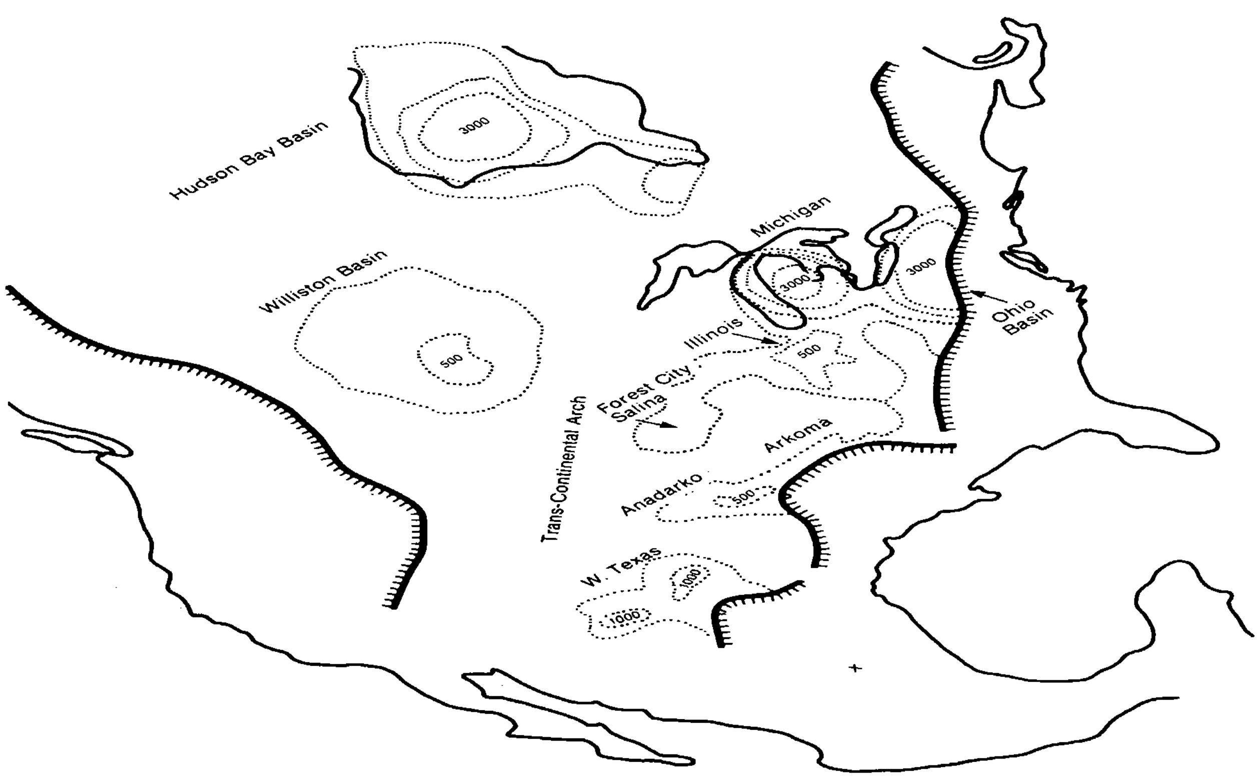

Figure 2. Map showing distribution and thickness (in feet) of Silurian strata in North America.

STRATIGRAPHY

The Hunton Group is a readily recognizable distinctive unit because it is stratigraphically sandwiched between the Woodford Shale, above, and the Sylvan Shale, below (Fig. 3). The Hunton is composed of sequences of dolomite, limestone, and calcareous shale. Based primarily on outcrop surveys, the Hunton is divided into a number of formations. The Chimneyhill Subgroup, at the base, is comprised of the Ordovician Keel Formation and the overlying Silurian Cochrane and Clarita Formations (Amsden, 1961, 1975,1980). The Chimneyhill is overlain successively by the Silurian Henryhouse and the Devonian Haragan and Bois d'Arc Formations (Fig. 4). In central and southern Oklahoma, the Bois d'Arc is overlain by the Frisco Formation, and in the Arkoma basin the uppermost Hunton is composed of the Sallisaw Formation (Penters Chert). Some workers do not include the Sallisaw Formation within the Hunton Group.

The overlying Woodford Shale and equivalent strata are part of an extensive sequence of hydrocarbon source beds that provided oil and gas for many of the Paleozoic petroleum reservoirs. The Woodford underlies most of Oklahoma, ranging in thickness from a feather edge in northern Oklahoma to >700 ft in southern Oklahoma. It contains conodont fauna of Late Devonian to Early Mississippian age.

The Woodford is easily distinguished on outcrop by its dark-gray to black color and its cherty composition, and in the subsurface it is an excellent "marker bed" because of its distinctive "high" gamma-ray response (Fig. 3).

The underlying Sylvan Shale is typically a greenish-gray to greenish-brown marine shale. On average the Sylvan is <50 ft thick, although it attains a thickness of >300 ft in the Arbuckle Mountains.

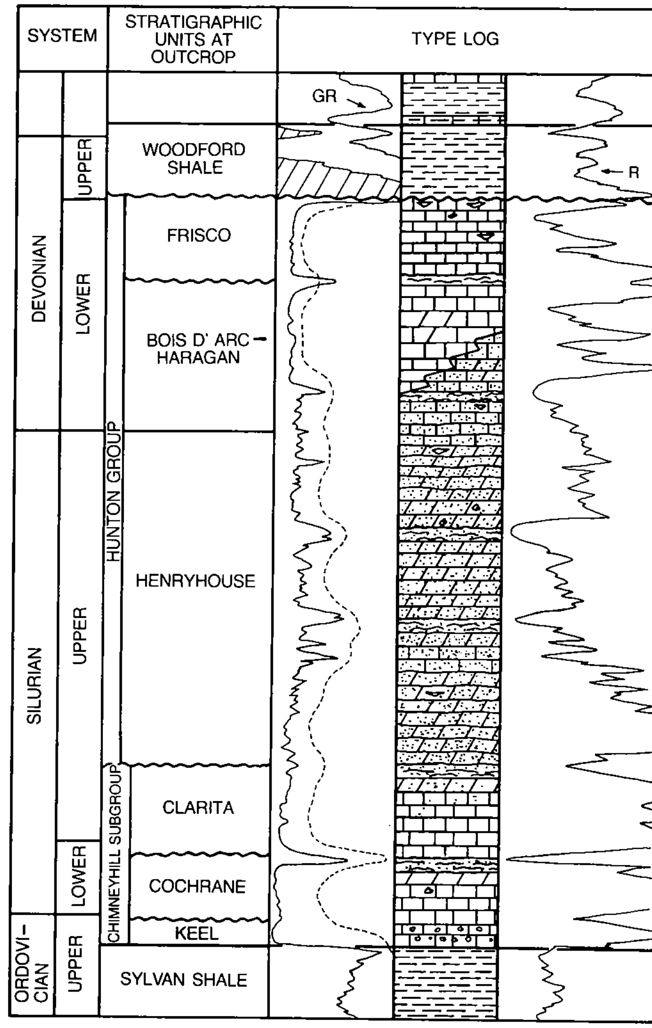

Figure 3. Type log of the Hunton Group in central Oklahoma (excluding the Sallisaw Formation).

PETROGRAPHY

More than 100 cores from Texas, Oklahoma, and Arkansas were described and evaluated during the course of this study. Recognition of lithofacies in these cores is based on texture, sedimentary structures, constituents, and geometry.

Lithofacies are classified according to Dunham's organization of inferred depositional textures (Fig. 5). Mud-supported rocks are much more abundant than grain-supported rocks, although a gradation exists between the two types. The Chimneyhill Subgroup and the Frisco tend to have more grain-supported textures, whereas the Henryhouse to Bois d'Arc section is typically mud-supported. The precursor to the Penters Chert was most likely mud-supported, but it is difficult to determine the original fabric due to the amount of alteration and brecciation.

Burrows and bioturbation are the most common types of depositional structures. Ripple bedding and cross bedding are typical of pelmatozoan and oolitic grainstones and packstones. Algal laminations are common and are often associated with fenestral fabric. Thin- to medium-bedded strata are present on outcrop and in core.

The most common constituents are skeletal fragments and lime mud (micrite). Peloids and ooids are common in high-energy facies. Skeletal fragments consist mostly of pelmatozoan material (dominantly crinoids). Certain portions of the Hunton contain abundant brachiopods (such as the Henryhouse) and abundant bryozoans (such as the Frisco). Other skeletal grains include algae, trilobites, corals, mollusks, and ostracods.

Terrigenous constituents are typically clay, with some quartz silt and sand. Penecontemporaneous constituents include dolomite, anhydrite, and glauconite.

Selected Core Analyses

The following cores were selected to represent various stratigraphic units and facies.

Gulf No. 1 Streeter

Most of the Hunton (Frisco 6,954–7,043 ft; Henryhouse 7,043–7,085, 7,105–7,268; and Cochrane 7,282–7,298 ft) is represented in this core (Fig. 6). This well is in the West Edmond field, in sec. 20, T. 13 N., R. 4 W.

The Frisco Formation (6,954–7,043 ft) is an oil-stained limestone that ranges from mud/wackestone to grainstone (Fig. 7). The fossil content includes pelmatozoans, bryozoans, brachiopods, and trilobites (Fig. 8). Porosity is either intraparticle or vuggy. The vugs are actually solution-enlarged intraparticle pores. The environment of deposi-

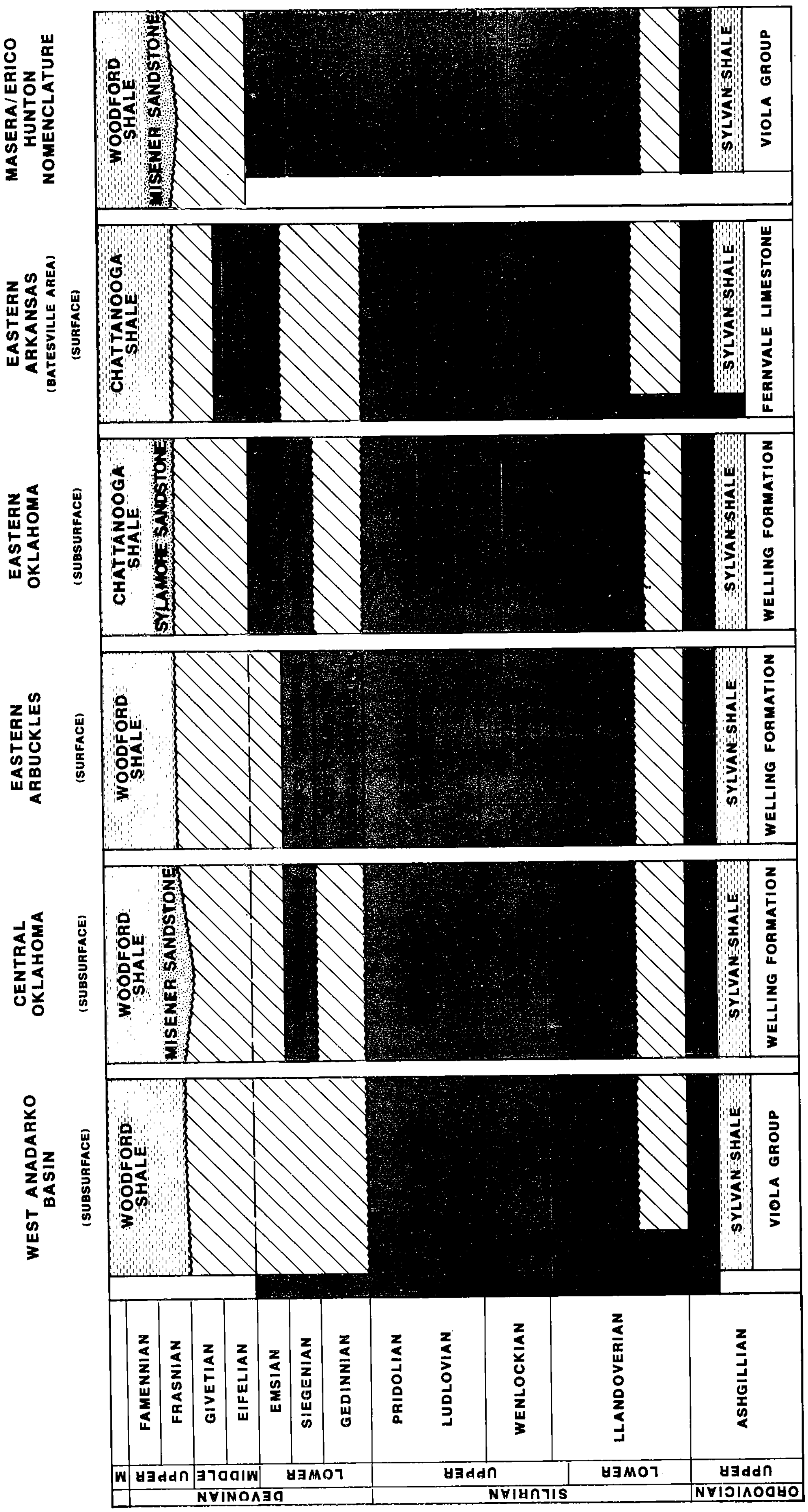

Figure 4. Correlation chart of the Hunton Group (revised from Amsden, 1975).

Depositional Texture recognizable					Depositional texture not recognizable
Original components not bound together during depositions				Original components were bound together during deposition... as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or questionably organic matter and are too large to be interstices.	Crystalline carbonate (Subdivide according to classifications designed to bear on physical texture or diagenesis.)
Contains mud (particles of clay and fine silt size)			Lacks mud and is grain-supported		
Mud-supported		Grain-supported			
Less than 10% grains	More than 10% grains				
Mudstone	Wackestone	Packstone	Grainstone	Boundstone	

(DUNHAM,1962)

Figure 5. Dunham's classification chart (Dunham, 1962).

tion was probably subtidal, and the range in texture probably corresponds to the range in strength of the currents. Mud-mound complexes are suggested by the abundant crinoid fragments.

The upper part of the Henryhouse (7,043–7,148 ft) is a gray lime wackestone, with an oolitic grainstone at 7,052–7,055 ft. The wackestone is slightly dolomitic in part of the interval. Fossils in the Henryhouse are pelmatozoans, brachiopods, trilobites, bryozoans, and ostracods. Effective porosity is present in only the oolitic zone. A lagoonal facies (7,043–7,052 ft) with sparse fossils, suggestive of restricted conditions, overlies the oolitic shoal and represents progradation. The oolite is a high-energy deposit and probably formed as a shoal. Underlying the oolitic grainstone is a subtidal deposit (7,055–7,148 ft), based on the diverse and delicately preserved fauna, very poor sorting, abundant mud, and stratigraphic position.

From 7,148 to 7,180 ft, the Henryhouse is a pelmatozoan-rich packstone with some micrite and ooids. This interval probably represents an upper-subtidal environment.

The interval from 7,180 to 7,268 ft (approximate base of the Henryhouse) is generally a gray wackestone/mudstone and packstone, basically limestone (except for a dolomitic bed at 7,245–7,254 ft). Fossils in the limestone are brachiopods, pelmatozoans, trilobites, bryozoans, and ostracods. This interval is interpreted to have formed in a subtidal setting well below wave base, as indicated by the diverse fauna and delicately preserved features. The dolomitic unit (7,245–7,254 ft) is interpreted as intertidal facies because of small vertical burrows intersecting thin laminations.

The lower cored interval (7,282–7,298 ft) is a gray lime wackestone, which is part of the Chimneyhill Subgroup. Low-energy subtidal conditions are indicated in this interval by the diverse fauna and well-preserved delicate features.

Sunray No. 1 Frans

This well is located in sec. 3, T. 15 N., R. 16 W.; Custer County, Oklahoma. The total interval that was cored is from 14,505 to 14,720 ft (Fig. 9). When comparing the cored interval to the wireline log, the core needs to be adjusted downward 5 ft to correspond to log depths. The Henryhouse Formation of the Hunton Group comprises the Franz core. The Henryhouse consists of two facies that show aggradation and progradation, with an intertidal facies overlying a subtidal facies.

The upper facies (14,505–14,680 ft) is a medium-gray, burrowed dolomudstone/wackestone (Fig. 10). Thin breccias are present at 14,513–14,514 ft and 14,575 ft. Replacive-calcite nodules and vugs are present along with less common vug-filling and nodular anhydrite (Fig. 11). Visible

Gulf Streeter No. 1
Sec. 20, T13N, R4W

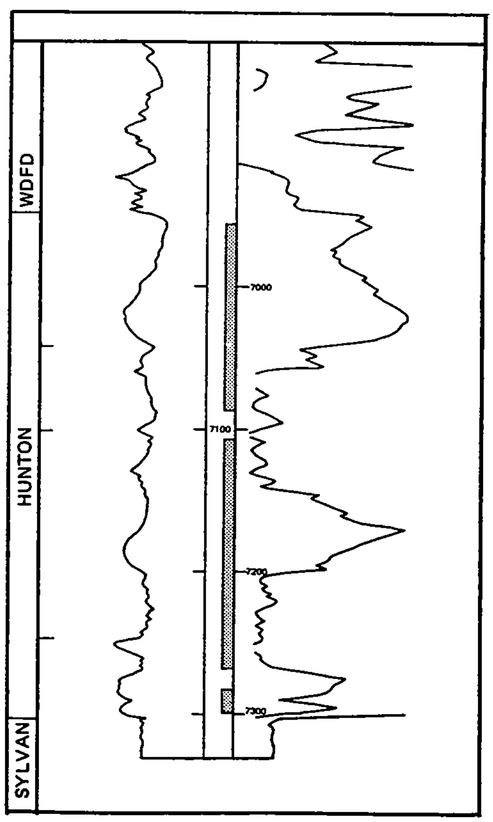

Figure 6. Well log of the Gulf No. 1 Streeter showing location of cored interval.

porosity is developed in burrowed zones; there are scattered occurrences of disseminated dead oil. This facies is interpreted to have been deposited in an intertidal environment.

The lower facies (14,680–14,720 ft) is a medium- to dark-gray, lime wackestone/mudstone. Fossil fragments are common, as is lamination to very thin bedding. Chert occurs as scattered nodules. The depositional environment is interpreted to have been in a subtidal facies.

Sun Oil No. 11 Kilcrease

This well is located in sec. 36, T. 5 N., R. 7 E., Pontotoc County, Oklahoma. The Hunton Group formations in the core are Henryhouse, Clarita, Cochrane, and Keel, along with the underlying Sylvan (Fig. 12). The core needs to be moved down ~5 ft to correspond to the wireline log.

The upper part of the Henryhouse (3,902–3,916 ft) consists of a greenish-gray to gray dolomitic lime mudstone/wackestone (Figs. 13,14). Dolomite content is 40–50%.

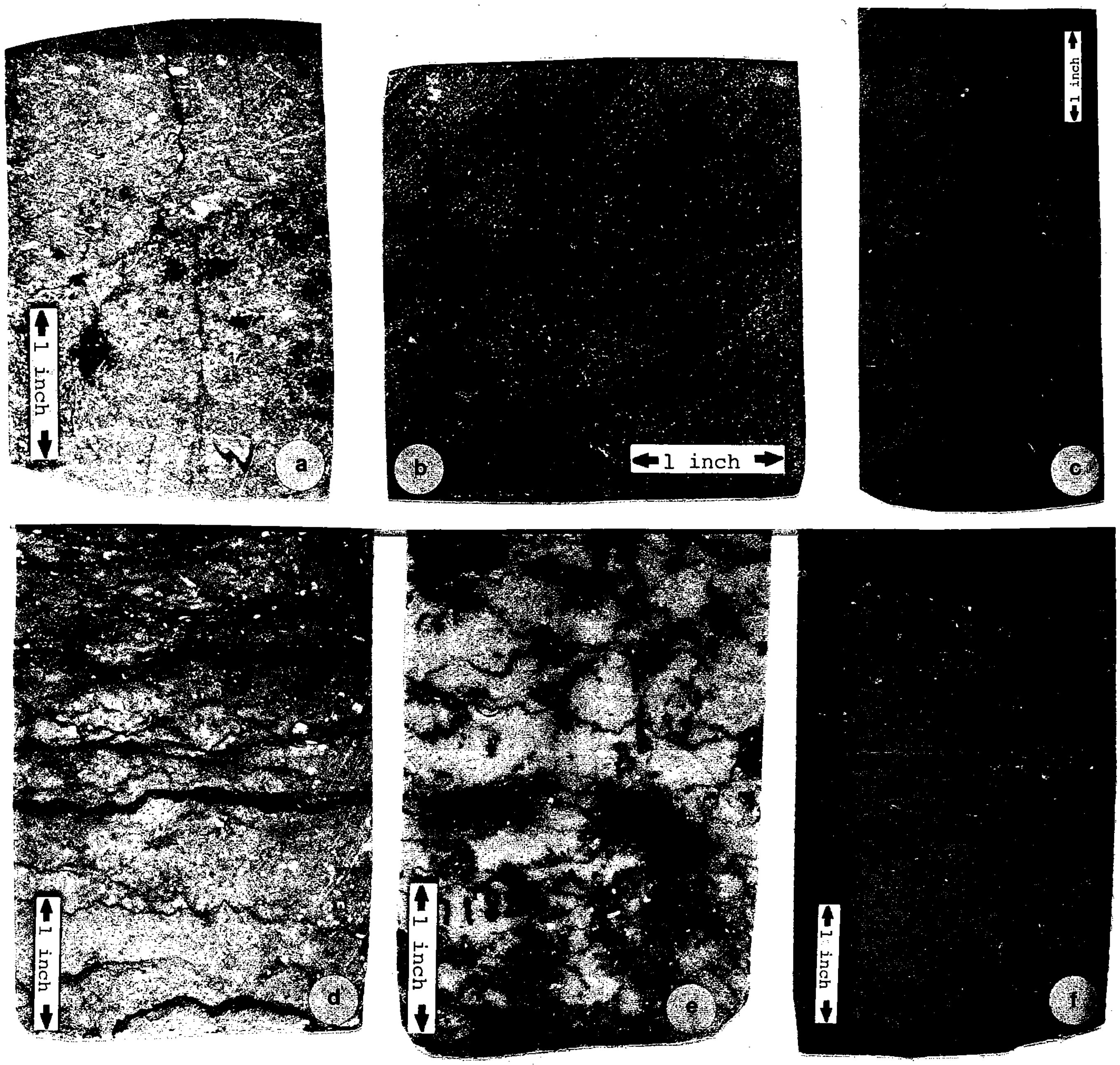

Figure 7. Photographs of selected core from the Gulf No. 1 Streeter: **(a)** 6,980 ft—Crinoidal packstone with some oriented grains and solution features (dark patches). Frisco Formation, moderate-energy, subtidal environment. **(b)** 7,053 ft—Oil-stained oolitic grainstone. Henryhouse Formation, high-energy, subtidal/intertidal shoal. **(c)** 7,071 ft—Wackestone with hummocky bedding formed by incipient stylotization. Low-energy, subtidal environment. **(d)** 7,160 ft—Crinoidal packstone with low-amplitude stylolites. Henryhouse Formation, low- to moderate-energy, subtidal environment. **(e)** 7,185 ft—Wackestone with solution features (dark areas). Low-energy, subtidal environment. **(f)** 7,249 ft—Dolomudstone. Bioturbated and slightly laminated. Cochrane Formation, intertidal environment.

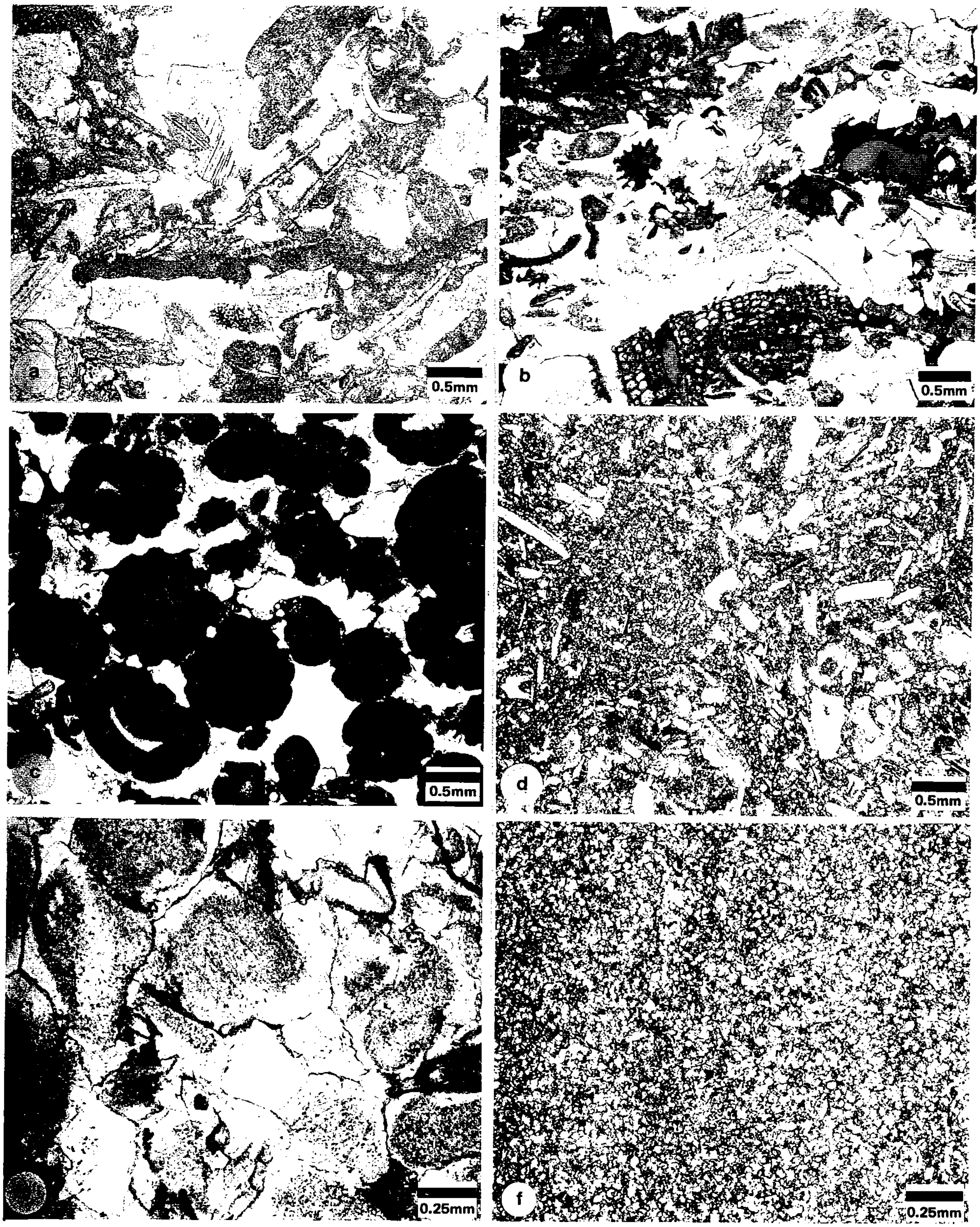

Figure 8. Photomicrographs from analysis of core from the Gulf No. 1 Streeter: **(a)** 7,018 ft—Enlarged intraparticle pores in a grainstone. Moderate- to high-energy, subtidal environment. Plane-polarized light (PPL). **(b)** 7,032 ft—Grainstone with oriented grains, hydrocarbons in pores, and prominent syntaxial overgrowths. Frisco Formation, high-energy, subtidal environment. PPL. **(c)** 7,055 ft—Grainstone with coated grains and interparticle porosity. Henryhouse Formation, high-energy, subtidal/intertidal shoal. PPL. **(d)** 7,058 ft—Poorly sorted dolomitic wackestone/packstone. Low-energy, subtidal environment. PPL. **(e)** 7,150 ft—Crinoidal packstone/grainstone, cemented by syntaxial overgrowths. Henryhouse Formation, moderate-energy, subtidal environment. PPL. **(f)** 7,250 ft—Dolomudstone with possible burrow. Intertidal environment. PPL.

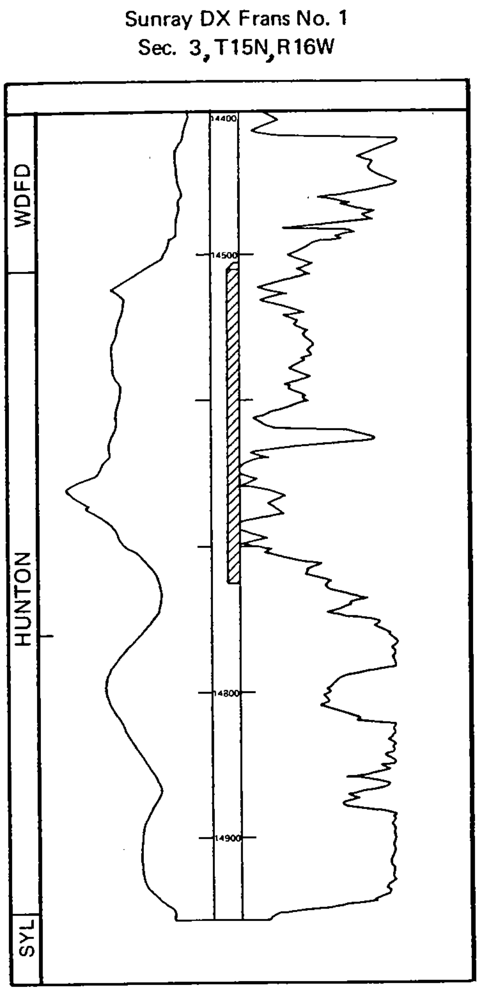

Figure 9. Well log of Sunray No. 1 Frans showing location of cored interval.

Characteristic features are algal laminae, fenestral (birdseye) fabric, and abundant pyrite (Figs. 14,15). Deposition was probably in an intertidal setting.

The lower part of the Henryhouse (3,916–3,955 ft) consists primarily of bioturbated, light- to medium-gray, lime mudstone/wackestone that is dolomitic in part. Dominant macrofeatures are nodular bedding due to compaction and incipient stylolitization, deformed horizontal laminae, and bioturbation. The fauna includes fragments of pelmatozoans, brachiopods, and trilobites. These features indicate a subtidal depositional environment.

The Clarita, from 3,955–3,993 ft, is a pink, crinoidal, lime packstone/grainstone. The uppermost part (3,955–3,961 ft), as well as thinner sections below, is a breccia. Brecciation is due to karstification, with the cavities filled by detrital carbonate (dolomite and limestone) and siliciclastics (mudstone and sandstone—the Misener overlies the Hunton in this well). Some voids are also filled with calcite. The filled solution cavities are both vertically and horizontally oriented. Sedimentary features in the nonbrecciated Clarita include burrows, sparse bedding/laminae, and cross bedding. Faunal content consists mainly of crinoids, brachiopods, and trilobites, with minor bryozoans. This unit is interpreted to have been deposited as part of a subtidal shoal.

The Cochrane (3,993–4,018 ft) is a glauconitic, lime wackestone with mudstone. In this interval, breccias are also present and probably formed during the karst episode that affected the Clarita. Cavities in the Cochrane are also filled by detrital carbonate and siliciclastics along with calcite. Pelmatozoans and brachiopods are the dominant fossil elements. The Cochrane probably was deposited in a subtidal environment.

The Keel (4,018–4,025 ft) is an oolitic, lime grainstone

Figure 10. Photographs of selected core from the Sunray No.1 Frans: **(a)** 14,531 ft—Burrowed dolomudstone with calcite-filled vugs; low porosity. Intertidal environment. **(b)** 14,645 ft—Burrowed dolowackestone with well-developed porosity. Dead oil (black) in some of the pores. Henryhouse Formation, intertidal facies. **(c)** 14,706 ft—Subtle lamination, dark chert (lower right), and sparry calcite nodules in a lime wackestone/mudstone. Henryhouse Formation, subtidal facies. **(d)** 14,714 ft—Faintly laminated, lime wackestone/mudstone, with randomly oriented fossils; minimal burrowing. Subtidal environment.

with some packstone; the upper part has also been affected by brecciation. Horizontal bedding and cross bedding are the main sedimentary structures. Crinoids are abundant in the lower part of the Keel. Glauconite is present in laminae-like streaks. This unit probably represents an upper-subtidal shoal.

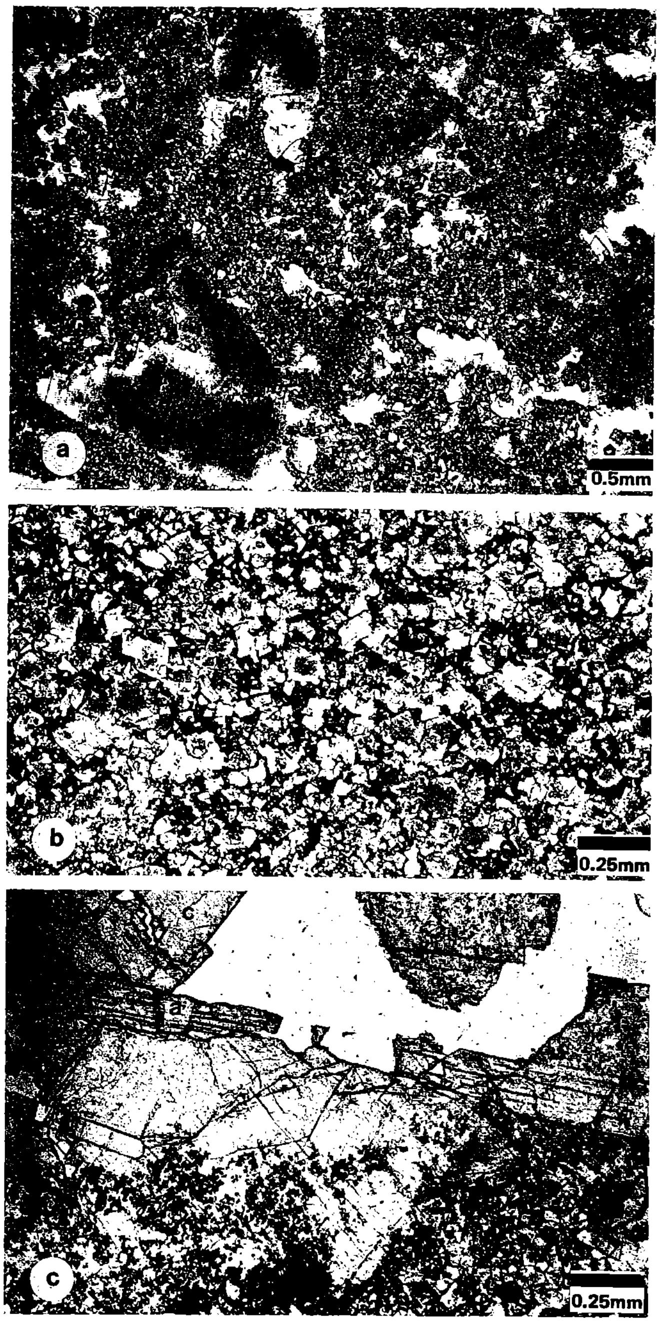

Figure 11. Photomicrographs from analysis of core from the Sunray No. 1 Frans: **(a)** 14,505 ft—Good intercrystalline and moldic porosity, partial dissolution of crinoid-arm plates in burrowed dolowackestone. Henryhouse Formation, intertidal facies. Plane-polarized light (PPL). **(b)** 14,531 ft—Dolomite with dead oil in intercrystalline pore spaces; porosity is best developed where coarser rhombs are located. Henryhouse Formation, intertidal facies. PPL. **(c)** 14,563 ft—Vug filled with anhydrite (a) and calcite (c). Henryhouse Formation, upper intertidal facies. PPL.

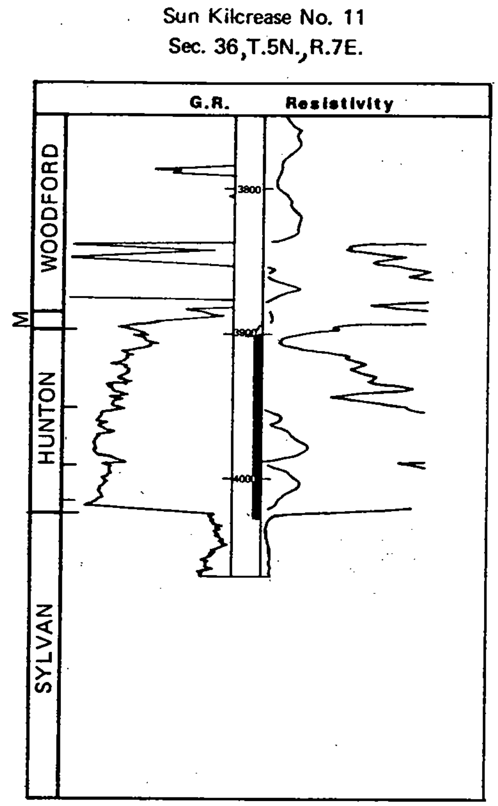

Figure 12. Well log of the Sun No. 11 Kilcrease showing location of cored interval.

DEPOSITIONAL ENVIRONMENT

Based on core and outcrop descriptions, the Hunton was deposited on a broad ramp in a shallow epicontinental sea (Fig. 16). Overall slope was generally toward the south, with the deepest water in the region of the Oklahoma aulacogen; the off-shelf basin was probably located south and southeast of the aulacogen.

Carbonate ramps have been described by Roehl (1967) and Laporte (1968) for ancient carbonate environments. Unfortunately, few, if any, modern analogues exist, although some areas such as the Persian Gulf have similar geometry. It is thought that the local slopes associated with the shoaling areas on the ramp were generally steeper than the very gentle slope of the sea floor.

Depositional Model

Based on the petrographic examination and evaluation of outcrops, cores, and samples, several distinct environments and related lithofacies can be identified. The basic depositional environments are supratidal, which occurred above high tide; intertidal, which occurred below high tide and above low tide; and subtidal, which occurred below low tide. These environments can be further subdivided into upper and lower parts (Fig. 17). Adjacent environments, such as the transitional area between intertidal and subtidal facies, can form distinctive lithofacies and herein are treated as separate facies. Within the subtidal environment there is a facies unique to the Frisco Formation, and that is crinoidal mud mounds. The constituents, textures, and structures of each facies are described below:

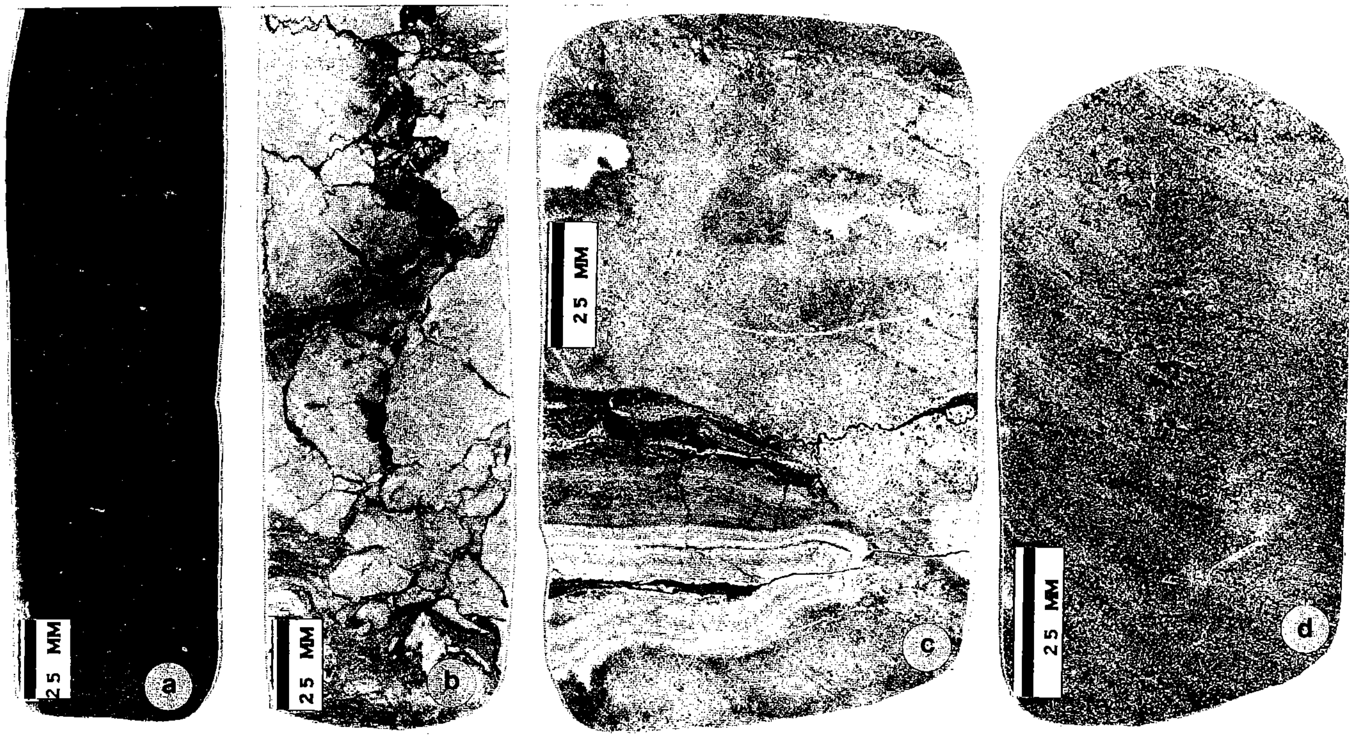

Figure 13. Photographs of selected core from the Sun No. 11 Kilcrease: **(a)** 3,907 ft—Dolomitic mudstone/wackestone with fenestral fabric and algal lamination. Henryhouse Formation, intertidal environment. **(b)** 3,994 ft—Collapse breccia to crackle breccia, composed of wackestone and filled with glauconitic sediments. Cochrane Formation, subtidal environment. **(c)** 4,012 ft—Coarsely crystalline calcite-filling cavity. The rock is glauconitic wackestone. Subtidal environment. **(d)** 4,022 ft—Oolitic grainstone. Keel Formation, subtidal shoal.

Upper intertidal to supratidal.—Algally laminated mudstones predominate in this environment. Fenestral fabrics are common and there is a paucity of fossils and burrowing. Anhydrite nodules and replaced evaporites, such as silica nodules with relict anhydrite crystals, are present in this facies (Chowns and Elkins, 1974; Beardall, 1983).

Upper subtidal to lower intertidal.—This facies is typically a crinoidal wackestone. Brachiopods are also present; trilobites and bryozoans may be present, but are rare. Burrowing is the most common sedimentary feature; it probably enhances permeability of this facies for later dolomitization, which forms a reservoir-quality lithology.

Lagoons. Lagoons may form in this area and the sediments that are deposited typically are peloidal mud/wackestones.

Subtidal.—This facies can also be subdivided into upper and lower (open-marine) systems, in response to wave and current energy. Typical textures are wackestone to mudstones, with rare packstones, and they contain a diverse fauna of crinoids, brachiopods, trilobites, bryozoans, and ostracods. Delicate, thin-shelled fossils are well preserved in the lower subtidal facies. Common sedimentary structures include burrows and hummocky to nodular beds.

Shoal. Oolitic and peloidal grainstones, along with skeletal grainstones, comprise this facies. Cross bedding is common; other sedimentary structures are thin beds and ripple laminae. Where dolomitized, this facies is an excellent reservoir.

Crinoidal mud mounds. This depositional environment has been identified in, and is probably restricted to, the Frisco Formation. It formed where crinoidal- and bryozoan-rich bioherms acted as baffles to trap lime mud, thus forming a mud-supported mound facies with associated grain-supported flanking and capping facies (Fig. 18; Medlock, 1984).

Cyclicity

The above facies occur in shallowing-upward cycles or parasequences. A complete sequence is shown in Figure 19; however, it is important to recognize that complete sequences are not commonly observed, due to progradation and disconformity or unconformity. As previously discussed, the Hunton is composed of a series of progradational, aggradational sequences that built generally southward on a ramp. The composition of each particular sequence is dependent on its position on the ramp. For example, a sequence from the Henryhouse Formation on the outer ramp (e.g., southern Oklahoma outcrops) is relatively thin (<25 ft) and is typically composed of open-marine calcareous shales and mudstones that shallow upward into subtidal mudstones and wackestones (Fig. 20). Farther up the ramp (e.g., Anadarko basin and shelf), equivalent Henryhouse strata are represented by thicker sequences (>50 ft) composed, from bottom to top, of the following: (a) lower subtidal mudstones and wackestones, (b) shallow subtidal wackestones, packstones, and grainstones, and (c) lower intertidal wackestones and pack-

Figure 14. Full-core photographs of the lower Henryhouse (3,947–3,955) and upper Chimneyhill (3,955–3,970) in the Sun No. 11 Kilcrease. Note sandstone-filled karst features, which are dark material from 3,956–3,961 ft.

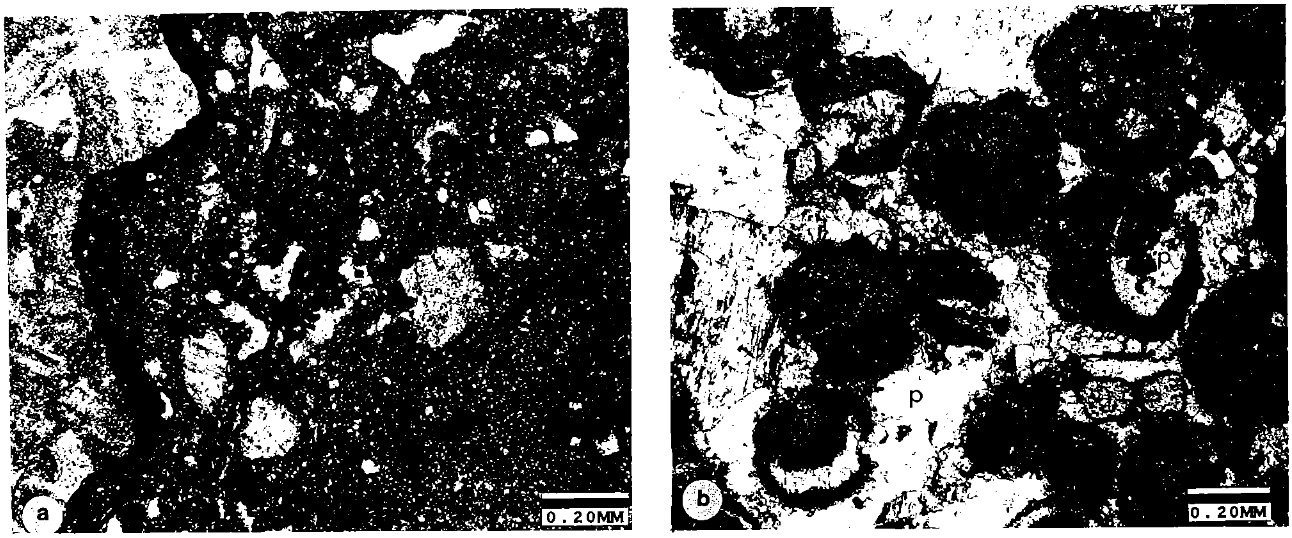

Figure 15. Photomicrographs from analysis of core from the Sun No. 11 Kilcrease: **(a)** 3,957 ft—Collapse breccia; cavity filled with clastic material and packstone. Clarita Formation, subtidal shoal environment overprinted by karstification. Plane-polarized light (PPL). **(b)** 4,024 ft—Inter- and intra-ooid porosity (p) in a grainstone. Keel Formation, subtidal shoal. PPL.

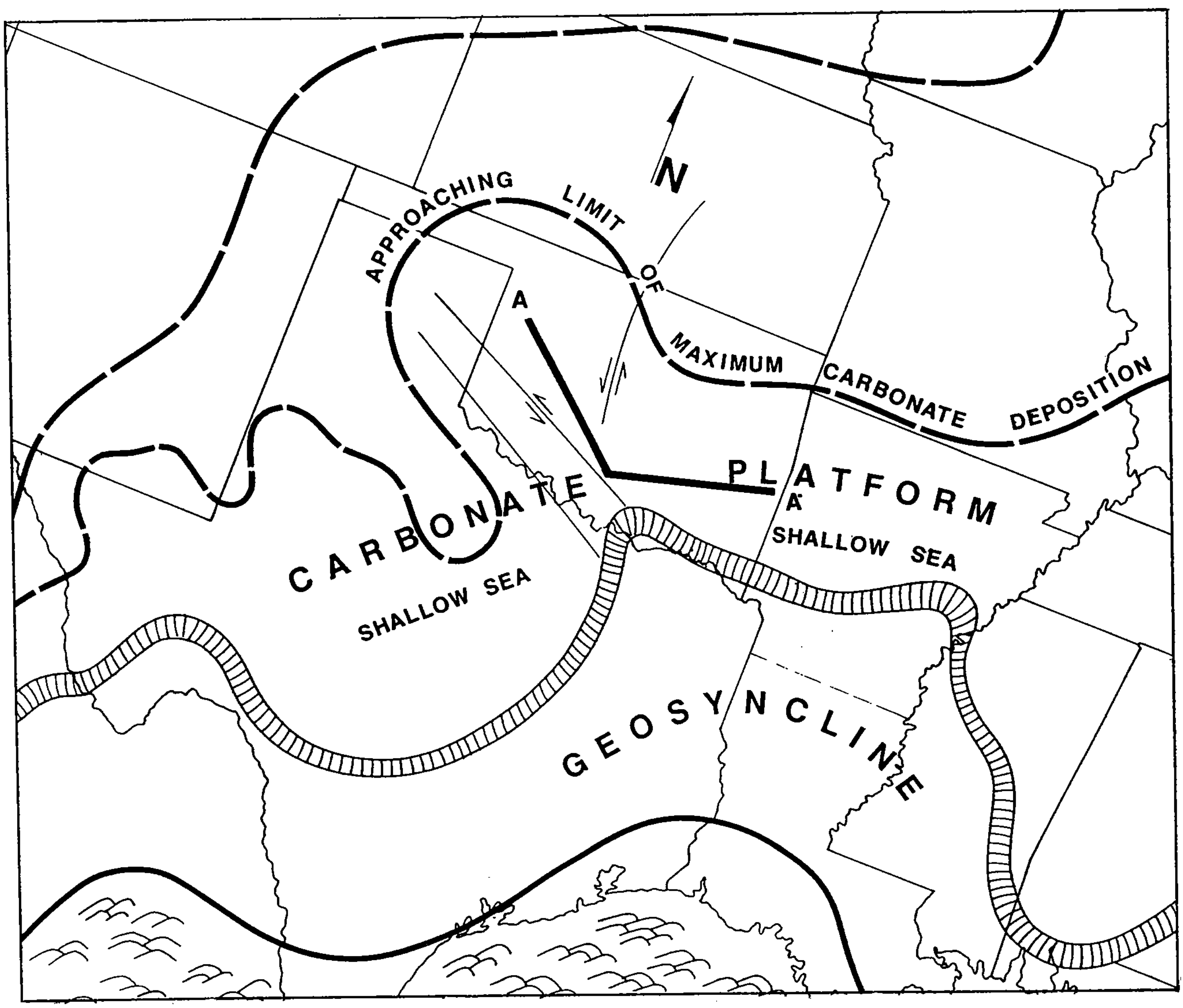

Figure 16. Depositional setting during Hunton time in the Midcontinent.

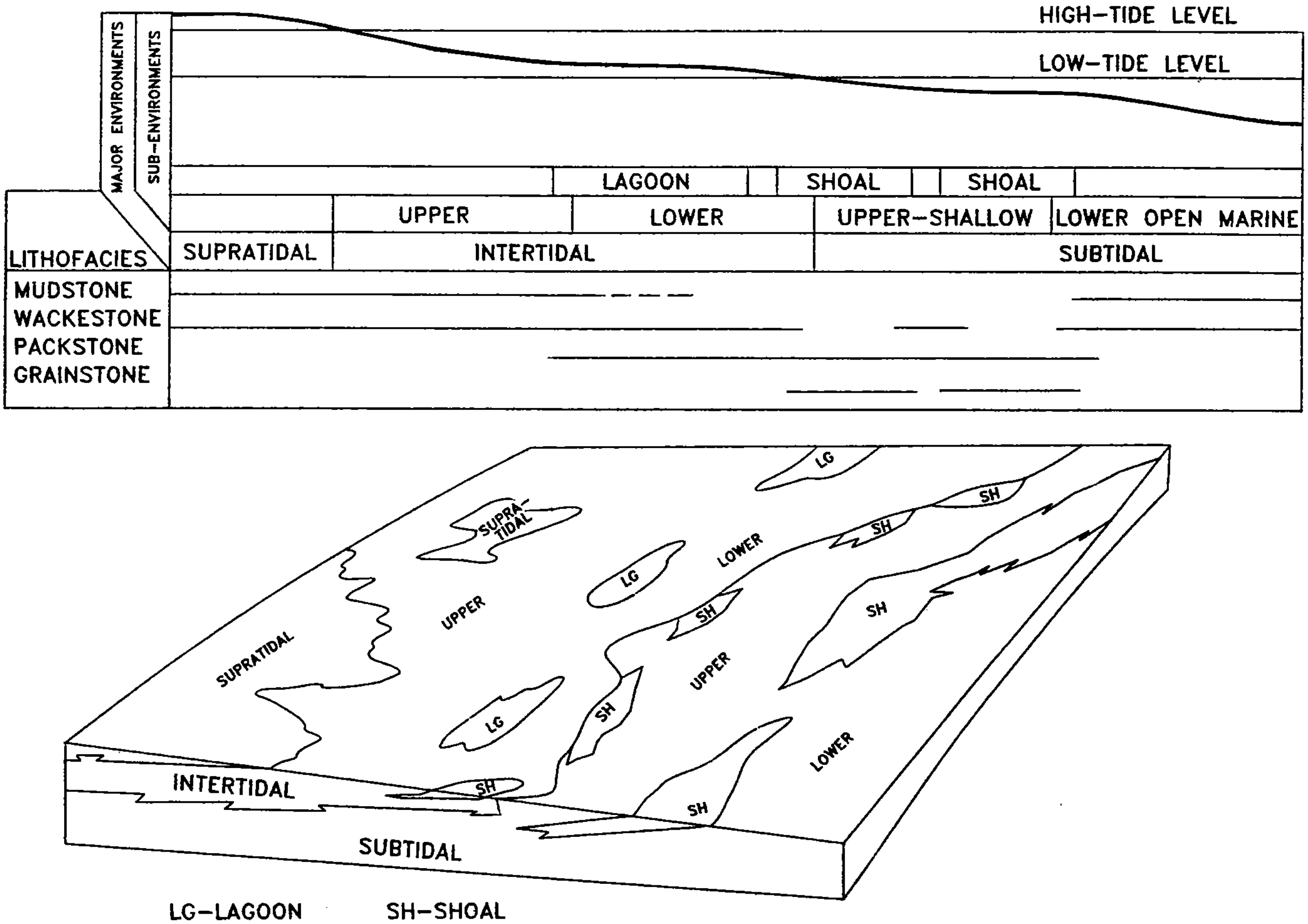

Figure 17. Depositional model for Chimneyhill through Bois d'Arc strata.

stones. On the upper ramp (e.g., northwest Oklahoma shelf), the sequences are usually composed of intertidal to supratidal deposits that are sometimes truncated by local intra-Henryhouse unconformities.

SEQUENCE STRATIGRAPHIC MODEL

One of the keys in developing an understanding of a potential pay is to integrate all data from cores, outcrops, samples, logs, etc., into a comprehensive model that can be used to evaluate reservoir characteristics for exploration and exploitation purposes. It is particularly important to understand facies vs. log response, even if it's only qualitatively, to provide a framework for detailed correlation and discovery.

Facies vs. Log Response

In the Chimneyhill through Bois d'Arc section, the most consistent factor which controls log response is shaliness. Fine clastics typically occur in the lower energy environments, such as the open marine and lower subtidal settings. Fine clastics are also found in the upper intertidal to supratidal environments, due to windblown deposition, and they may be winnowed by currents into the low-energy subtidal facies.

A typical sequence or parasequence has a low-energy, high-energy, and low-energy configuration from base to top. The effect is that the upper subtidal to lower intertidal environments have the cleanest (lowest) gamma-ray response and the highest resistivity; the open marine to lower subtidal settings have the poorest (highest) gamma-ray response and the lowest resistivity; and the upper intertidal to supratidal responses are in between, and can be quite "shaly" in areas where there is a concentration of windblown clastics (Fig. 21). Of course, there are many variations and exceptions to those observations; this is especially true in areas of high porosity and high water saturations, where the gamma-ray log will show "clean," but the resistivity will be low.

Unconformities

The pre-Woodford unconformity is the hiatus most commonly associated with the Hunton Group. There are also several intra-Hunton unconformities, such as those found at the Silurian–Devonian contact and the base of the Frisco. This later unconformity is the most important because it represents the period of maximum erosion and truncation of the Hunton, even though the time gap is only 1–2 million years. The amount of erosion is substantiated by the stratigraphic relationship of the Frisco to the underlying units (i.e., Frisco can be found resting unconformably on Chimneyhill to Bois d'Arc strata; Fig. 22).

Correlation

An extensive network of cross sections was built that included strike and dip sections at a density of at least one per township. The area covered by these cross sections includes the Anadarko basin in Texas and Oklahoma, the Arkoma basin in Oklahoma and Arkansas, and central and southern Oklahoma. Although Hunton outcrop nomenclature has been used by others extensively on the surface, with some accuracy, in the subsurface the use of these names has been extensively misapplied and has led to great confusion in external correlations of the Hunton Group. For example, in central Oklahoma and the Anadarko basin, the uppermost porous unit of the Hunton is often named, and correlated with the Bois d'Arc; in fact, however, the Bois d'Arc has been truncated over most of this area, and the uppermost porous unit is usually Henryhouse or Frisco.

Evaluation of these cross sections and integration of core and outcrop data first revealed that there was not a typical Hunton section for correlation purposes; in fact, at least three type logs are required to understand the stratigraphic relationships of the intra-Hunton sequences. Figure 22 shows the correlations and sequence relationships from west to east among the type sections for the Anadarko basin, southern Oklahoma, and the Arkoma basin.

RESERVOIR CHARACTERIZATION

Porosity, and thus reservoir development, is facies dependent in the Hunton, although there are several processes which influence porosity development. An evaluation of the diagenetic history of the Hunton reveals that the two most important processes are dolomitization and dissolution (Fig. 23).

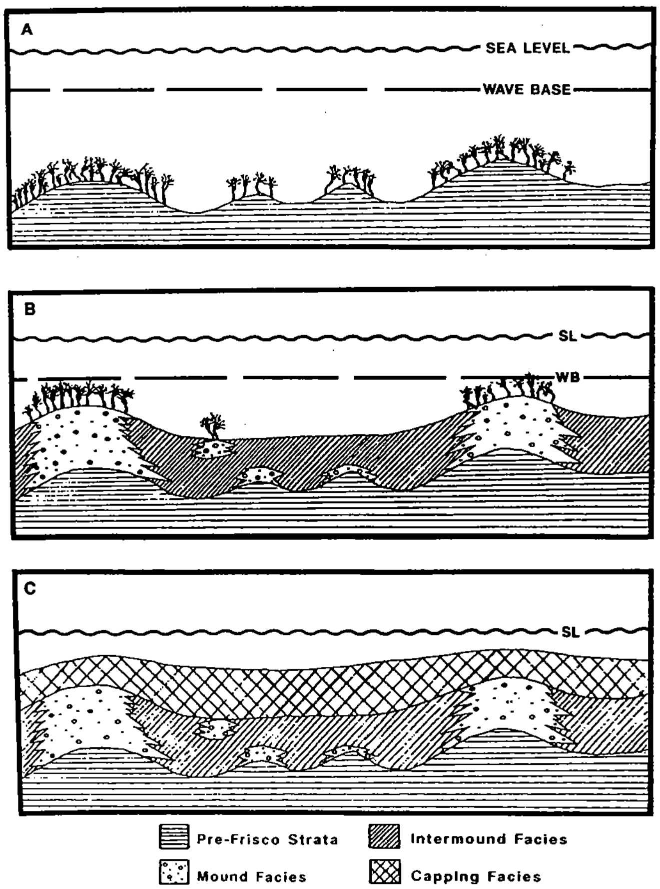

Figure 18. Depositional model for the Frisco Formation.

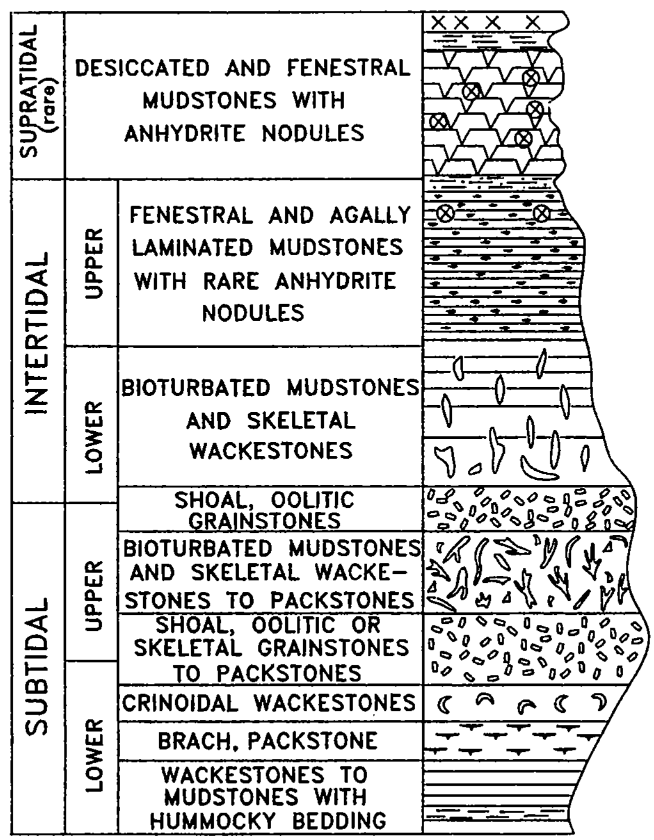

Figure 19. Idealized sequence showing relationship of lithofacies to depositional environments.

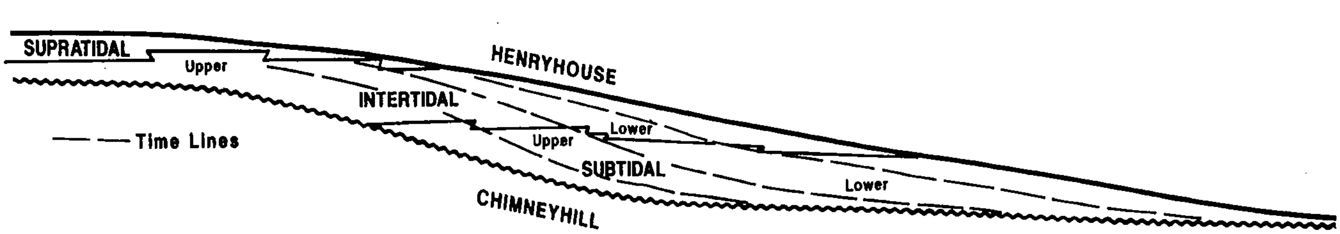

Figure 20. Depositional model showing details of progradation and aggradation.

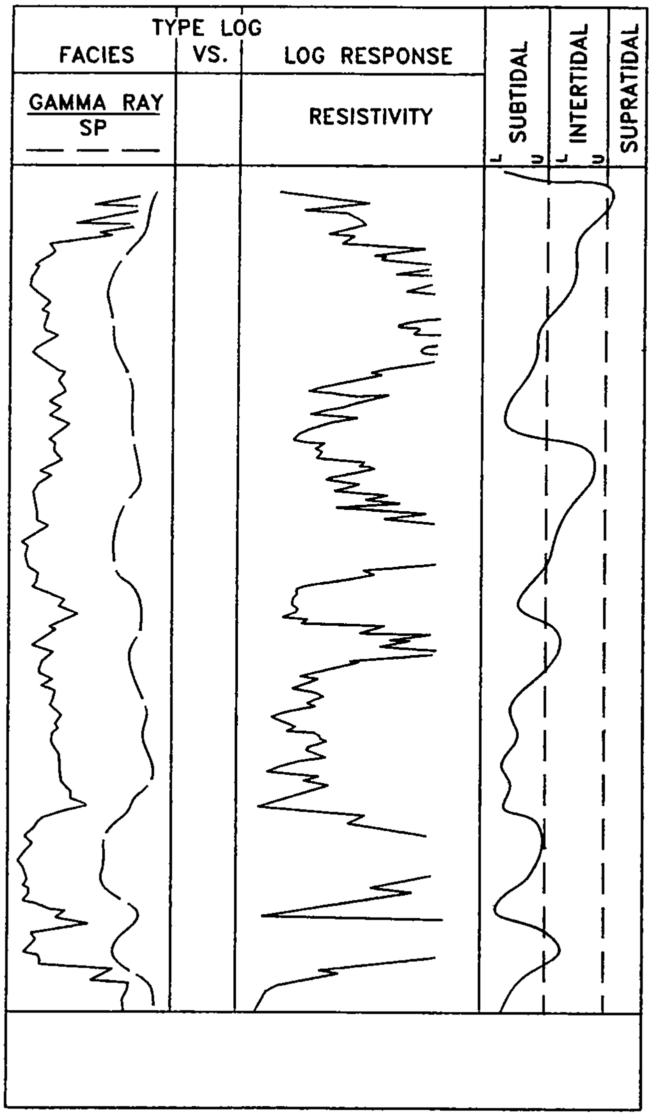

Figure 21. Log showing typical response of SP, gamma-ray, and resistivity curves to depositional environments.

Porosity Development

Significant porosity formation in the Hunton usually occurs in oolitic, dolomitized grainstones (Keel and Henryhouse), and dolomitized, burrowed wacke/packstones (Cochrane, Henryhouse, and Haragan/Bois d'Arc). These units are typically subjected to at least three stages of dolomitization and multiple stages of dissolution in the form of karstification or connate fluids (Manni, 1985) The relationship of dolomitization and facies is shown in Fig. 24.

Porosity development in the Frisco is different than in most other parts of the Hunton; most porosity in the Hunton is related to dolomitization, whereas the Frisco is rarely dolomitized. It typically has inter- and intraparticle porosity combined with moldic to vuggy porosity.

The Penters is unusual in that it has been subjected to intense karstification, typically associated with porosity occlusion; however, the high percentage of chert combined with dolomite develops an extensive fracture system that can drain hydrocarbons from the intercrystalline porosity in the dolomite.

Trap and Geometry

Most of the Hunton fields would be classified as structural/stratigraphic traps, with emphasis on stratigraphic. The most common type of reservoir configuration in the Hunton is where porous facies are truncated by either the pre-Woodford unconformity and/or intra-Hunton unconformities along a structural nose. This is particularly true of most of the Henryhouse section in the Anadarko basin, which develops reservoirs wherein the traps are formed by truncation (by unconformity) and both the seal and the source rock are the Woodford Shale (Fig. 25).

Another common reservoir configuration in the Hunton is trapping by permeability barriers caused by facies changes or karst profiles along structural noses or faults (Fig. 25).

In the Arkoma basin, fractured reservoirs and Hunton fields are more commonly related to structural configuration (e.g., Bonanza field).

CONCLUSIONS

The Hunton Group is a shallow-marine carbonate composed of limestones, dolomites, and calcareous shales that prograded and aggraded onto a ramp. These carbonates are cyclic and can be divided into sequences within formational boundaries that are recognizable in outcrops, cores, and logs. These sequences commonly are separated by disconformities or unconformities and can be correlated regionally. Furthermore, reservoir development in the Hunton Group is facies dependent and reservoir-producing lithofacies can be correlated and mapped (Figs. 26,27).

Use of core, outcrop, and log relationships, integrated with a comprehensive sequence-stratigraphy model, can greatly enhance success for exploration programs or field development.

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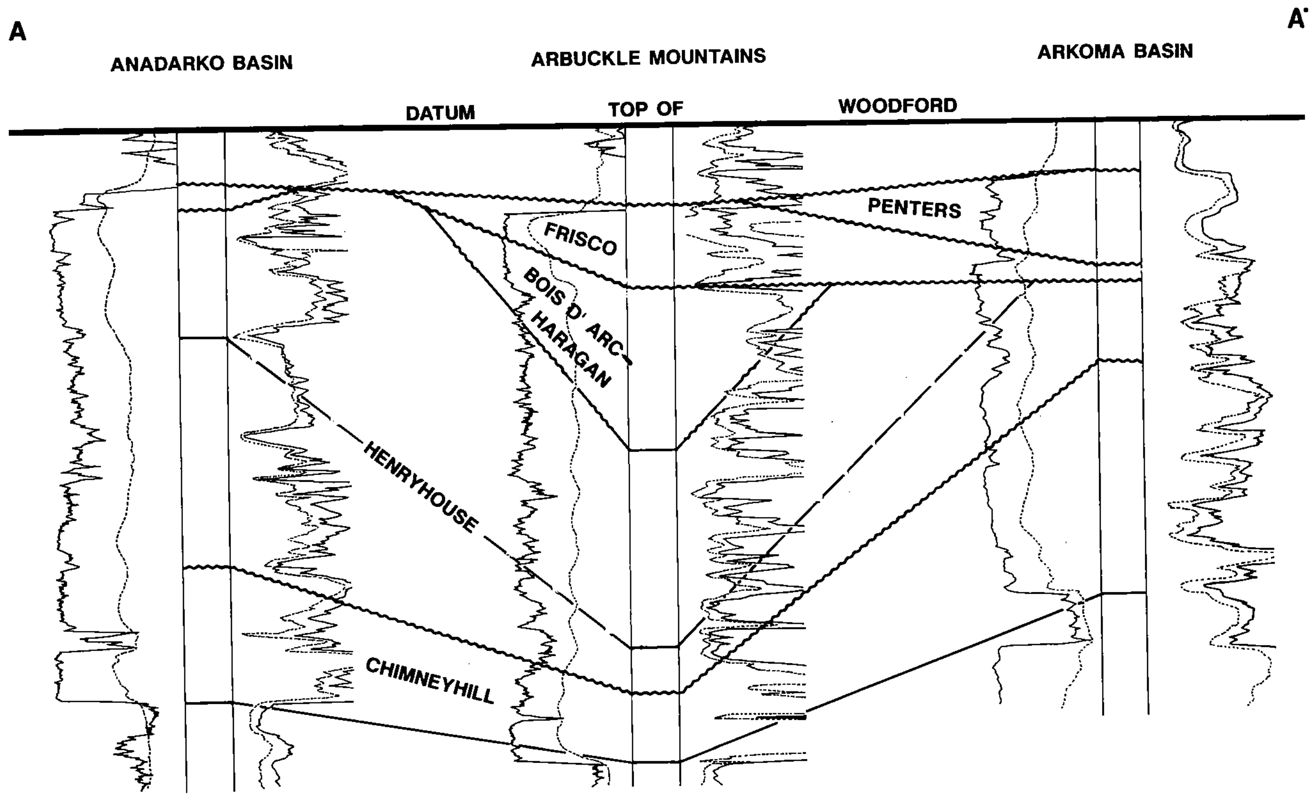

Figure 22. Cross section showing the correlation of type logs for the Anadarko basin, Arbuckle Mountains (southern Oklahoma), and Arkoma basin.

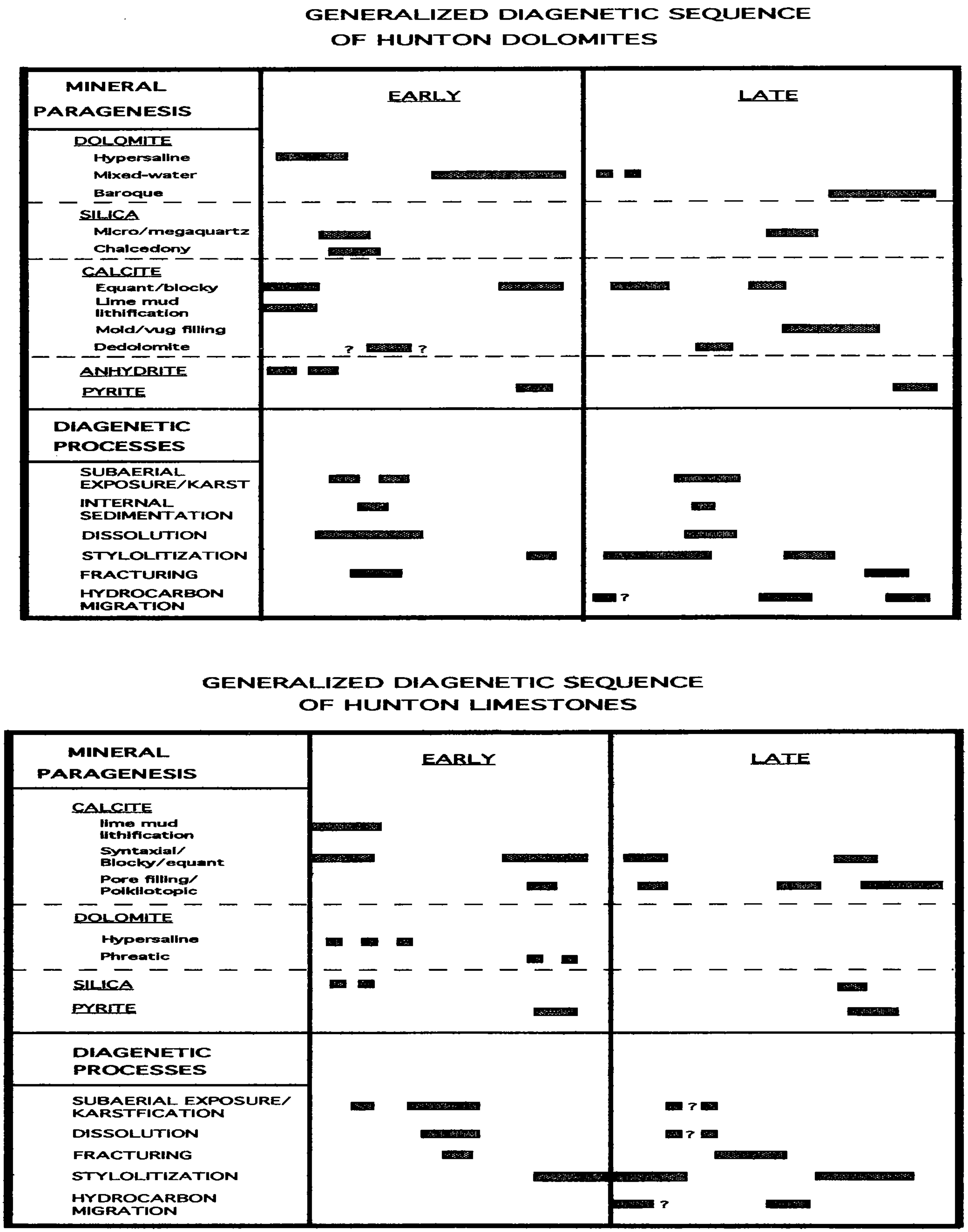

Figure 23. Generalized diagenetic sequence of Hunton dolomites (above) and Hunton limestones (below).

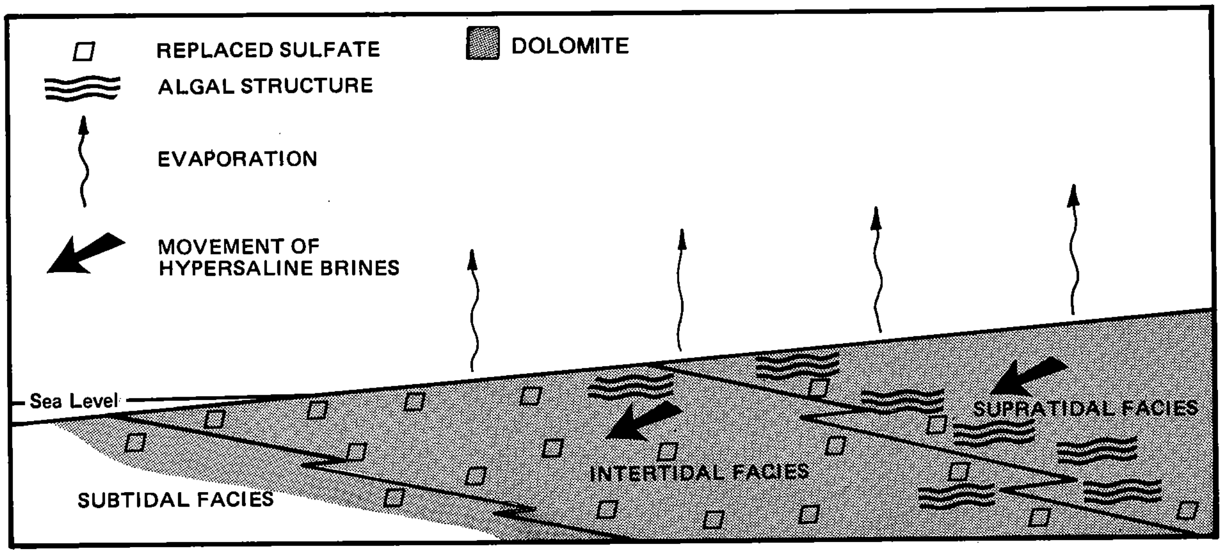

Figure 24. Diagram showing the relationship of dolomitization to depositional facies.

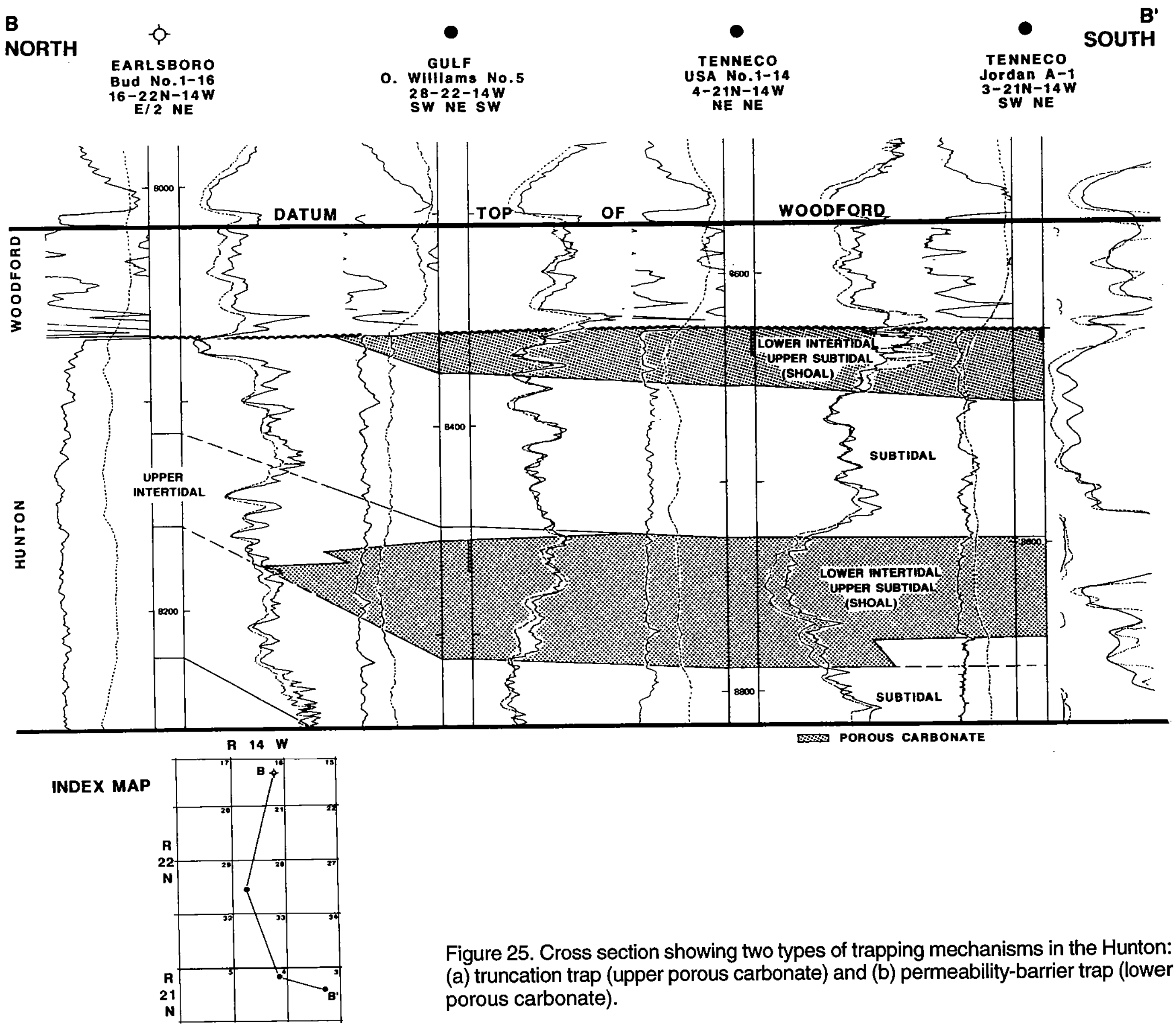

Figure 25. Cross section showing two types of trapping mechanisms in the Hunton: (a) truncation trap (upper porous carbonate) and (b) permeability-barrier trap (lower porous carbonate).

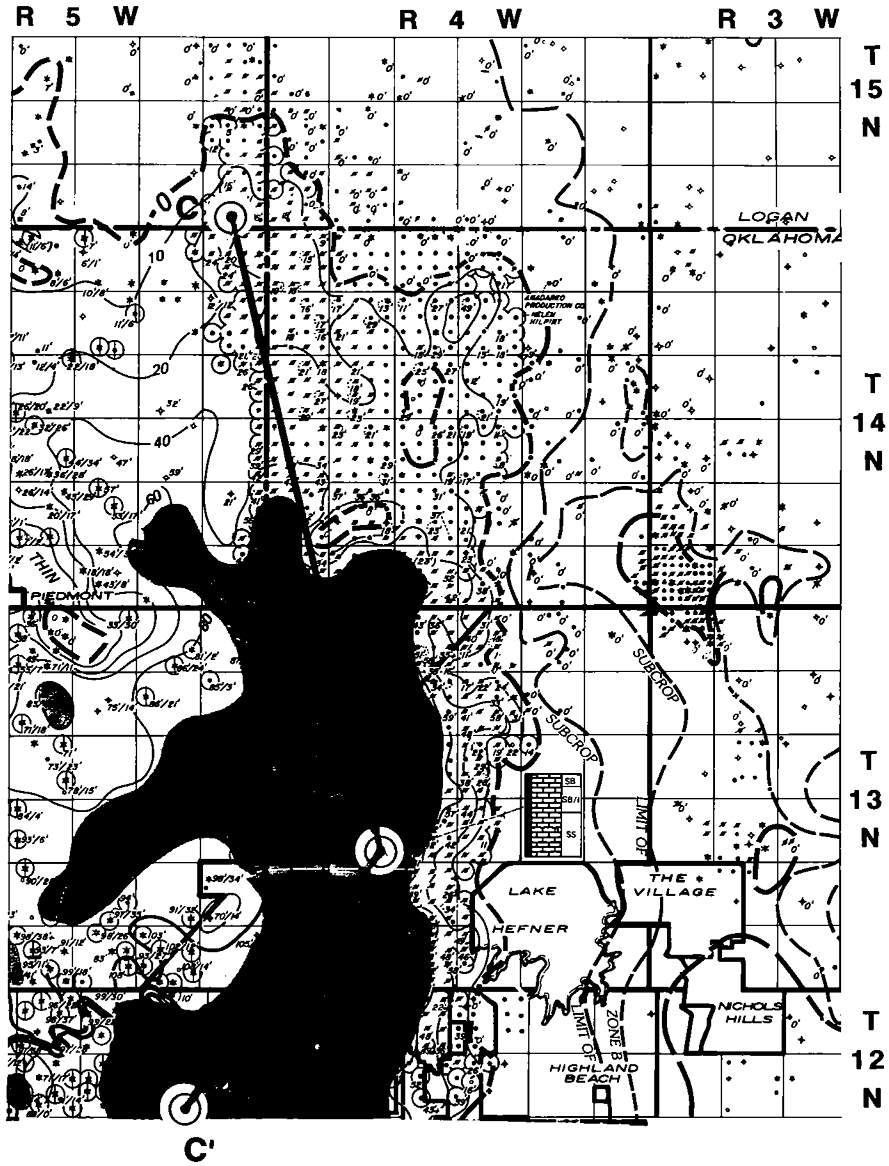

Figure 26. Isopach map of Frisco limestone at West Edmond field (smaller squares are 1 mi²). Cross section C–C′ shown in Figure 27.

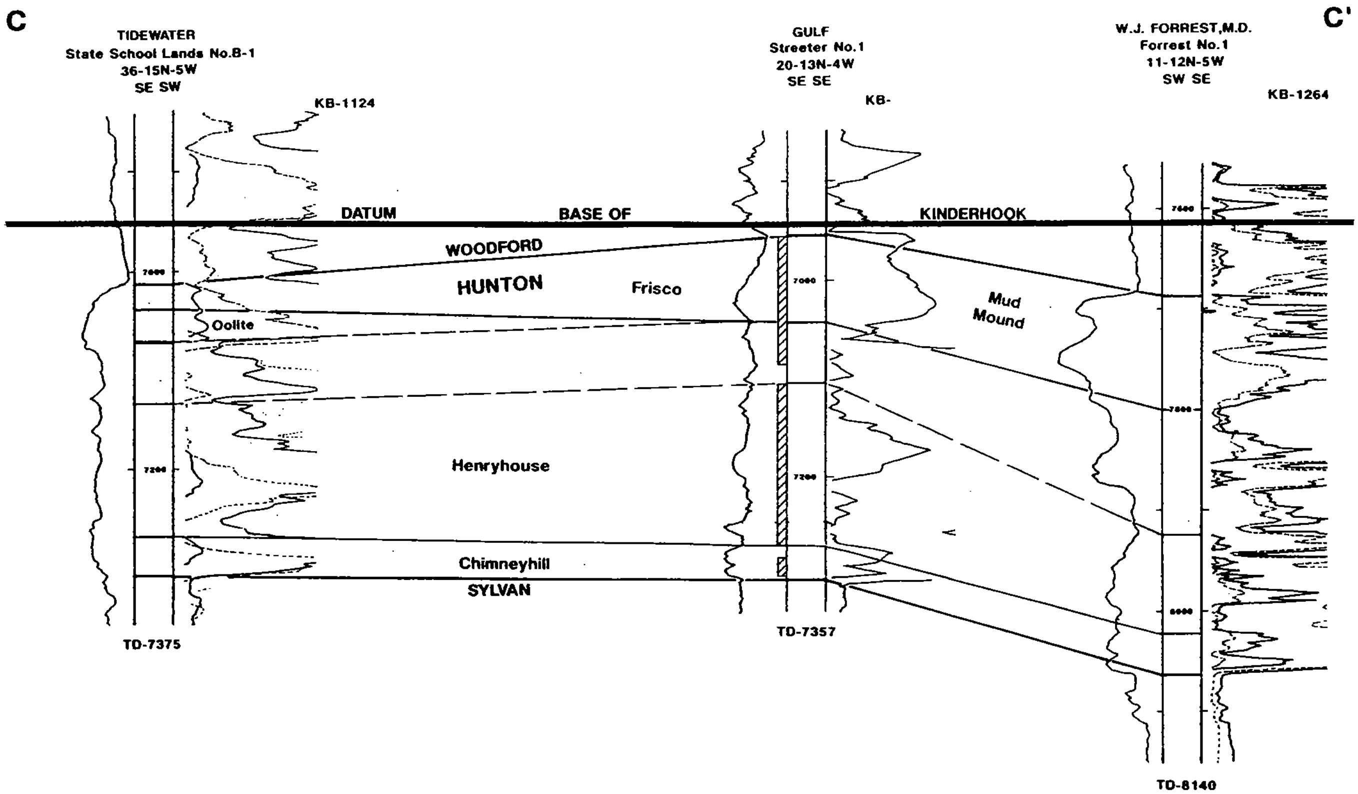

Figure 27. North–south stratigraphic cross section C–C′ showing the relationship of the Frisco reservoir and underlying oolitic Henryhouse reservoir in the West Edmond field (see Fig. 26 for location).

PART 2

Field Trip

Hunton Group Field Trip to the Arbuckle Mountains, Oklahoma

Field-Trip Leaders:

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INTRODUCTION

The Ordovician–Silurian–Devonian Hunton Group has long been a stratigraphic unit of interest in Oklahoma, due to the prolific production of oil and gas associated with its multiple reservoirs. Equivalent zones, such as the Fusselman Formation in the Midland basin, have also been prolific producers.

Hunton strata are part of a large ramp on which were deposited Ordovician through Devonian carbonates that covered much of North America. In the Arbuckle Mountain region, the Ordovician–Silurian section of the Hunton can be divided, in ascending order, into the Chimneyhill Subgroup (which consists of the Ordovician Keel Formation and the Silurian Cochrane and Clarita Formations) and the overlying Henryhouse Formation. The Devonian Hunton is composed of the Haragan, Bois d'Arc, and Frisco Formations (Fig. 1).

The Paleozoic outcrops along I-35 (Appendix) in southern Oklahoma have become classic exposures in a relatively short time and have been the site of numerous field trips. During construction of I-35 in the late 1960s, Dr. Robert O. Fay was commissioned by the U.S. Bureau of Public Roads, through the Oklahoma Geological Survey, to "salvage anything of scientific value and to study and map new exposures of rock" (Fay, 1969,1989). Figure 2 is a stratigraphic section developed by Fay (1989) that shows the stratigraphic position and thickness of all units exposed in the I-35 road cut.

Outside of the Arbuckle Mountains, there are relatively few good outcrops of the Hunton Group in Oklahoma. There are a few interesting outcrops in northeastern Oklahoma on the west flank of the Ozark uplift. In northern Arkansas, however, there are excellent outcrops along the Buffalo and White Rivers on the south flank of the Ozark uplift. One of the primary problems, regarding Hunton depositional facies, that must be understood when viewing outcrops in southern Oklahoma, is that the Hunton

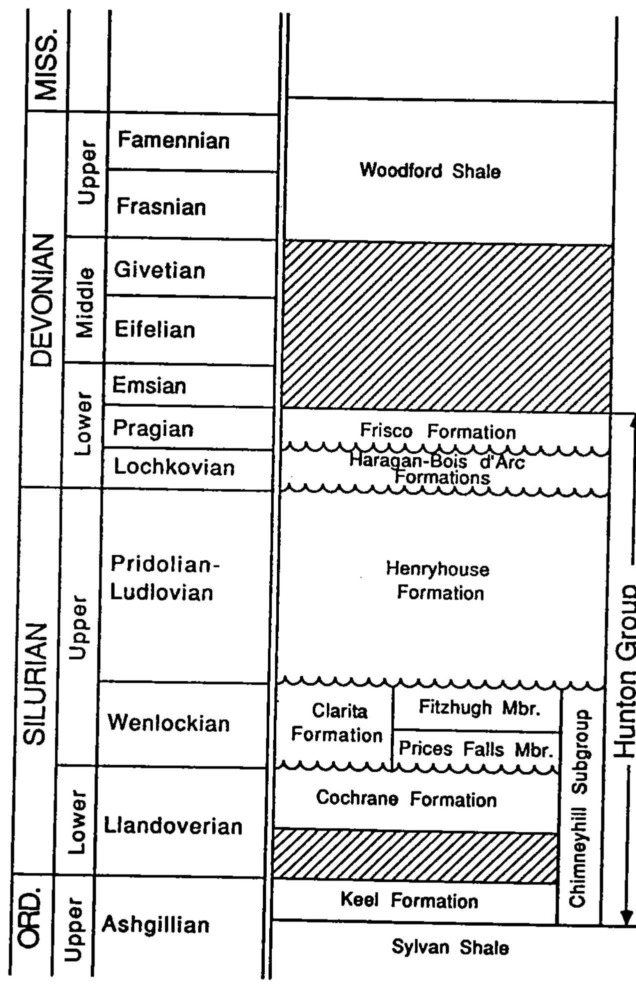


Figure 1. Stratigraphic subdivisions of uppermost Ordovician, Silurian, and Devonian Systems recognized on outcrop in the Arbuckle Mountains and Criner Hills of southern Oklahoma (Barrick and others, 1990).

Al-Shaieb, Zuhair; Fritz, R. D.; Barrick, J. E.; Medlock, P. L.; and Puckette, James, 1993, Hunton Group field trip to the Arbuckle Mountains, Oklahoma, *in* Johnson, K. S. (ed.), Hunton Group core workshop and field trip: Oklahoma Geological Survey Special Publication 93-4, p. 183–212.

STRATIGRAPHIC SECTION EXPOSED ON I-35 THROUGH THE ARBUCKLE ANTICLINE AND MAPPED AREA

SYSTEM	GROUP	GROUP, FORMATION AND MEMBER	MAP SYMBOL	THICKNESS (ft) SOUTH FLANK	NORTH FLANK
QUATERNARY		Alluvium and Terrace deposits	Qat		
		UNCONFORMITY			
PERMIAN - PENNSYLVANIAN		Pontotoc Group *undifferentiated*	Po/ℙp		
		UNCONFORMITY			
PENNSYLVANIAN		Collings Ranch Conglomerate	ℙcr		3000 estimated
		UNCONFORMITY			
MISSISSIPPIAN		Goddard Shale	Mg	2500	
MISSISSIPPIAN		Delaware Creek Shale (formerly "Caney" Shale)	Md	425	
MISSISSIPPIAN		Sycamore Limestone	Ms	358	221
DEVONIAN		Woodford Shale	MDw	290	274
DEVONIAN	Hunton Group (DSOh)	Bois d'Arc Limestone	*	9	19
DEVONIAN	Hunton Group (DSOh)	Haragan Formation	*	25	16
SILURIAN	Hunton Group (DSOh)	Henryhouse Formation	*	191	72
SILURIAN	Hunton Group (DSOh); Chimneyhill Subgroup	Clarita Limestone	*	12	16
SILURIAN	Hunton Group (DSOh); Chimneyhill Subgroup	Cochrane Limestone	*	13	4+
ORDOVICIAN	Hunton Group (DSOh); Chimneyhill Subgroup	Keel Limestone	*		7
ORDOVICIAN		Sylvan Shale	Os	305	275
ORDOVICIAN		Viola Group	Ov	684	710
ORDOVICIAN	Simpson Group	Bromide Formation	Obr	420	346
ORDOVICIAN	Simpson Group	Poolville Limestone Member	*	120	80
ORDOVICIAN	Simpson Group	Mountain Lake Member	*	300	266
ORDOVICIAN	Simpson Group	Tulip Creek Formation	Otc	395	297
ORDOVICIAN	Simpson Group	McLish Formation	Oml	475	397
ORDOVICIAN	Simpson Group	Oil Creek Formation	Ooc	747	
ORDOVICIAN	Simpson Group	Joins Formation	Oj	294	
ORDOVICIAN	Upper Arbuckle Gp	West Spring Creek Formation	Ow	1515	
ORDOVICIAN	Upper Arbuckle Gp	Kindblade Formation	Ok	1410	
ORDOVICIAN	Upper Arbuckle Gp	Cool Creek Formation	Occ	1300	
ORDOVICIAN	Upper Arbuckle Gp	McKenzie Hill Formation	Omh	900	
ORDOVICIAN	Upper Arbuckle Gp	Butterly Dolomite	Ob	297	
CAMBRIAN	Lower Arbuckle Group	Signal Mountain Formation	Єsm	415	
CAMBRIAN	Lower Arbuckle Group	Royer Dolomite	Єry	717	
CAMBRIAN	Lower Arbuckle Group	Fort Sill Limestone	Єfs	155	
CAMBRIAN	Timbered Hills Group	Honey Creek Limestone	Єhc	105	
CAMBRIAN	Timbered Hills Group	Reagan Sandstone	Єr	240	
CAMBRIAN		Colbert Rhyolite	Єc	4500 drilled	7500 estimated

* Formation or member shown only on cross section

Figure 2. Correlation chart for the Arbuckle anticline showing thickness of formations on outcrop (Fay, 1989).

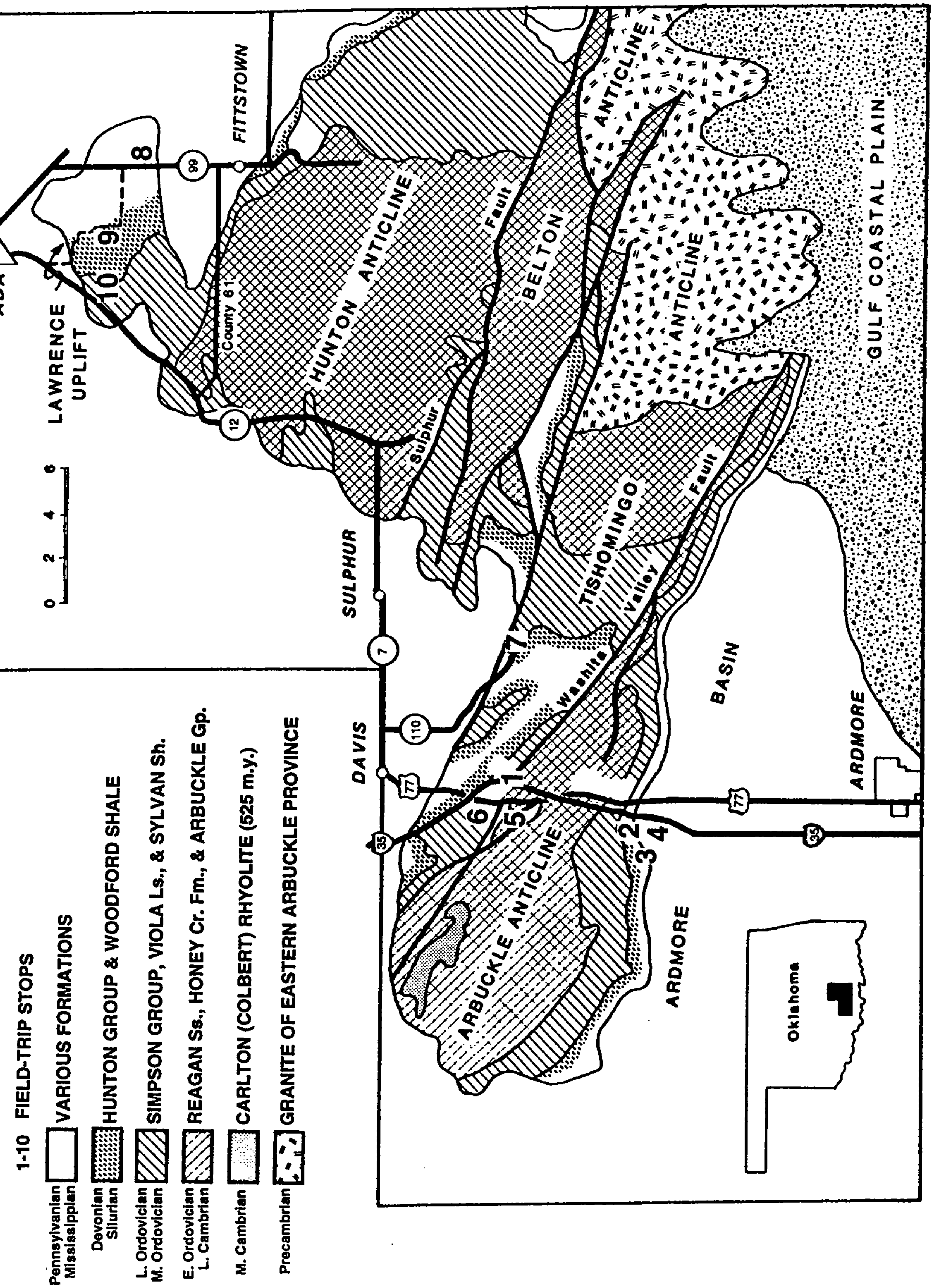


Figure 3. Generalized geologic map of the Arbuckle Mountains showing field-trip stops 1–10 (modified from Ham, 1969).

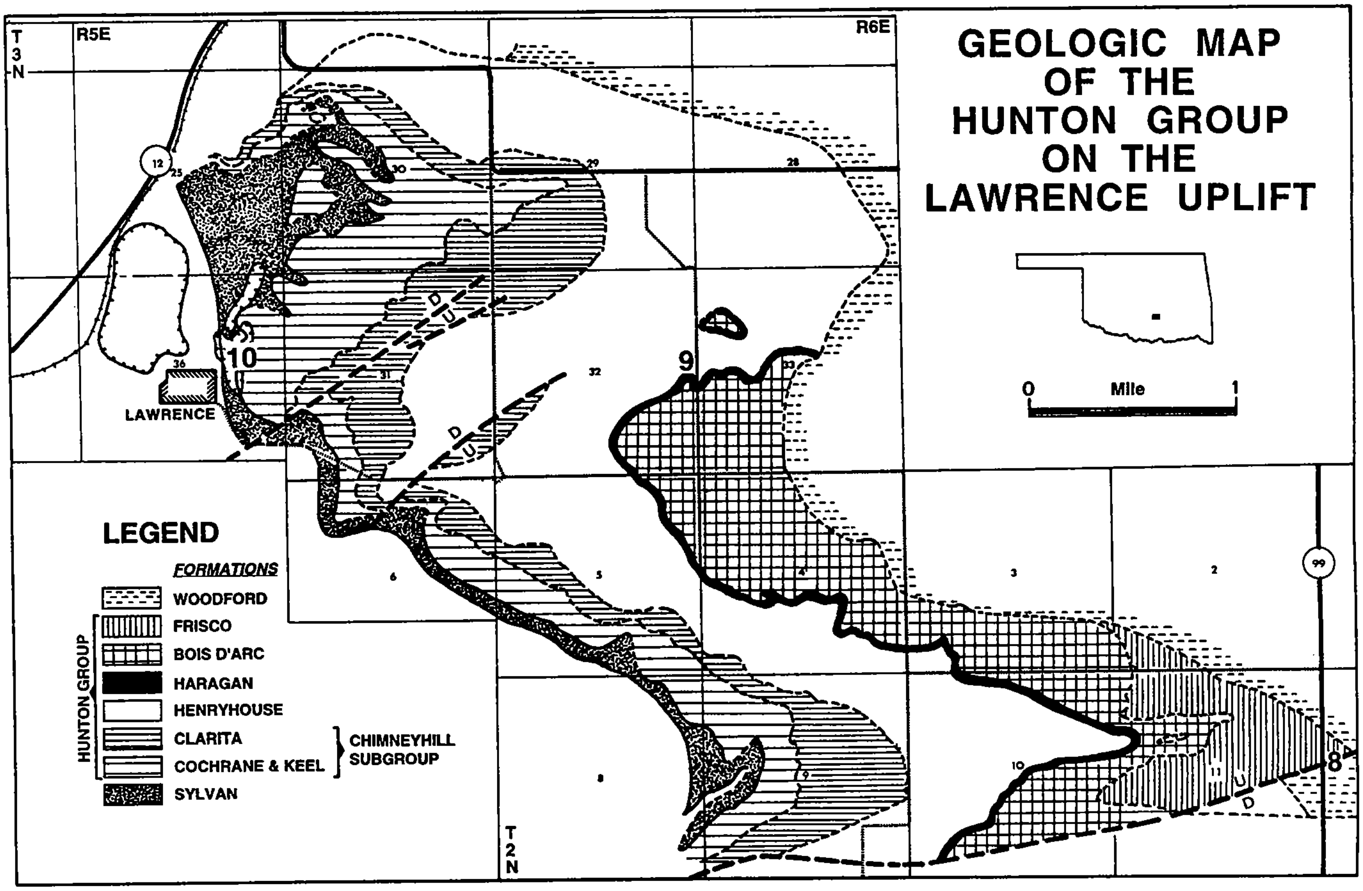


Figure 4. Geologic map of the Hunton Group on the Lawrence uplift showing field-trip stops 8–10 (modified from Amsden, 1960).

was deposited on a ramp, and the facies change significantly updip and downdip along the ramp. Therefore, the facies that we see in cores from the Anadarko basin are usually not represented in the outcrop. This is especially true of the Henryhouse Formation, which contains excellent reservoirs in the lower intertidal/upper subtidal facies on the Anadarko shelf. In contrast, Henryhouse outcrops in the Arbuckle uplift are dominantly thin, argillaceous, lower subtidal facies that are not reservoir quality rocks.

The primary purposes of this brief field excursion are to: (1) study the stratigraphy and post-depositional history of the Hunton Group, (2) examine the cyclic nature of the Hunton carbonates, and (3) examine the relationship of sequence stratigraphy to correlation and reservoir development. The general geology of the Arbuckle Mountains and the planned field-trip stops are shown in Figures 3 and 4. The general itinerary of field-trip stops is given in Table 1, and a detailed road log is given in Table 2.

Table 1. — Proposed Field-Trip Itinerary to Examine the Hunton Group in the Arbuckle Mountains

7:00	Check-out	Leave for Arbuckle Mountains
9:00	Stop 1	**Scenic Turnout — North Flank of Arbuckle Mountains** Arbuckle uplift overview
9:45	Stop 2	**Sylvan Formation–Hunton Group Contact** Chimneyhill Subgroup
10:00	Stop 3	**Silurian–Devonian Contact in the Hunton Group** Henryhouse, Haragan, and Bois d'Arc Formations
10:30	Stop 4	**Woodford Shale** Source rock for the Midcontinent
10:45	Stop 5	**Turner Falls Scenic Overlook (Optional)** Travertine deposition
11:00	Stop 6	**Hunton Group on U.S. Highway 77** Chimneyhill Subgroup, Henryhouse and Haragan Formations
12:00	Lunch	**Lake of the Arbuckles Overlook**
1:00	Stop 7	**Hunton Anticline Quarry** Structure and lithology of the Bois d'Arc Formation
2:30	Stop 8	**Frisco–Bois d'Arc Contact on Bois d'Arc Creek** Facies and depositional environment of Frisco mud mounds
3:30	Stop 9	**Road Cut Near the South Fork of Jackfork Creek** Henryhouse facies
4:00	Stop 10	**Hunton Group in the Lawrence Quarry South of Ada** Chimneyhill Subgroup and Sylvan Shale

Table 2. — Road Log for the Hunton Group Field Trip to the Arbuckle Mountains

Cumulative mileage	*Interval*	
0.0*	0.0	Proceed west on Lindsey St.
2.0	2.0	Intersection of Lindsey St. and Interstate 35 (I-35). Proceed south on I-35.
3.0	1.0	Bridge over the Canadian River.
31.9	28.9	Paoli exit. Continue south on I-35.
38.6	6.7	Pauls Valley exit. Continue on I-35.
47.5	8.9	Wynnewood exit. Continue south on I-35.
53.5	6.0	Rest area.
55.9	2.4	Junction of I-35 and State Highway 7 to Davis and Sulphur. Continue south on I-35.
56.3	0.4	Entering Murray County.
58.4	2.1	Weigh station.
60.6	2.2	Junction of I-35 and U.S. Highway 77 (US-77). Proceed south on I-35.
61.5	0.9	Road cut through the Viola and Simpson Groups.
62.1	0.6	Road cut through the Collings Ranch Conglomerate.
62.7	0.6	***Stop 1*** — Scenic turnout. Arbuckle Uplift Overview.
64.4	1.7	Junction of I-35 and US-77. Continue south on I-35.
65.5	1.1	Scenic turnout.
65.9	0.4	Road cut through stromatolite-bearing Cool Creek Limestone.
66.2	0.3	Carter County line.
66.9	0.7	Road cut in the Oil Creek Formation.
67.4	0.5	Road cut in the Viola Formation.
67.8	0.4	***Stops 2, 3, and 4*** — Sylvan Shale, Hunton Group, and Woodford and Sycamore Formations. South-dipping beds of the Sylvan through Sycamore Formations are exposed on the west side of the southbound lane of I-35. ***Stop 2*** is the Sylvan Shale–Hunton Group contact and the Chimneyhill Subgroup. ***Stop 3*** is the Silurian–Devonian contact in the Hunton Group (Henryhouse, Haragan, and Bois d'Arc Formations). ***Stop 4*** is the Hunton Group–Woodford Shale contact (Woodford Shale and overlying Sycamore Limestone).
69.0	1.2	Exit 42 at junction of I-35 and State Highway 53. Leave I-35.
69.1	0.1	Stop sign. Turn left, proceed under overpass, turn left, and continue north on I-35.

*Mileage begins at the intersection of Asp and Lindsey Streets on the campus of The University of Oklahoma, in Norman.

TABLE 2. — *Continued*

Cumulative mileage	*Interval*	
70.5	1.4	Milepost 44. Road cut in the Sycamore Limestone.
70.9	0.4	Road cut in the Viola Group limestone.
71.5	0.6	Road cut in the lower Simpson Group.
72.3	0.8	Road cut in the Arbuckle Group.
73.7	1.4	Junction of I-35 and US-77. Exit I-35 and turn left to go north on US-77.
75.7	2.0	***Stop 5*** — Turner Falls Overlook. Travertine deposit in Honey Creek Limestone and stromatolite-bearing Cool Creek Formation.
75.9	0.2	Fault contact between Pennsylvanian Collings Ranch Conglomerate and the Ordovician Cool Creek Formation.
76.6	0.7	Entrance to Turner Falls Park on left.
77.7	1.1	***Stop 6*** — Hunton Group. Exposure on west side of US-77 represents a nearly complete section of strata from the Upper Ordovician Sylvan Shale (south end of road cut) to Upper Devonian Woodford Shale (north end of road cut).
78.1	0.4	Washita River crossing.
81.4	3.3	Intersection with State Highway 7 (OK-7). Turn right and proceed east on OK-7.
83.5	2.1	East through the town of Davis to State Highway 110 (OK-110) intersection. Exit OK-7 and take OK-110 south toward Dougherty.
86.5	3.0	Curve in road.
87.6	1.1	Intersection; continue left on OK-110.
90.5	2.9	Exit left (northeast) to Arbuckle Dam.
91.2	0.7	Lunch site; Lake of the Arbuckles overlook.
91.9	0.7	Return to OK-110 and turn left.
92.2	0.3	Make a left turn on "gravel road."
92.4	0.2	Rock Creek Bridge.
92.7	0.3	Turn right to go to the Hunton quarry. *CAUTION*: rough road.
93.1	0.4	***Stop 7*** — Hunton anticline: exposures of the Woodford Shale, Bois d'Arc Limestone, and the Haragan Formation.
101.3	8.2	Return to OK-7 and turn east (right).
106.6	5.3	Entering Sulphur, Oklahoma.
108.2	1.6	Artesian spring on right side of the road.
108.5	0.3	Intersection with U.S. Highway 177 (US-177). Turn left (north) on US-177.
108.8	0.3	Intersection with OK-7. Turn right and continue east on OK-7.
115.2	6.4	Intersection with State Highway 1 (OK-1). Turn left on OK-1 and proceed north.
121.7	6.5	Entering Pontotoc County; continue on OK-1.
123.7	2.0	Roff, Oklahoma. Immediately north-east of town is a glass-sand pit in the Simpson Oil Creek Formation.
125.0	1.3	Intersection with County Highway 61. Take County Highway 61 east.
136.4	11.4	Intersection with State Highway 99 (OK-99). Turn left and proceed north on OK-99.
139.0	2.6	***Stop 8*** — Frisco-Bois d'Arc contact on Bois d'Arc Creek and the mound facies of the Frisco Formations. Faulted contact between the Woodford Shale and Frisco Formation. Proceed north on OK-99.
139.5	0.5	Intersection with country road, turn west (left).
141.2	1.7	Bend in road, Haragan outcrop on left.
141.6	0.4	Intersection. Turn right (north).
142.0	0.4	Proceeding north: Woodford Shale on right side of the road.
142.6	0.6	Intersection. Turn west (left).
143.6	1.0	Intersection with locked gates. Turn right and proceed north. Driving on Bois d'Arc Limestone.
144.1	0.5	***Stop 9*** — Henryhouse exposed at road cut near the South Fork of Jackfork Creek.
144.2	0.1	South Fork of Jackfork Creek.
144.3	0.1	Crest of hill. Henryhouse is exposed in ditches on both sides of the road.
144.5	0.2	North Fork of Jackfork Creek.
144.6	0.1	Road turns west and then northwest.
145.0	0.4	Road turns north.
145.3	0.3	Intersection. Turn left (west).
146.0	0.7	Intersection. Turn right and proceed north.
146.5	0.5	Road turns west.
147.5	1.0	Road turns north.
148.0	0.5	Go under conveyer that carries rock from Lawrence Quarry to Ideal Cement plant in Ada.
148.1	0.1	Intersection of OK-1. Turn left and proceed southwest on OK-1.
149.5	1.4	Turn left into Lawrence Quarry entrance.
		Entering Lawrence Quarry. *CAUTION*: Watch out for large trucks.
150.7	1.2	***Stop 10*** — Exposed contact of the Chimneyhill Subgroup and the underlying Sylvan Shale. Excellent exposures of the Keel (including the Ideal Quarry Member) and Cochrane Formations of the Chimneyhill Subgroup.
		End of trip. Return to Norman via OK-1 to Ada and then OK-19 to Stratford, Pauls Valley, and I-35.

STOP 1

SCENIC TURNOUT — NORTH FLANK OF THE ARBUCKLE MOUNTAINS

Near center sec. 31, T. 1 S., R. 2 E.
Murray County, Oklahoma
(Scenic turnout, south-bound lane on I-35)

Arbuckle Uplift Overview

The Arbuckle Mountains have been of interest to geologists because of their excellent outcrops of Precambrian and Cambrian igneous rocks and of Paleozoic carbonate and clastic rocks. This area has been intensively studied for years, and many publications are available that address various aspects of its geological features. The following description for Stop 1 is condensed from Ham (1969).

The Arbuckle Mountains, a large inlier of folded and faulted Paleozoic and Precambrian rocks, are covered on the east, north, and west by gently westward-dipping Pennsylvanian and Permian strata, and on the south by gently southward-dipping Early Cretaceous sediments of the Gulf Coastal Plain (Figs. 3,5).

The Arbuckle Mountains consist of 11,000 ft of fossiliferous Late Cambrian through Devonian strata that constitute the greatest area of exposure of this sequence in all the Midcontinent region. Reference to the Arbuckle uplift outcrops as the Arbuckle "Mountains" is misleading because ~80% of the area consists of gently rolling plains. Only in the western area—that of the Arbuckle anticline—is the topographic relief sufficient to evoke comment.

The geology is characterized mostly by outcrops of carbonate rocks. Immediately to the east begins the 200-mi-long exposure of the Ouachita Mountains, principally a flysch sequence that is quite unlike the Arbuckles in stratigraphic and structural development; and 100 mi to the west are the Wichita Mountains, unlike either the Arbuckles or Ouachitas and characterized chiefly by extensive outcrops of Cambrian (not Precambrian) igneous rocks. The primary emphasis of Arbuckle Mountains geology lies in its early Paleozoic carbonates and late Paleozoic clastics, deposited partly upon a craton of Precambrian granites and partly in an adjoining geosynclinal basin, the whole welded by Pennsylvanian orogeny and epeirogeny into a single tectonic/geologic unit.

Rocks of Pennsylvanian age crop out around most of the Arbuckle Mountains, and are preserved within them in the Mill Creek syncline, Wapanucka syncline, and Franks graben. Dornick Hills (Wapanucka and Atoka) strata within and adjoining the Arbuckle Mountains generally are non-conglomeratic, but the Desmoinesian and younger Pennsylvanian rocks are conglomerate-bearing and record the beginning and close of mountain building in the Arbuckle region.

The four principal conglomerate sequences within, and contiguous with, the Arbuckle Mountains are the "Franks," Deese, Collings Ranch, and Vanoss conglomerates. All but the Vanoss, the youngest, are preserved in synclinal grabens and are moderately to strongly folded and faulted.

Clearly, the Deese and "Franks" contain erosional products derived from the first great period of uplift in the Arbuckle Mountains, which began as broad domal folding of the Hunton anticline in early Desmoinesian (early Deese) time. It also is clear that these conglomerates were closely folded, locally overturned, and faulted by later Pennsylvanian orogeny. This later deformation, to which the name Arbuckle orogeny has long been applied, produced the structurally complex Arbuckle anticline and the major folds of the Ardmore basin; it was certainly the most intense deformation to affect the Arbuckle Mountain region (Fig. 6). The folding can be dated from the evidence seen in the Ardmore basin as post-Hoxbar, from the Mill Creek syncline and Lake Classen area as post-Deese (and possibly post-Hoxbar), and from the Franks graben as post-Francis (middle Missourian).

From a comparison with widely distributed marine rocks north of the mountains, the date of the Arbuckle orogeny can be correlated with the unconformity at the base of the Ada Formation, in the middle part of the Virgilian Series. Conglomerates of the Ada Formation consist dominantly of limestone pebbles from the Arbuckle Group and have been derived almost exclusively by erosion of the Arbuckle anticline. Rocks of the Ada Conglomerate are very similar to those of the Collings Ranch Conglomerate at its type locality in the Arbuckle Mountains, where it is unconformable upon vertically dipping beds of the Arbuckle anticline. This suggests that the Collings Ranch and Ada Conglomerates are equivalents and that the strongest pulse of Arbuckle orogeny was mid-Virgilian.

The Collings Ranch and Vanoss Conglomerates rest with pronounced angular unconformity on older rocks that were steeply folded during culmination of the Arbuckle orogeny. These conglomerates, resulting from this uplift and deposited peripheral to the Arbuckle Mountains, are exposed now only around the Arbuckle anticline and northward into the Sulphur area. Most of these coarse clastic sediments were derived from the Arbuckle and Tishomingo anticlines.

A late phase of Arbuckle orogeny, chiefly faulting and uplift without strong folding, came in the last stage of deposition of the Collings Ranch strata. At this time the conglomerate was folded and faulted into a graben. Some of the faults bordering the graben extend northwestward to the edge of the mountains, where they pass underneath the Vanoss Conglomerate, thereby showing the Vanoss to be younger than the Collings Ranch.

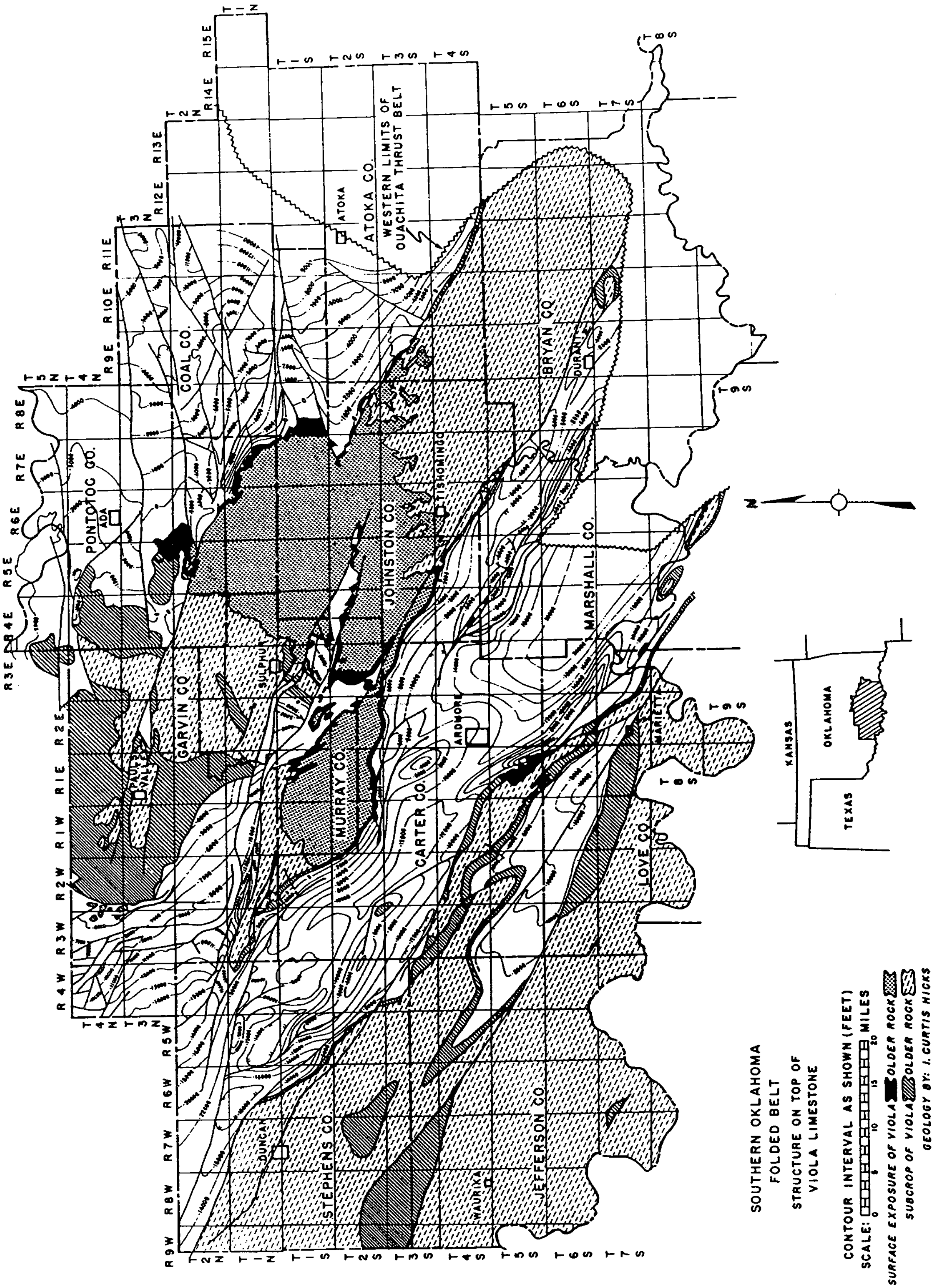


Figure 5. Regional tectonic map of the area around the Arbuckle Mountains (Hicks, 1971).

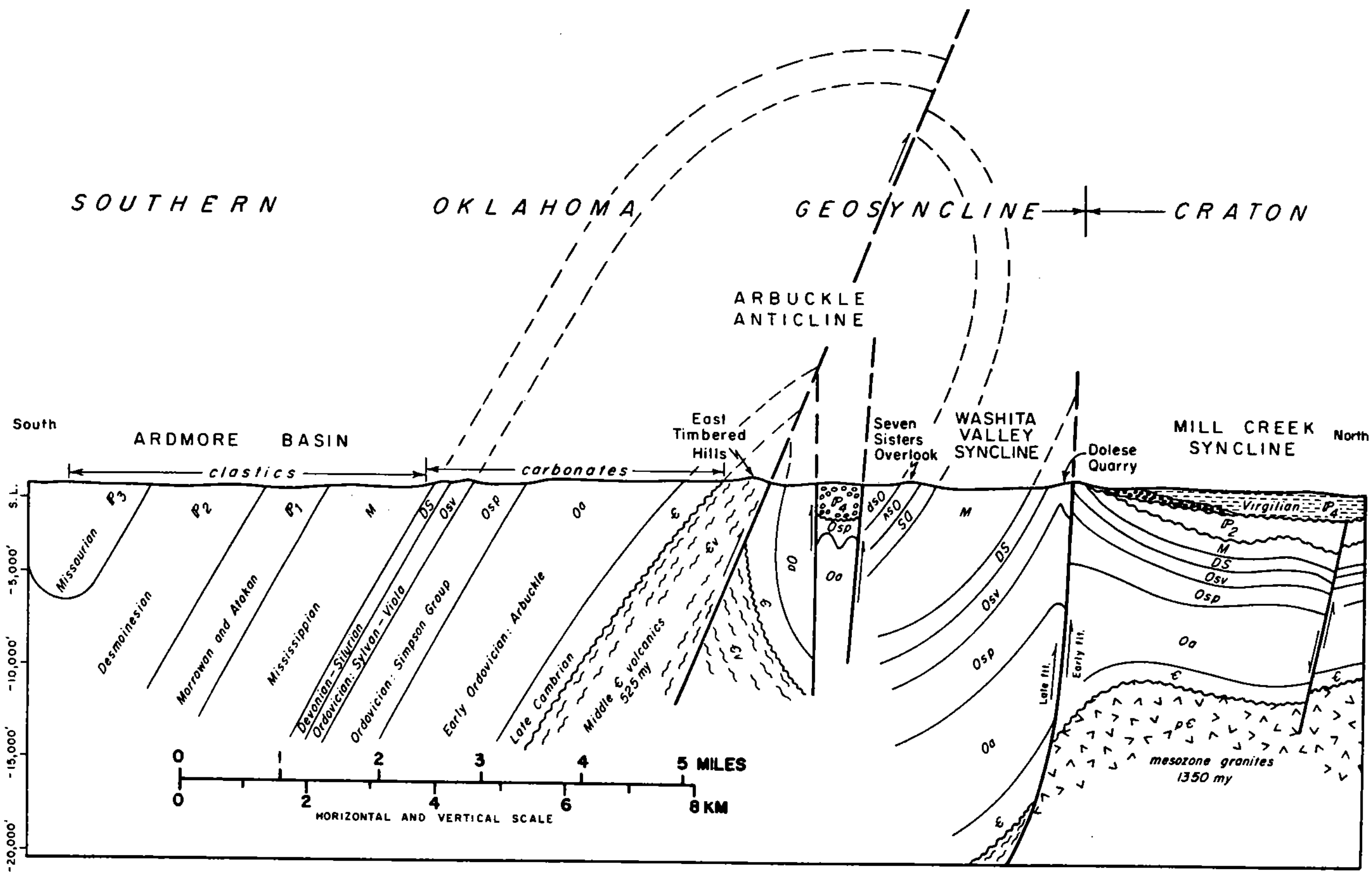


Figure 6. Structural cross section through the Arbuckle anticline along I-35 (modified from Ham, 1969).

STOP 2

SYLVAN FORMATION–HUNTON GROUP CONTACT

SE¼ sec. 25, T. 2 S., R. 1 E.
Carter County, Oklahoma
(Marker No. 4 — 3 ft above base of Cochrane Formation)

Chimneyhill Subgroup

The Hunton Group on the south flank of the Arbuckle anticline is found in two major exposures. The first (northernmost) consists of the Clarita and Cochrane Formations of the Chimneyhill Subgroup. The second is composed of the Henryhouse and Haragan–Bois d'Arc Formations. An important unconformity separates the Silurian from the Devonian at this locality. The Hunton is overlain by a thick section of Woodford Shale, which is overlain by the Sycamore Limestone. Both the Woodford and Sycamore are heavily fractured at this locality.

The Chimneyhill overlies at least 300 ft of the Ordovician Sylvan Shale at this locality. The Sylvan is typically greenish gray and is mostly covered at this site. The relatively soft Sylvan generally forms valleys in its outcrops, and samples of somewhat-weathered Sylvan can be observed in the creek bed immediately below the prominent carbonates of the overlying Chimneyhill Subgroup.

Stratigraphy and Lithofacies

The Chimneyhill, which typically forms a resistant ridge at the base of the Hunton on outcrop, is rather thin (<25 ft) on the south flank of the Arbuckle uplift. In the Anadarko basin to the north the Chimneyhill is often in excess of 100 ft thick.

There is no evidence of the Ordovician-age Keel oolite at this locality; rather, the base of the Hunton is represented by the Cochrane Formation, which is composed of a light-gray to tan, fine- to medium-grained, fossiliferous mudstone/wackestone with a few intervals of fossiliferous packstone/grainstone. Manni (1985) proposed that the Cochrane was deposited on a "ridge and swale" topography developed on a carbonate ramp with an extremely gentle slope. Glauconite indicates a quiet environment (swale), very slow deposition, and locally oxidizing and reducing conditions. Since glauconite is present mostly in the Cochrane, the iron was available during Cochrane deposition and not during Clarita time (the Clarita has little or no glauconite). The Cochrane is approximately 12–15 ft thick at this location. In the subsurface of the Anadarko basin, it can be very porous and is an extensive aquifer system with porous intervals >50 ft thick. Unfortunately, due to the extensive porosity and permeability network, few traps are formed, except along truncation boundaries or on anticlinal structures.

The Clarita is a light-gray to light-tan, fine-grained to coarse-grained, glauconitic pelmatazoan packstone/grainstone. The most distinctive features of the Clarita are the abundant crinoid fragments and pinkish color. The Clarita is typically a poor-quality reservoir in the Anadarko basin, although there is some petroleum production related to fractures in both the Anadarko and Arkoma basins.

STOP 3

SILURIAN–DEVONIAN CONTACT IN THE HUNTON GROUP

SE¼ sec. 25, T. 2 S., R. 1 E.
Carter County, Oklahoma
(Marker No. 3 — 98 ft below top of Bois d'Arc Formation)

Henryhouse, Haragan, and Bois d'Arc Formations

The Henryhouse–Haragan–Bois d'Arc section is nearly 225 ft thick at this locality (Fig. 7). Approximately 120 ft of Henryhouse is exposed, with the lower Henryhouse covered down to the top of the Chimneyhill Subgroup.

Stratigraphy and Lithofacies

The Henryhouse is represented by marly, open-marine to shallow-subtidal limestones and shales (Fig. 8). The Henryhouse section is typically a gray to tan, fine-grained, thin-bedded fossiliferous wackestone to packstone (and sometimes grainstone) that is interbedded with calcitic tan to gray shale. The lower 70 ft of the Henryhouse is covered, but based on other outcrops in the area it is suspected to be dominantly shale and mudstones.

The Haragan–Bois d'Arc section is similar, except that the top 10 ft of the 35 ft of strata exposed is a resistant, tan to bluish-gray, cherty, fossiliferous wackestone to grainstone. These uppermost beds are thicker than those of the Henryhouse and have less shale interbeds, although there are some distinctive lemon-yellow shale partings.

Biostratigraphy

The Upper Silurian Henryhouse Formation is similar in lithofacies and biofacies to the overlying Lower Devonian Haragan Formation, and the boundary between these units is based largely on differences in the shelly faunas. At this outcrop, ~65 ft of the marlstones of the upper Henryhouse and lower Haragan Formations are well exposed, and, upon examination, the problem in precisely locating the contact can be fully appreciated. The most distinctive lithologic feature of the lowermost Haragan in the central Arbuckles, is a unit of skeletal wackestones and packstones that generally forms a low ridge in the poorly exposed marlstone sections; it is 21–28 ft below the top of the exposed Hunton. Based on brachiopods, the Henryhouse–Haragan contact is ~29.5 ft below the top of the Hunton. A specimen of the Haragan species *Meristella atoka* was recovered 28.5–29 ft below the top, and the Henryhouse form *"Lissatrypa" henryhousensis* was collected 30.8–31.2 ft below the top.

Detailed sampling for conodonts across the boundary at this section constrains the magnitude of the hiatus between the Henryhouse and Haragan, but the faunas are not sufficiently distinct to precisely locate the boundary (Barrick and Klapper, 1992). The latest Silurian (late Pridolian) conodont, *Oulodus elegans detorta*, occurs as much as 30.3 ft below the top of the Hunton. This species ranges to the top of the Silurian at the Devonian stratotype at Klonk, in the former Czechoslovakia. The first definitive Devonian species, *Icriodus postwoschmidti*, appears 28.5 ft below top of the Hunton. *Icriodus postwoschmidti* and two other species of the Haragan conodont fauna, *I. eolatericrescens* and *Pedavis biexoramus*, characterize the Early Devonian (early Lochkovian) *eurekaensis* Zone. The *eurekaensis* zone, known from the Cordilleran region of North America, is the second conodont zone above the base of the Devonian.

The same change in conodont faunas occurs across the Henryhouse–Haragan boundary elsewhere in the central Arbuckle Mountains, as well as in sections on the Lawrence uplift (Barrick and Klapper, 1992). The maximum interval of time that can be missing at the boundary includes that represented by the basal Devonian *woschmidti* Zone, and possibly an equally small part of the latest Silurian. Given the relatively brief interval of time missing at the Henryhouse–Haragan boundary, the dramatic change in the benthic fauna, including brachiopods, ostracodes, and trilobites, is even more remarkable (Amsden, 1988).

Cyclicity and Log-Facies Relationships

The Henryhouse and Haragan–Bois d'Arc section consists of several packages of carbonate sequences. Each sequence contains several shallowing-upward cycles, or parasequences, that are typically composed of mudstones and wackestones at the base to packstones and grainstones at the top. The beds also become thicker at the top of each cycle and are thinner and finer grained toward the base. These parasequences range from 7 to 25 ft thick in the Henryhouse and Haragan–Bois d'Arc.

At this stop, individual parasequences can be recognized

Figure 7. Henryhouse–Haragan/Bois d'Arc outcrop on the south flank of the Arbuckle anticline looking toward the west from the west lane of I-35 (Stop 3; outcrop is ~200 ft long).

by their weathering profile. The coarser-grained (shallow, high-energy) carbonates are more resistant and form ridges on the land surface. The finer-grained (deeper, low-energy) rocks form the soil-covered areas between the ridges.

A measured section produced vertically by detailed logging, with facies logged in the left-hand column and relative sea level and energy on the right, produces a graphic representation of the outcrop which is similar to a gamma-ray/spontaneous-potential electrical log and responds to the same basic parameters, primarily shaliness (Fig. 9). These measured sections, or "pseudo-E-logs," are typically logged at 1″ = 40′ (2.5″ scale) or 1′ = 100′ (1″ scale). This type of log can be correlated to electric logs with some degree of success.

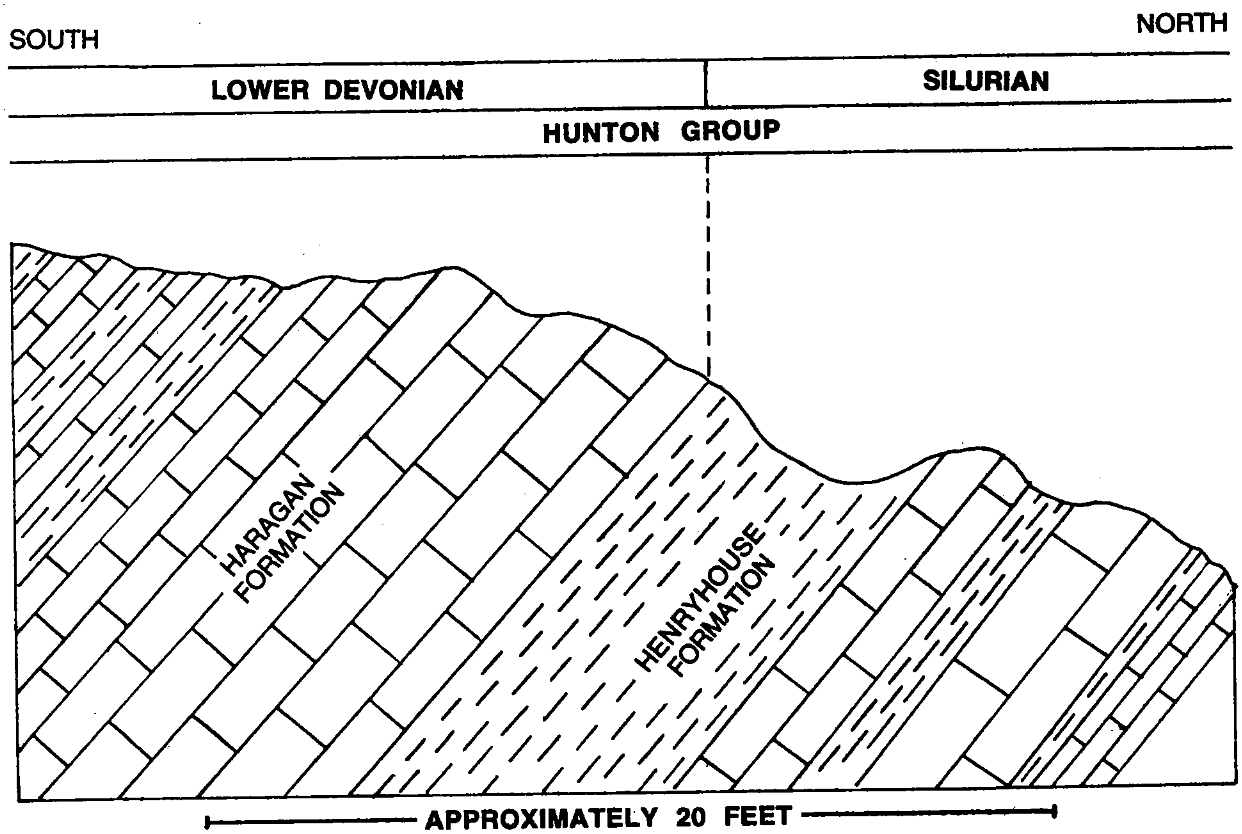


Figure 8. Measured section showing the weathering profile of the Henryhouse and Haragan parasequences on the south flank of the Arbuckle anticline (Stop 3).

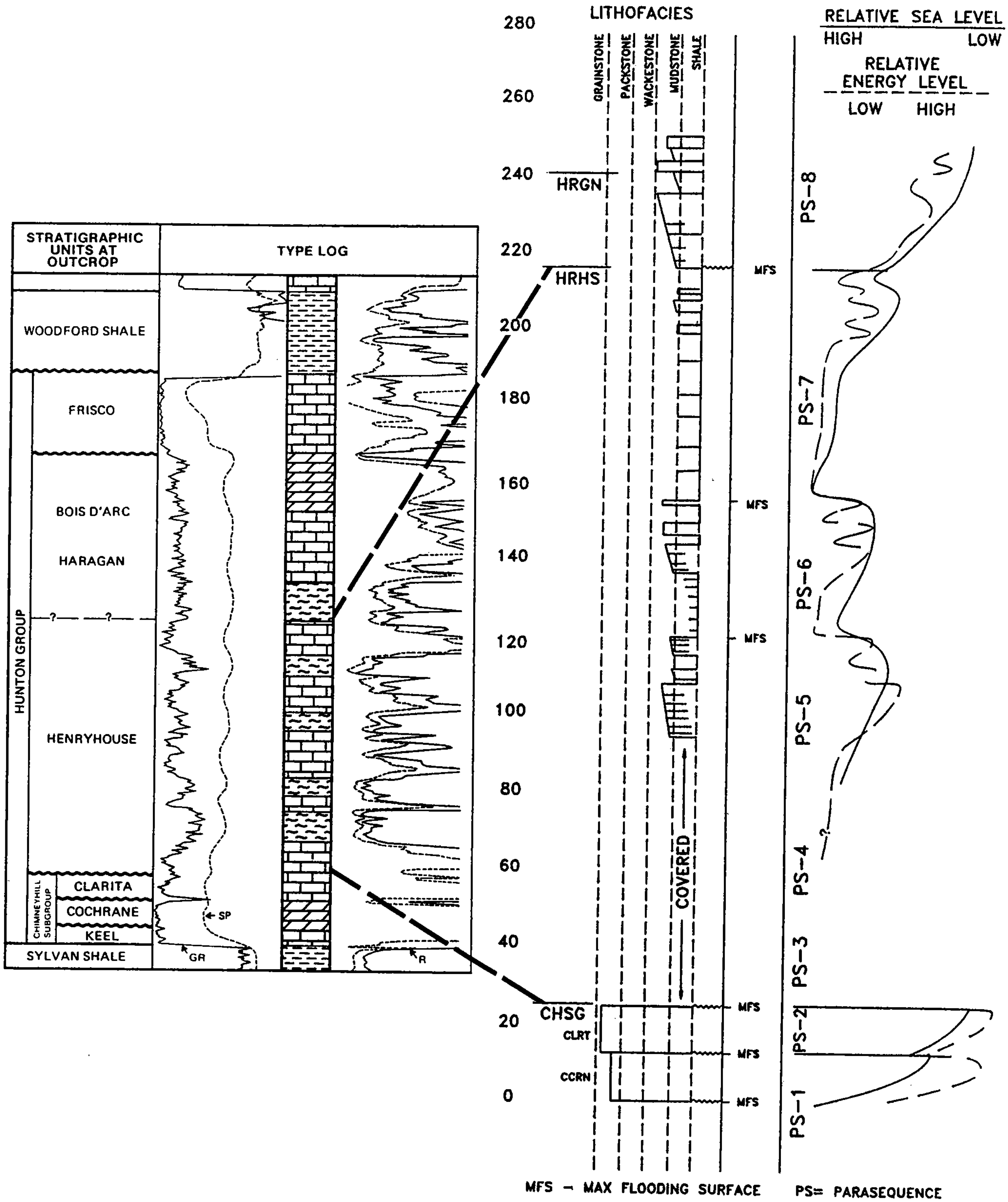


Figure 9. Comparison of type log (left) for southern Oklahoma to "Pseudo E-log" (right) derived from the outcrop of Henryhouse and Haragan/Bois d'Arc strata at Stop 3 ("Pseudo E-log" developed by field study supplemented by descriptions of Fay, 1989).

STOP 4

WOODFORD SHALE

SE¼ sec. 25, T. 2 S., R. 1 E.
Carter County, Oklahoma
(Marker No. 2 — 9 ft below top of Woodford Shale)

Source Rock for the Midcontinent

The Woodford Shale is latest Devonian to earliest Mississippian in age. The boundary has been determined by conodont analysis and is indicated by orange markers on the outcrop (Fig. 10).

At this locality, the Woodford is ~290 ft thick (Fay, 1989), and it consists of dark-gray to black, thin-bedded, cherty, phosphatic shale. The upper 102 ft of this highly organic shale is exposed in this outcrop, whereas the middle 137 ft is covered (Fay, 1989).

The Woodford is cyclic and may be divided into three different units. Within the upper part of the Woodford, the phosphate zone is ~10 ft below the Woodford–Sycamore contact. This zone is ~12 ft thick and is underlain by a nonphosphatic siliceous and fissile shale. This interval is characterized by alternating beds of siliceous and fissile material that are generally 1–4 in. thick. Toward the base of this exposure, these interbeds are thinner (1–2 in. thick) and eventually grade into a fissile shale. Sulfur scale is common on this exposure in the interbedded interval.

The Woodford is a thick, black, highly organic shale, and is one of the primary hydrocarbon source rocks in the U.S. It has been estimated by Comer and Hinch (1987) that 70% of the oil discovered in southern and central Oklahoma was derived from the Woodford Shale. In addition to being an important source rock, the Woodford (though it is commonly cherty and highly fractured) makes a good seal for Hunton petroleum reservoirs. The Woodford is also cyclic, as is evidenced by at least three shallowing upward sequences.

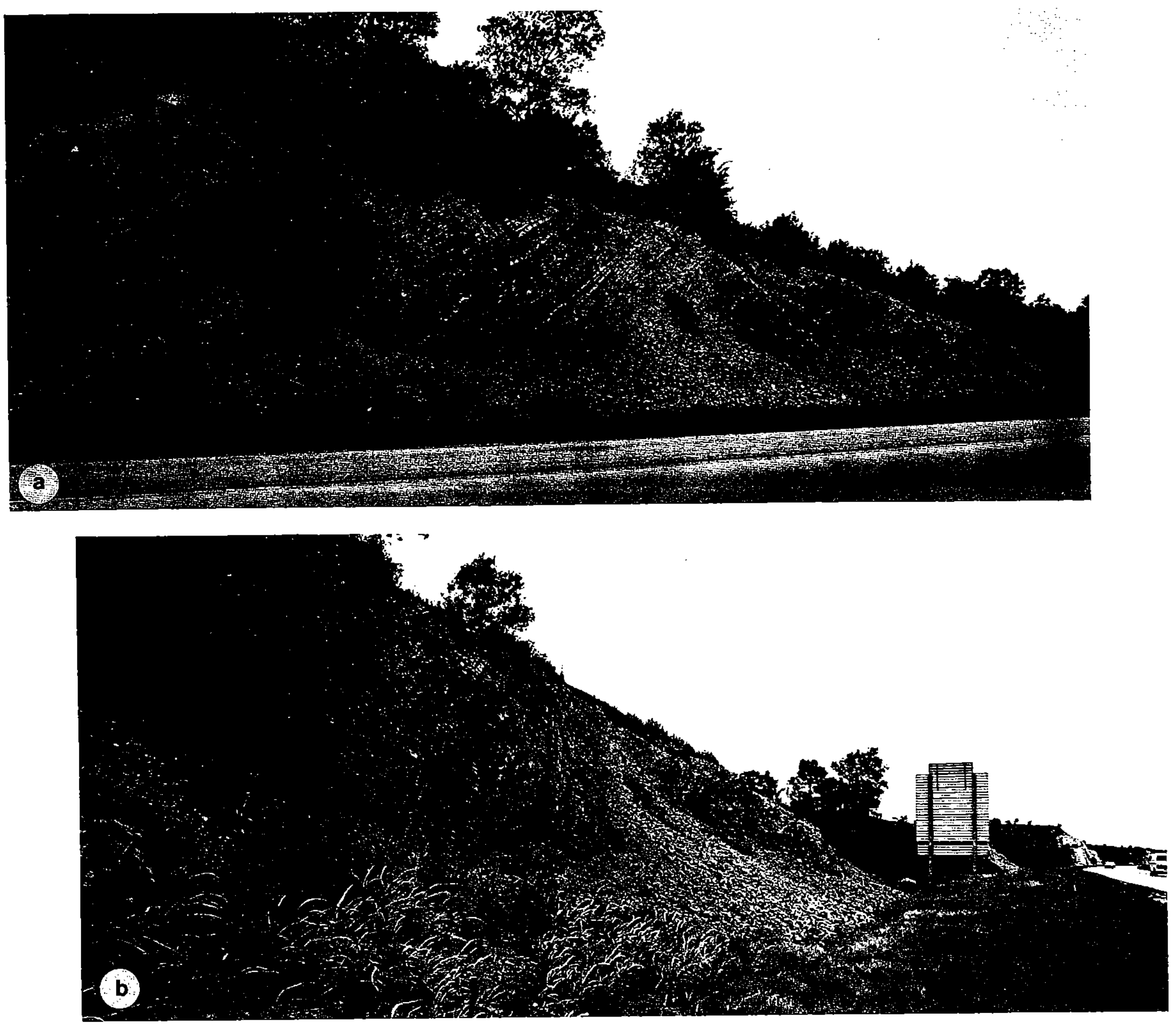


Figure 10. Woodford Shale outcrop on the south flank of the Arbuckle anticline looking toward (a) the west and (b) the north on I-35 (Stop 4; outcrop in photograph is ~300 ft long).

STOP 5

TURNER FALLS OVERLOOK (OPTIONAL)

E½ sec. 36, T. 1 S., R. 1 E.
Murray County, Oklahoma

Travertine Deposition

All streams flowing across limestone beds of the Arbuckle Mountains dissolve calcium carbonate, carrying much of it away in solution. At Turner Falls, on Honey Creek, an edifice of calcium carbonate has been deposited from the stream waters. The deposit consists of spongy-textured travertine. The thickness of the travertine gradually increased with the passage of water over it, until the falls reached a maximum height of 150 ft. Then, probably during middle Pleistocene time, increased rainfall caused Honey Creek to cut a chasm into the depositional platform it had just built, and the waterfall was reduced to half its former height. This beautiful waterfall is now in a stage where it is maintaining a steady state, receiving about as much calcium carbonate in the form of stream-floor deposits as is mechanically abraded during floods. Blue-green algae assist in precipitating the calcium carbonate.

Unlike a typical waterfall, which results from exhumation of a preexisting resistant rock layer by stream erosion, Turner Falls is a subsequent creation of Honey Creek. The creek itself provided the materials to construct the precipice over which it falls.

Behind the falls, as a scenic background and topographically the highest point in the area, are the East Timbered Hills, composed of Precambrian Carlton Rhyolite. Between the rhyolite and the overlook are complexly folded and faulted Late Cambrian and Early Ordovician rocks on the north limb of the Arbuckle anticline. A fault trace parallels the valley of Honey Creek, separating limestones of the Cool Creek Formation at the overlook from those of the McKenzie Hill Formation. Above the cliff and barely behind it is a small lateral valley, occupied by a few deciduous trees, which marks the trace of the Washita Valley fault. Beyond the fault is the Pennsylvanian Collings Ranch Conglomerate (Ham, 1969).

STOP 6

HUNTON GROUP ON U.S. HIGHWAY 77

NW¼ sec. 30, T. 1 S., R. 2 E.
Murray County, Oklahoma

Chimneyhill Subgroup, Henryhouse and Haragan Formations

This outcrop is located on the west side of U.S. Highway 77, just south of the I-35 interchange, and it presents a nearly complete exposure of strata from the Sylvan Shale through Silurian and Early Devonian strata, up to the lower part of the Woodford Shale (Fig. 11). This section is located at the north end of the Arbuckle anticline, with the oldest units located at the south end of the outcrop. This stop was described in detail by Amsden (in Ham, 1973), and the current description includes a portion of that material.

The upper Sylvan Shale, Chimneyhill Subgroup, and the Henryhouse and lower Haragan Formations are completely exposed; however, the upper Haragan and the Woodford are rather poorly exposed. Approximately 200 ft of Hunton strata are present, with only the Bois d'Arc and Frisco Formations (of Early Devonian age) absent. The Frisco, which is a well-defined organo-detrital limestone of middle Early Devonian age (Deerparkain–Seigenian), has been removed by post-Hunton/pre-Woodford erosion and is absent in this part of the Arbuckle Mountains. The Bois d'Arc Formation is absent due to local merging with the shaly mudstone to wackestone facies of the Haragan Formation.

Stratigraphy and Lithofacies

The Chimneyhill Subgroup includes, from oldest to youngest, the Keel, Cochrane, and Clarita Formations; it consists of organo-detrital limestones that can be distinguished from the underlying Sylvan Shale and overlying Henryhouse Formation because these units are relatively low in insoluble clay and silt-size detritus (Amsden, 1967). They constitute an incomplete sequence ranging in age from the Late Ordovician (late Ashigillian) to early Late Silurian (Wenlockian), with the individual formations separated by periods of erosion and, at least, local truncation. Nevertheless, the Chimneyhill Subgroup has proved to be a useful stratigraphic division in the subsurface, where data are often inadequate to distinguish the individual formations.

The oldest Hunton formation is the Ordovician Keel limestone, which is a distinctive, fossiliferous oolite that is cemented partially by spar and partially by micrite. The unit is 4 ft thick and has well-defined upper and lower contacts. The Keel is separated from the overlying Cochrane by a time interval spanning a considerable part of the Early Silurian (early and middle Llandoverian), during which time there was a least local erosion and truncation of the Keel (Amsden, 1960,1963). The contact between the Keel and Cochrane is very irregular at this outcrop, and the unconformity is readily observed.

The overlying Cochrane Formation is ~6 ft thick and comprises organo-detrital limestones with substantial glauconite. Brachiopods from this formation indicate a late Early Silurian (late Llandoverian) age, and point to a correlation with the Blackgum Formation of eastern Oklahoma and the Sexton Creek Formation of southeastern Illinois.

The Clarita Formation is ~14 ft thick and is completely exposed. This formation is divided into a lower Prices Falls Member, which is a thin but persistent shale or marl bed that can be recognized over most of the outcrop area, and an upper Fitzhugh Member, which consists primarily of fossiliferous mudstones and wackestones in the lower 10 ft, whereas the upper few feet is quite shaly and resembles the overlying Henryhouse Formation. Nonetheless, the Clarita–Henryhouse contact is lithologically well defined at this exposure, with the base of the Henryhouse being marked by a sharp increase in insoluble detritus (the basal Henryhouse beds average ~47% HCl insolubles; Amsden, 1978). There is also a change in the concentration of microfossils; the uppermost Clarita strata carry a rich fauna of arenaceous Foraminifera, whereas in the basal Henryhouse beds these fossils are much less numerous. Throughout most of the Arbuckle Mountains, the Clarita–Henryhouse boundary is a reasonably well-defined, mappable contact.

The Henryhouse Formation is ~100 ft thick and is mostly marlstone with a few thin calcareous shale beds in the lower part. The marlstone is composed typically of finely divided carbonate mixed with clay and silt-size insoluble detritus. The carbonate is mostly calcite with only scattered dolomite crystals. Fossils are scattered throughout the matrix in varying concentrations, but the matrix probably has a mud-supported fabric

The Henryhouse bears a large, well-preserved invertebrate fauna. Amsden (1951), Strimple (1963), Lundin (1965), Sutherland (1965), and Campbell (1967) assigned the Henryhouse to the Upper Silurian, although its exact position within this series generally has not been specified. The present outcrop is of special interest because it is one of the three localities from which Decker (1935) reported graptolites, and Decker assigned them an early Ludlovian age. The graptolite beds present a distinctive Henryhouse facies that is unusually high in detrital quartz and mica. The lower 18 ft of the Henryhouse is high in terrigenous detritus and includes four thin, graptolite-bearing shales.

The Henryhouse–Haragan contact here, as elsewhere in the Arbuckle Mountains, is poorly defined lithologically. However, this contact is biostratigraphically well defined and can be mapped on this basis throughout the outcrop area. On average, the Haragan has a slightly lower insoluble-detritus content; however, the difference is too small to serve as a useful distinction at the outcrop, in hand specimen, or in thin section. Despite this obscure lithostratigraphic break, the Haragan–Bois d'Arc fauna (Helderbergian) is separated from the underlying Silurian strata by an unconformity of some magnitude. In fact, over a large area in the southeastern part of the Arbuckle Mountains, Haragan–Bois d'Arc strata bearing numerous Early

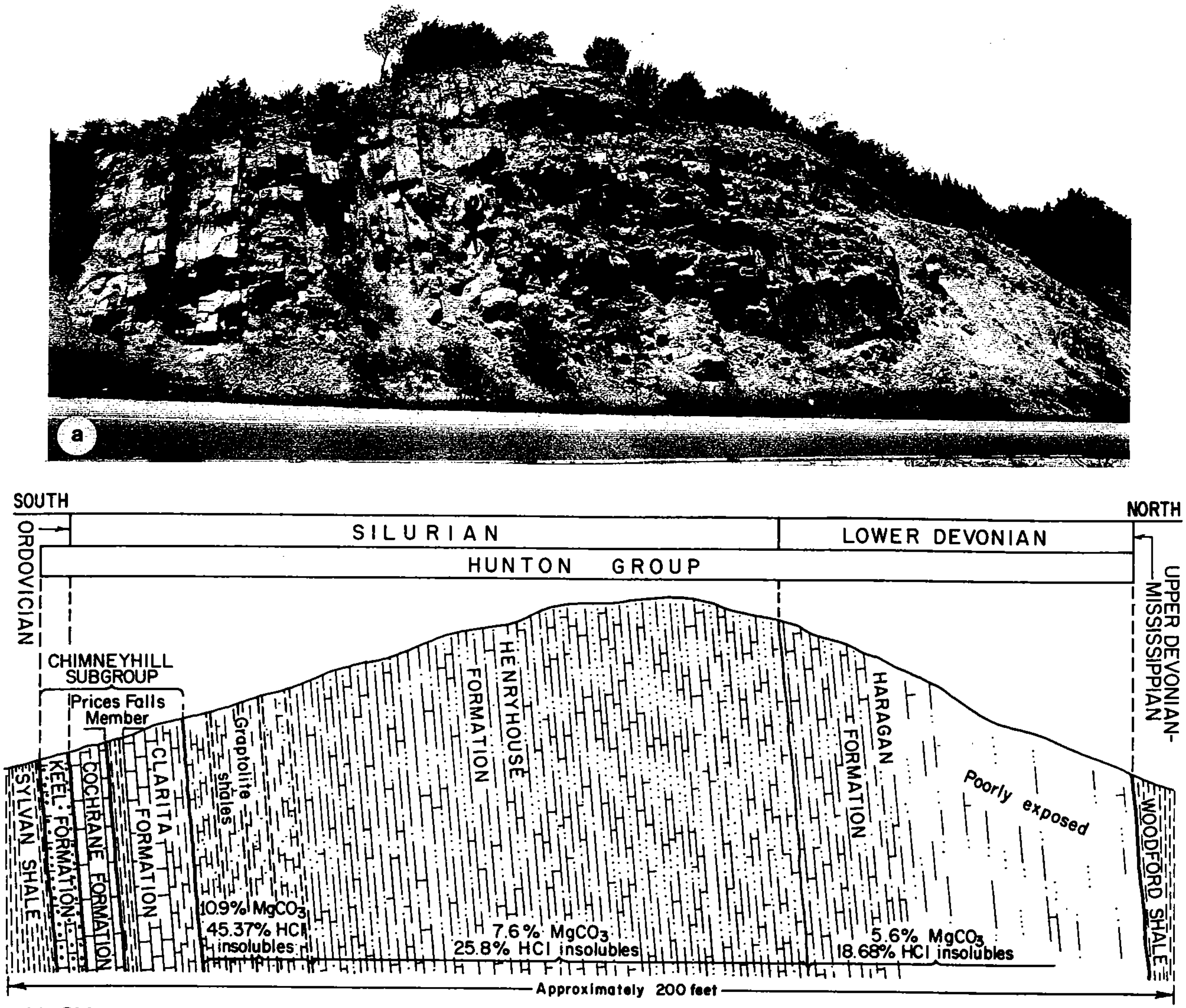


Figure 11. Chimneyhill Subgroup–Henryhouse–Haragan section on the north flank of the Hunton anticline; (a) looking west from Highway 77 (Stop 6; outcrop width is ~200 ft) and (b) measured section (from Amsden, 1978).

Devonian fossils rest on the Cochrane Formation of Early Silurian age, or locally on the Ordovician.

The Haragan Formation is ~60 ft thick. However, the upper part is poorly exposed and there is some evidence of structural disturbance in the lower part of the Woodford Shale. The exposed beds have a characteristic marlstone texture, similar to that described previously for the Henryhouse. The Henryhouse–Haragan contact is poorly defined lithologically (see previous discussion) and must be determined by means of fossils.

The Bois d'Arc Formation is not represented at this locality, although it is fairly well developed in nearby outcrops. The Haragan and Bois d'Arc represent facies of one another, and there is a complete gradation from the typical Haragan marlstone into the low-detrital packstones/grainstones (commonly with chert) of the Bois d'Arc. A minor faunal difference exists between the two lithofacies, but this is thought to be because of environmental changes rather than a time difference (Amsden, 1958a).

Biostratigraphy

The Hunton Group exposed here constitutes one of the most complete sections of uppermost Ordovician through Lower Devonian strata in the south-central United States. The oolitic Keel Formation, at the base of the Hunton, contains the latest Ordovician *Noixodontus* conodont fauna, which occurs in association with Hirnantian (late Ashgillian) brachiopods at other Keel sections and in the Mississippi Valley (Amsden and Barrick, 1986). The overlying Cochrane Formation is the least-well dated of the Hunton units. At this section, and many others in the central Arbuckle Mountains, the conodont faunas are sparse and nondiagnostic; but on the Lawrence uplift, and in the subsurface of Oklahoma, the Cochrane has yielded a *celloni* Zone fauna, making at least the upper part of the unit late Llandoverian (Early Silurian) in age (Barrick and others, 1990). In the subsurface, Amsden (1988) has obtained slightly older, but still late Llandoverian, brachiopods. From this limited information, it appears that a major hia-

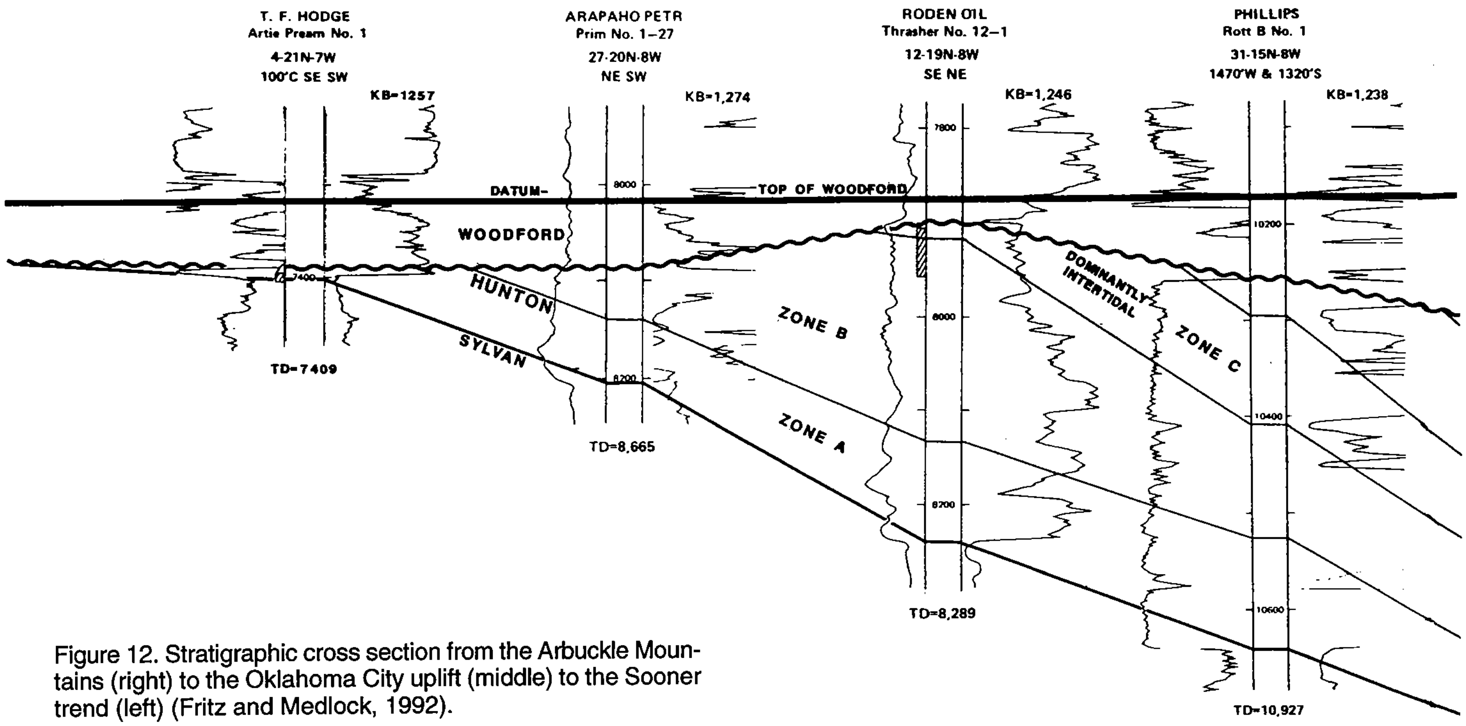


Figure 12. Stratigraphic cross section from the Arbuckle Mountains (right) to the Oklahoma City uplift (middle) to the Sooner trend (left) (Fritz and Medlock, 1992).

tus, encompassing early and middle Llandoverian time, separates the Cochrane from the Keel.

The Clarita Formation, including the Prices Falls and Fitzhugh Members, represents a condensed, but complete record of the Wenlockian (early Late Silurian), based on the succession of conodont faunas described by Barrick and Klapper (1976). The Prices Falls Member bears a conodont fauna, which Dr. Gilbert Klapper of the University of Iowa assigns to the late Llandoverian (Amsden, 1967, p. 944). The Fitzhugh Member carries an early Late Silurian (Wenlockian) brachiopod fauna similar to the one in the St. Clair Limestone of Arkansas (Amsden, 1968). Only a minor hiatus, involving at the most a portion of the *amorphognathoides* Zone, separates the Clarita from the Cochrane.

At this outcrop, the contact of the Henryhouse with the Clarita is a knife-sharp lithologic change, within a single bed, from slightly argillaceous to more-argillaceous limestone. The basal Ludlovian (middle Late Silurian) *crassa* Zone conodont fauna spans this contact and no hiatus can be demonstrated at this outcrop. In sections on the Lawrence uplift and in the Oklahoma subsurface, however, the upper part of the Clarita has been truncated by pre-Henryhouse erosion.

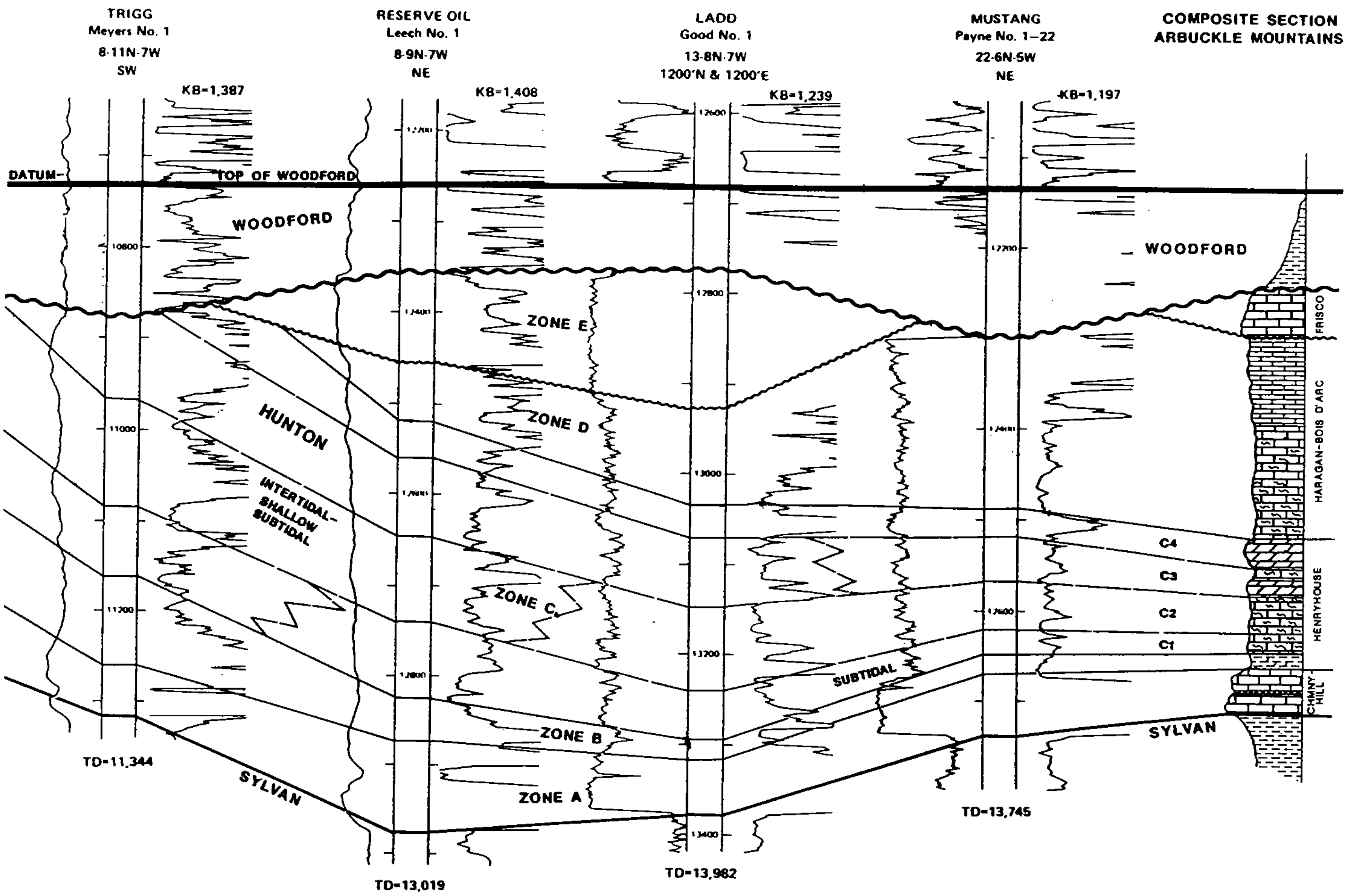
The marlstones of the Henryhouse Formation at this exposure provide an apparently uninterrupted representation of Ludlovian through Pridolian time. The basal 23 ft of the Henryhouse contains unusually large quantities of terrigenous detritus in the central Arbuckles and is rarely exposed. A complete sequence of Ludlovian conodont faunas occurs in the basal unit, spanning the *crassa, ploeckensis,* and *siluricus* Zones. Ludlovian graptolites occur in at least four levels. The upper Ludlovian *snajdri* Zone fauna appears in the more-resistant, less-argillaceous carbonates at 24.5 ft. In sections on the Lawrence uplift, the basal, extremely argillaceous and silty unit is absent, and carbonates bearing the *snajdri* fauna rest directly on the truncated top of the Clarita. A *siluricus* Zone fauna was recovered from the lower Henryhouse in Kingfisher County (Barrick and others, 1990), but little else is known about the age of the base of the Henryhouse in the subsurface.

Approximately 56 ft above the base of the Henryhouse, conodonts of the *eosteinhornensis* Zone, which spans the Pridolian (late Late Silurian), appear. The upper part of the Henryhouse bears this characteristic conodont fauna throughout the outcrop area (Barrick and Klapper, 1992). The Henryhouse–Haragan contact lies 89 ft above the base of the Henryhouse, in the poorly exposed alternating shale and limestone section near the top of the exposure. The contact can be approximated by locating the beds bearing the distinctive crinoid *Scyphocrinites* near the base of the Haragan. Although this section has been less-well sampled, the sequence of conodonts and shelly faunas (Barrick and Klapper, 1992) appears to be identical with that described for the contact in the earlier field-trip stop.

The Haragan bears a large, well-preserved invertebrate fauna, dominated by brachiopods (Amsden, 1958a) and containing many ostracodes (Lundin, 1965), trilobites, corals, and bryozoans. It is assigned an Early Devonian (Helderbergian, Gedinnian) age.

Correlation and Sequence Stratigraphy

Both the Henryhouse and Haragan–Bois d'Arc Formations are cyclic ramp carbonates. Each cycle is a shallowing-upward sequence that can be recognized on outcrops, cores, and logs by the gradual thickening of beds at the top of each cycle and by the increase in shaliness toward the bottom of each cycle. Each cycle or parasequence is



bounded by a disconformity, or in some cases by an unconformity. The Silurian–Devonian contact is represented by a regional unconformity that disrupts the cycles in the Henryhouse and Haragan Formations.

A cross section from the Arbuckle Mountains to the Oklahoma City uplift shows the relationships of the individual parasequences within the Hunton (Fig. 12). Note that each parasequence thickens northward up the ramp.

Reservoir Development

Northward, into the subsurface of the Anadarko basin, the Henryhouse Formation is thicker and is represented by shallow subtidal to lower intertidal facies. These facies develop excellent porosity and represent some of the best reservoirs in Oklahoma. The Bois d'Arc also forms excellent reservoirs in the area between the eastern Anadarko basin and southern Oklahoma.

STOP 7

HUNTON ANTICLINE QUARRY

NW¼SE¼ sec. 31, T. 1 S., R. 3 E.
Murray County, Oklahoma

Structure and Lithology of the Bois d'Arc Formation

This small quarry was opened during the construction of the Lake Arbuckle dam. It contains exposures of the Woodford Shale and the Haragan–Bois d'Arc Formations. A similar section, described and collected by T. W. Amsden, is located ~2.5 mi north of Dougherty, and ~1 mi northwest of Stop 7 (Amsden, 1958a,b).

Stratigraphy and Lithofacies

The most striking feature of this outcrop is the faulted anticline (Fig. 13). This is an incomplete section which begins (at the base) in the Haragan; the lower part of the Haragan is covered, and below this cover is the Sylvan Shale. No Chimneyhill strata were observed in this area, and they may be absent; however, there is considerable faulting, and it is possible that the Haragan is faulted against the Sylvan.

The Haragan is a yellow to gray, fossiliferous, silty mudstone that is distinctly different from the overlying Cravatt Member (basal member of the Bois d'Arc Formation). The Cravatt Member is yellow to bluish-gray, fossiliferous mudstone to wackestone, with prominent nodules of vitreous to tripolitic chert.

Overlying the Bois d'Arc at this locality is the Woodford Shale. The Woodford is best exposed on the northern flank of this fold.

Figure 13. Haragan/Bois d'Arc outcrop at the Hunton anticline quarry (Stop 7; outcrop width is ~600 ft).

STOP 8

FRISCO–BOIS D'ARC CONTACT ON BOIS D'ARC CREEK

NE¼ sec. 11, T. 2 N., R. 6 E.
Pontotoc County, Oklahoma
(Exposures in Bois d'Arc Creek at State Highway 99 bridge)

Facies and Depositional Environment of Frisco Mud Mounds

A nearly complete section from the upper part of the Henryhouse Formation, up through the Haragan and Bois d'Arc Formations, and into the Frisco Formation, is exposed along Bois d'Arc Creek; however, only the Frisco and Bois d'Arc Formations will be examined at this stop.

The exposure of the Frisco Formation found at Bois d'Arc Creek is especially important; it is the best outcrop example of the contact between the Frisco and the Bois d'Arc, and it shows the character of the Frisco mound facies. In fact, the Frisco mound facies has essentially been exhumed at this locality by removal of the overlying Woodford Shale (Fig. 14).

Stratigraphy and Lithofacies

The Frisco consists of a series of biohermal mounds that formed on the eroded surface of the pre-Frisco strata. At this exposure, distinctive facies serve to identify mound and intermound rocks. The core of the mound is typically represented by massive, muddy lime beds, with randomly occurring, rather well-preserved crinoid and bryozoan fragments, along with brachiopods, trilobites, and corals (Fig. 15). This muddy facies apparently developed well

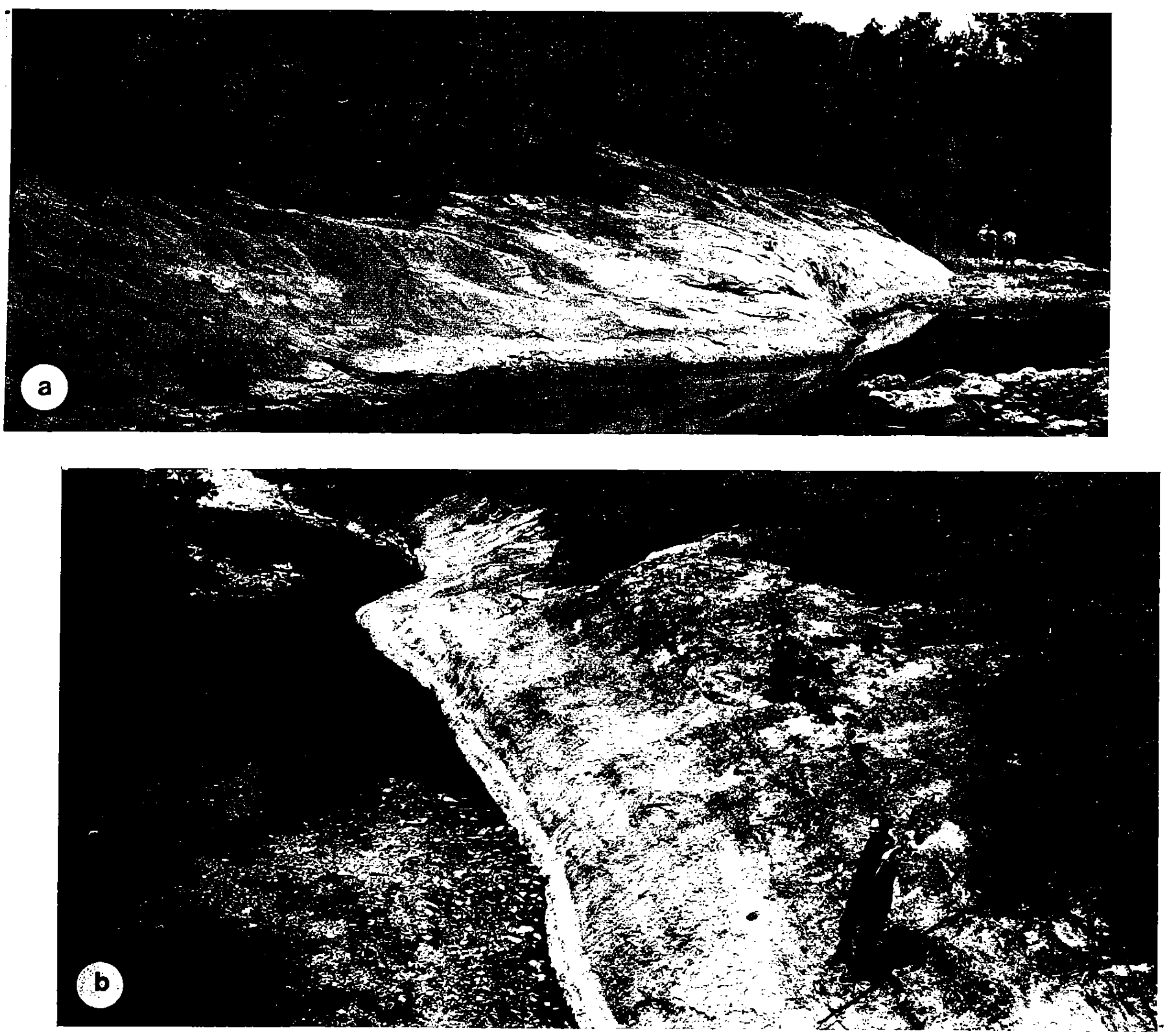


Figure 14. Frisco Formation outcrop along Bois d'Arc Creek (Stop 8) showing the morphology of the mud mound from (a) a side view (looking east) and (b) a side view.

under wave base, where the crinoids and bryozoans act as a "diffuse baffle" that allowed settling of lime muds. As the mound was built up to wave base, the mud was winnowed and a grain-supported facies developed. As the mound aggraded into stronger currents and wave action, dunes and sand waves developed on the crest or the relatively gentle flanks. Finally, as the mound approached sea level, growth stopped until rejuvenation by relative sea-level rise produced by subsidence and/or eustatic changes in sea level. The intermound facies is light-colored, moderately sorted, bioclastic packstone/grainstone with abundant crinoid fragments. Currents sorted the grains and winnowed out the mud in this facies. Intermound beds often onlap the mound facies (Fig. 16a). A "capping facies" that is lithologically very similar to intermound rocks is observed above the mound facies.

These facies can be readily observed along Bois d'Arc Creek. The grainstone facies, for example, can be examined in outcrops on the west side of the highway, immediately across from the parking area. The core mudstone and flanking wackestone/packstone facies can be examined farther west along the creek bed (Fig. 16b). An examination of the Frisco facies to the west along the creek reveals that the Frisco is not one large mound but is a complex of coalescing mounds that are typically 10–20 ft thick. Farther west along the creek, the contact between the Frisco and Bois d'Arc can be recognized by the change in color and weathering profile; i.e., the Frisco is dark, massive, and rounded in appearance, whereas the Bois d'Arc is medium to light gray and bedded.

Large, well-preserved crinoid fragments are present along the unconformable contact between the Frisco and Bois d'Arc Formations. Although close examination reveals some weathering profiles and possible karst, the contact appears to be, for the most part, unaltered. Regionally, the unconformity is angular in nature, and there is some suggestion that the paleotopography of the Bois d'Arc surface may have been a controlling factor in the placement and development of the Frisco mud mounds; i.e., after erosion and marine transgression of the Bois d'Arc, the subsea paleohighs may have been points of initial development of the mound facies.

Biostratigraphy

At this locality the Henryhouse is truncated by the Haragan Formation, as described by Amsden (1988, and earlier papers), and a couple of meters of the top of Henryhouse, present at nearby sections, may be missing here, based on study of the conodont biofacies (Barrick and Klapper, 1992). The upper Henryhouse contains the Late Silurian, Pridolian, *eosteinhornensis* conodont fauna.

The extremely thin Haragan (4 ft thick) is overlain by the cherty Cravett Member of the Bois d'Arc Formation. Although Early Devonian conodonts occur in both units, the first zonal-diagnostic species does not appear until nearly 13 ft above the base of the Cravett. Here, *Icriodus postwoschmidti* occurs in thin, skeletal packstone layers at the transition from the Cravett into the skeletal grainstones of the Fittstown Member of the Bois d'Arc, and ranges through at least the lower 13 ft of the Fittstown. *Icriodus postwoschmidti* occurs near the base of the Haragan at the sections in the central Arbuckles; thus, the Cravett Member

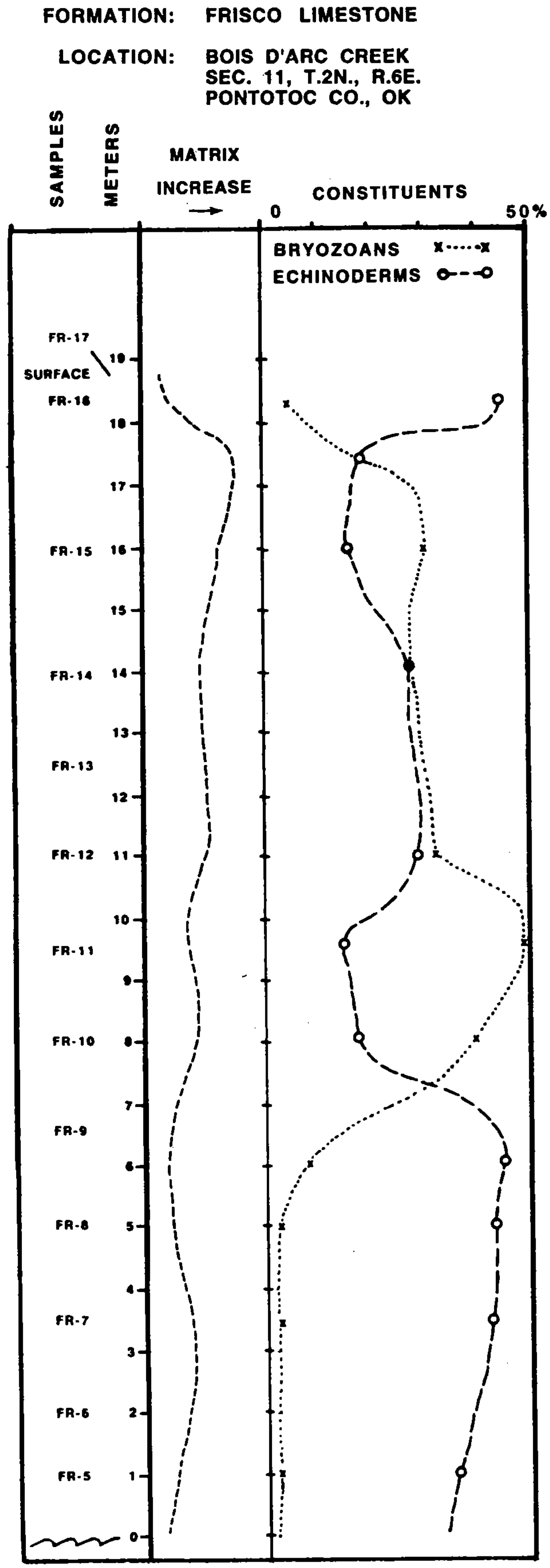


Figure 15. Measured section showing composition and constituents in a Frisco mud-mound sequence (after Harrison, 1987).

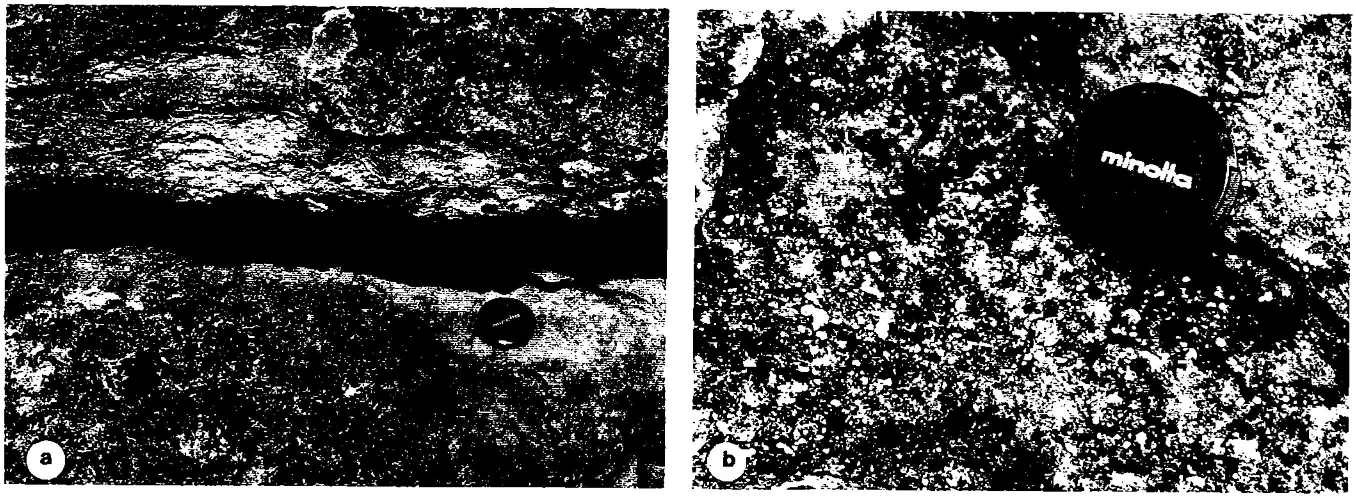


Figure 16. Frisco limestone facies: (a) contact between mound core and overlying capping facies, and (b) grainy texture of flanking facies.

of the Bois d'Arc and the lower part of Fittstown Member can be assigned to the same conodont zone as the Haragan, the early Lochkovian *eurekaensis* Zone (Barrick and Klapper, 1992).

In the coarse skeletal grainstones in the upper part of the Fittstown Member along Bois d'Arc Creek, representatives of the conodont genus *Ancyrodelloides* appear (Barrick and Klapper, 1992). This taxon is restricted to the *delta* Zone, the Lochkovian zone that immediately overlies the *eurekaensis* Zone. This shows that the upper part of the Fittstown Member of the Bois dArc is slightly younger than the Haragan Formation.

The overlying Frisco Formation yields abundant conodonts, but the high-energy environment in which the skeletal grainstones accumulated have broken most specimens beyond confident identification. Klapper (in Amsden, 1985, p. 2) indicated that *Icriodus claudiae* may be present in the Frisco. This species occurs in Pragian (Early Devonian) strata elsewhere, and a minor hiatus involving at least a single conodont zone appears to be present at the base of the Frisco (Barrick and others, 1990).

Reservoir Development

The Frisco is unusual in that its petroleum reservoirs are in limestone, whereas most of the Hunton reservoirs are in dolomite. Apparently the depositional setting and/or the geochemistry were not conducive to dolomitization during Frisco time. The Frisco does have excellent moldic, intra-, and inter-particle porosity; it can be a prolific producing horizon, as is evidenced by production at the West Edmond field.

STOP 9

ROAD CUT NEAR THE SOUTH FORK OF JACKFORK CREEK

NE¼NE¼SE¼ sec. 32, T. 3 N., R. 6 E.
Pontotoc County, Oklahoma

Henryhouse Facies

This outcrop consists of ~8 ft of the Henryhouse Formation. The Henryhouse is primarily light-gray, fossiliferous, silty carbonate. This exposure is located near the top of the Henryhouse, close to the boundary with the overlying Haragan–Bois d'Arc (Fig. 17). The Haragan is thin in this area (6 ft thick; Amsden, 1960) and is lithologically similar to the Henryhouse. The faunal differences between the Henryhouse and Haragan are very distinct, and biostratigraphy is used to locate the boundary with precision (Amsden, 1957). The resistant beds at the top of the hill are Bois d'Arc limestone.

Figure 17. Henryhouse Formation outcrop along the road cut near the South Fork of Jackfork Creek (Stop 9; outcrop width is ~40 ft).

STOP 10

HUNTON GROUP IN THE LAWRENCE QUARRY SOUTH OF ADA

sec. 36, T. 3 N., R. 5 E.
Pontotoc County, Oklahoma

Chimneyhill Subgroup and Sylvan Shale

This stop contains excellent exposures of the contact between the Chimneyhill Subgroup and the underlying Sylvan Shale (Fig. 18). These rocks are exposed along the east wall of the quarry where they dip gently (~6°) to the east–southeast.

Stratigraphy and Lithofacies

The Chimneyhill strata represented at this stop are the Keel Formation (including the Ideal Quarry Member of Amsden, 1957) and the Cochrane Formation. In outcrop, the more-resistant Cochrane and Keel cap the wall of the excavation. The Keel appears lighter colored and more massive than the overlying darker and rubbly-appearing Cochrane. Petrographic analysis reveals the Ideal Quarry, Keel, and Cochrane units have distinct lithologies that reflect their differing depositional environments.

The Ideal Quarry Member is closely related to the Keel and has generally been considered a distinct lithology in the basal part of the Keel (Reeds, 1911; Maxwell, 1936; Amsden, 1957,1960). The Ideal Quarry is typically a 1–5-ft-thick, fossil-rich (well over 50% fossils; Amsden, 1960) limestone. The fossil material consists mostly of brachiopods, gastropods, and pelmatozoans. The matrix commonly is clay- to silt-size carbonate, and less commonly is sparry calcite. Ooids are rare at the base and increase in number toward the boundary with the overlying Keel. A significant feature of the Ideal Quarry Member is the coating of the fossils and grains by concentric layers of calcareous material. This coating has produced irregular to spherical grains that resemble oolites. These grains are apparently oncolites formed by algae.

The carbonate mud, fossil content, and algal coatings suggest the Ideal Quarry Member was deposited in an upper subtidal setting. The abundance of fossils suggests the environment was quiescent enough to support a rich fauna, yet shallow enough (within the photic zone) to support significant algal growth.

The type locality for the Keel Formation is also at the site of the Lawrence quarry (Amsden, 1960). Here it is approximately 5–7 ft thick and is a light-colored, oolitic limestone. The contact between the Ideal Quarry Member and the upper part of the Keel Formation is gradational and often difficult to determine in outcrop. However, the Keel is oolite rich and relatively fossil poor, whereas the Ideal Quarry is an oncolitic, bioclastic limestone. The symmetry of the Keel oolites suggests they were deposited in a wave-agitated shoaling environment, where mechanical and chemical action induced sphericity. Most ooids are 1 mm in diameter, but some pisolites exceed 3 mm. The most common matrix is sparry calcite, but microcrystalline calcite is also a significant cement. Cross-bedded and graded ooids are evident in the Keel, suggesting wave or current action. Silicification is common in the Keel, especially in the upper part. In some places the silicification is more intense, and this results in nodules and lenses of oolitic chert.

The cross-bedded, oolitic nature of the Keel suggests it was deposited as a carbonate shoal in an upper subtidal to lower intertidal setting. The high energy of this environment encouraged oolite precipitation, but was not conducive to faunal proliferation.

The type locality for the Cochrane Formation is nearby, in sec. 5, T. 2 N., R. 6 E., Pontotoc County (Amsden, 1960). The Cochrane is ~18 ft thick at the type locality, but is somewhat thinner in the quarry wall where it has been truncated by erosion.

Figure 18. Chimneyhill Subgroup outcrop in the Lawrence Quarry (Stop 10) showing (a) Keel and Cochrane Formations overlying the Sylvan Shale, and (b) close-up of contact between the Keel and Cochrane (hammer 1 ft below contact).

The Cochrane is distinctly different from the underlying Keel. The Cochrane is a richly fossiliferous, glauconitic limestone. Amsden (1957) reported that it is composed of 50–60% fossils. The matrix is composed primarily of microcrystalline and sparry calcite cement, with lesser amounts of micrite. Glauconite is almost universally present in the Cochrane as rounded, millimeter-size grains. Chert is common in the quarry exposures and forms a nodular zone at the Cochrane/Keel contact. An unconformity separates the Cochrane and the Keel Formations. Amsden (1960) suggests this unconformity does not represent a large time span, but it does involve localized uplift and sufficient erosion to remove the Keel and Ideal Quarry strata, and permit the Cochrane to rest directly on the Sylvan. The Cochrane is normally overlain by the Clarita Formation, but the Clarita is absent at this exposure.

The glauconitic and bioclastic nature of the Cochrane suggest that it was deposited in a shallow-marine setting where water chemistry favored the precipitation of glauconite.

The brecciated Cochrane and Keel rocks in the quarry wall suggest the Hunton was subaerially exposed and subjected to dissolution processes. The exact timing of the paleokarst genesis is not certain, but it may have formed during the pre-Frisco or pre-Woodford epeirogenies.

SUMMARY OF FIELD TRIP

Although most of the key Hunton outcrops were examined during this trip, only outer-ramp rocks were observed. Therefore, it would not be correct to assume that this is a typical Hunton section, especially inasmuch as little of the reservoir-rock facies was observed. On the contrary, northward into the Anadarko basin the Hunton, especially the Chimneyhill–Henryhouse section, changes dramatically to mid- and inner-ramp deposits that are porous and develop excellent reservoirs.

Nevertheless, data from detailed outcrop evaluations, tempered with realization of their position on the Hunton carbonate model, integrated with core and log data, can provide valuable information on stratigraphy, lithofacies, and biofacies. This information can then be correlated into the overall sequence-stratigraphy model to develop an understanding of Hunton reservoir development and potential exploration opportunities.

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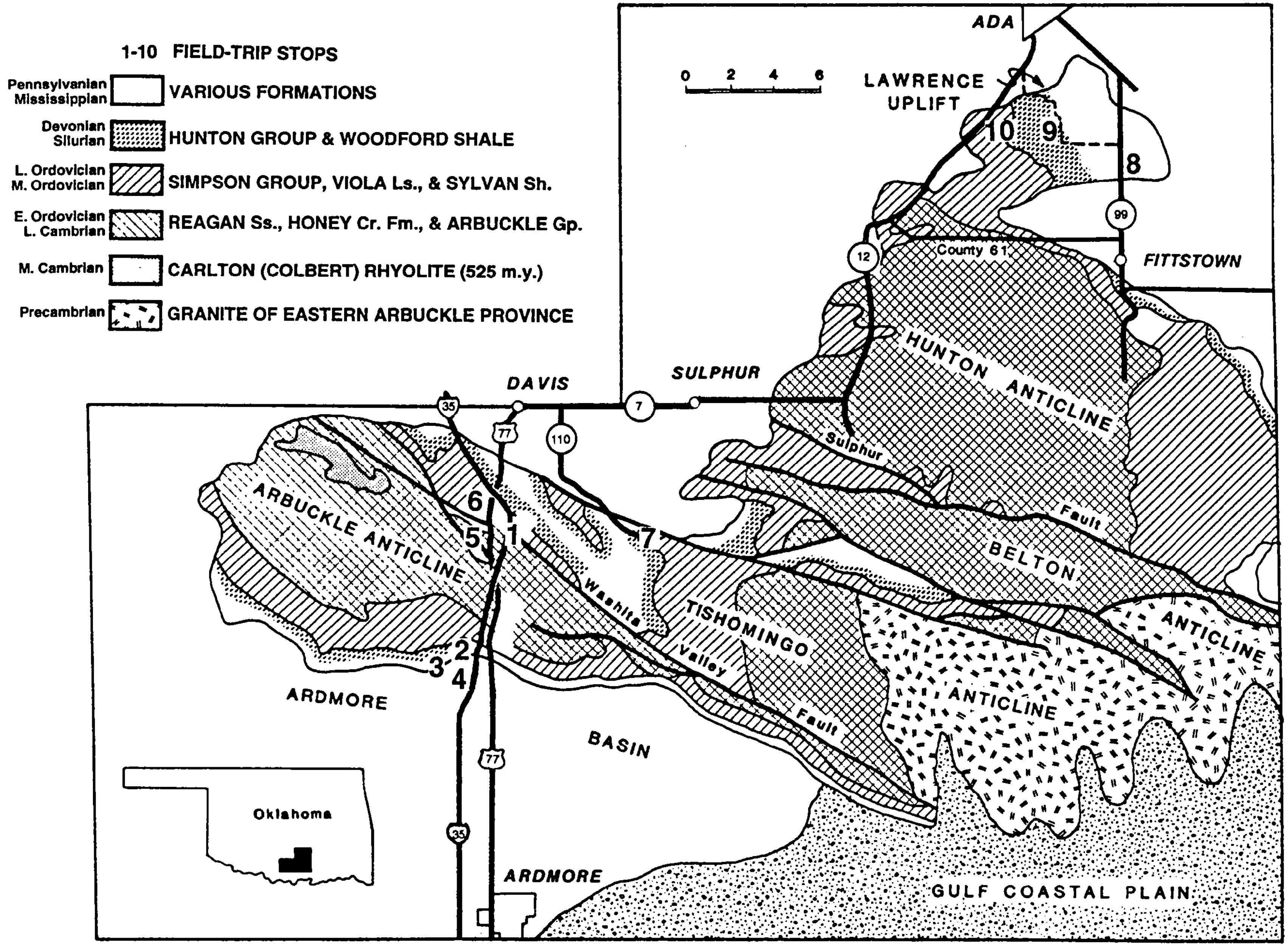
APPENDIX — Road Log Along Interstate 35
(from Fay, 1989)

South miles	North miles	Station no.	Marker no.	Lane* W–C–E–77	
0	7.3	2312+83′		C	Bridge—On Goddard Formation
		2316		W	Goddard Formation–Delaware Creek (Caney) Shale contact
0.15	7.15	2320+75′		W	Delaware Creek (Caney) Shale outcrop on west side (Fig. 1)
		2323+20′		C	**44 Mile Post**
0.2	7.1	2323+25′		W	Delaware Creek (Caney) Shale–Sycamore Limestone contact (Fig. 2)
		2323+80′	(1)	W	**Marker No. 1.**—Sycamore Limestone, 13 ft below top
0.3	7.0	2329+52′		W	Sycamore Limestone–Woodford Shale contact (Figs. 3–5)
		2329+70′	(2)	W	**Marker No. 2.**—Woodford Shale, 9 ft below top
0.4	6.9	2333+83′		W	Woodford Shale–Hunton Group contact (Fig. 6)
		2333+98′	(3)	W	**Marker No. 3.**—Bois d'Arc Limestone, 1 ft below top
0.5	6.8	2337+80′		W	Hunton Group–Sylvan Shale contact
		2337+80′	(4)	W	**Marker No. 4.**—Cochrane Limestone of Chimneyhill Subgroup of Hunton Group, 3 ft above base
0.6	6.7	2342+70′		W	Sylvan Shale–Viola Group contact
		2342+91′	(5)	W	**Marker No. 5.**—Viola Group, 15 ft below top
0.65	6.65	2345+23′		W	Oil seep in Viola Group, east side of west lane (Fig. 7)
0.8	6.5	2352+10′		W	Viola Group–Pooleville Limestone contact
		2353+15′	(6)	W	**Marker No. 6.**—Pooleville Limestone Member of Bromide Formation, Simpson Group, 1 ft below top
0.9	6.4	2359+30′		C	Center of bridge over rural road
1.2	6.1	2370+25′		E	Basal McLish sandstone–Oil Creek limestone contact
		2376		C	**45 Mile Post**
		2378+59′	(7)	E	**Marker No. 7.**—Oil Creek Formation, 49 ft above base (Fig. 8)
1.34	5.96	2379+25′		E	Oil Creek Formation–Joins Formation contact
1.45	5.85	2383+35′		E	Joins Formation–Arbuckle Group West Spring Creek Formation contact
		2383+90′		E	Algal bed in West Spring Creek Formation, beginning of West Spring Creek road cut on east side of east lane
		2384+15′	(8)	E	**Marker No. 8.**—West Spring Creek Formation of Arbuckle Group, 67 ft below top
		2387+37′	(9)	E	**Marker No. 9.**—West Spring Creek Formation, C-D-P Zone, or *Ceratopea–Diparelasma–Pomatotrema* Zone, 284 ft below top
1.5	5.8	2387+48′		E	Sandstone bed below C-D-P Zone—top of red bed sequence in West Spring Creek Formation (Figs. 9–10)
1.65	5.65	2395+06′		E	Graptolite bed, *Didymograptus protobifidus*, in bluish-gray limy shale a few inches thick (marker for dating West Spring Creek Formation as Early Ordovician in United States) (Fig. 11)
1.8	5.5	2401+15′		E	Base of red beds in West Spring Creek Formation
1.85	5.45	2404+40′		E	West Spring Creek Formation–Kindblade Formation contact, tan dolomitic siltstone on gray limestone (Fig. 12)
		2404+42′	(10)	E	**Marker No. 10.**—Kindblade Formation, 2 ft below top
2.2	5.1	2423+15′	(11)		**Marker No. 11.**—Axis of anticline in lower Kindblade Formation, 100 ft above base (Fig. 13)
2.25	5.05	2425+30′		E	Fault contact between Kindblade Formation and Cool Creek Formation
		2428+80′		C	**46 Mile Post**
2.45	4.85	2435+80′		W	Lower Cool Creek Formation outcrop
		2437+05′	(12)	W	**Marker No. 12.**—Cool Creek Formation, about 700 ft below top
2.6	4.7	2444+80′		W	Cool Creek Formation–McKenzie Hill Formation contact

*W = West Lane, generally west side; C = Center of Median between four lanes; E = East Lane, generally east side; 77 = U.S. Highway 77.

South miles	North miles	Station no.	Marker no.	Lane* W–C–E–77	
2.9	4.4	2459+40′		W	McKenzie Hill Formation–Butterly Dolomite contact
3.0	4.3	2466+90′		W	Butterly Dolomite–Signal Mountain Formation contact
3.2	4.1	2474+30′	20	W	**Marker No. 20.**—Signal Mountain Formation–Royer Dolomite, just above contact
		2481+60′		C	**47 Mile Post**
		2483+15′	21	W	**Marker No. 21.**—Royer Dolomite, 150 ft above base
3.45	3.85	2485+80′		W	Royer Dolomite–Fort Sill Limestone contact, on slope south of ditch
3.48	3.82	2488+50′		W	Chapman Ranch thrust fault—covered and cannot be seen—Fort Sill Limestone on Royer Dolomite; approximate position is at fence line
3.7	3.6	2505		C	Bridge, above Highway 77, Cattle Pens Interchange
		2534+40′		C	**48 Mile Post**
4.3	3.0	2534+80′		W	Royer Dolomite–Fort Sill Limestone contact—highly distorted beds
		2535+20′	22	W	**Marker No. 22.**—Fort Sill Limestone, 9 ft below top (Fig. 14)
		2541	23	E	**Marker No. 23.**—Royer Dolomite, 2 ft above base
4.4	2.9	2542+20′		W	Fault—Fort Sill Limestone–Royer Dolomite
4.7	2.6	2560+20′		W	Royer Dolomite starts to dip north
4.9	2.4	2564+80′		W	Fault—Royer Dolomite–Signal Mountain Formation
		2566+80′	24	W	**Marker No. 24.**—Signal Mountain Formation
4.95	2.35	2569+20′		W	Fault—Signal Mountain Formation–Butterly Dolomite—part of Washita Valley fault system
		2570	25	E	**Marker No. 25.**—Butterly Dolomite
5.07	2.23	2575+30′		W	Fault—Butterly Dolomite–Cool Creek Formation
5.15	2.15	2579+50′		W	Cool Creek Formation–Kindblade Formation contact; Cool Creek is a massive oolitic zone
		2581	26	W	**Marker No. 26.**—Lower Kindblade Formation
		2587+20′		C	**49 Mile Post**
5.35	1.95	2590+60′		W	Kindblade Formation, dipping steeply northward (normal)
5.4	1.9	2592+50′		W	Kindblade Formation—Displacement of drill holes shows evidence of recent movements; see fold on east side of west lane
		2596+50′	27	E	**Marker No. 27.**—Kindblade Formation, 25 ft below top, on east turnout, west side of north end
5.5	1.8	2598+55′	28	W	**Marker No. 28.**—Kindblade Formation, 52 ft below top, on west turnout, west side of south end—tectonic breccia in Kindblade (Fig. 20)
5.55	1.75	2599+30′		W	Kindblade Formation–West Spring Creek Formation contact, on west turnout, west side of south end, just north of Marker No. 28 about 75 ft
5.6	1.7	2603+80′		W	Fault—West Spring Creek Formation–Collings Ranch Conglomerate, on west turnout, west side of north part, north of road to geologic sign
		2604+50′	29	W	**Marker No. 29.**—Collings Ranch Conglomerate, 70 ft north of fault, on west turnout, west side of north part
5.7	1.6	2610	30	E	**Marker No. 30.**—Collings Ranch Conglomerate, synclinal axis (back cover, bottom)
		2628+75′	16	W	**Marker No. 16.**—Collings Ranch Conglomerate, 25 ft south of fault
6.1	1.2	2629		W	Fault—Collings Ranch Conglomerate–Bromide Formation; Collings Ranch dips 11° S—Bromide overturned, dipping south about 61°—Collings Ranch flat on top of Bromide (back cover, top)
		2629+22′	19	W	**Marker No. 19.**—Mountain Lake Member of Bromide Formation, fault contact, 126 ft below Viola Group
		2630+14′	17	W	**Marker No. 17.**—Pooleville Limestone Member of Bromide Formation, 36 ft below Viola Group, faulted
6.15	1.15	2630+50′		W	Bromide Formation Pooleville Limestone Member–Viola Group contact—vertical dip with Collings Ranch Conglomerate resting unconformably above synclinal axis in Viola Group
6.2	1.1	2633+50′		W	
		2633+70′	18	W	**Marker No. 18.**—Viola Group, 175 ft above base
6.25	1.05	2637+20′		W	Viola Group–Bromide Formation Pooleville Limestone Member contact

South miles	North miles	Station no.	Marker no.	Lane* W–C–E–77	
		2638+50′	(31)	E	**Marker No. 31.**—Bromide Formation Pooleville Limestone Member, 15 ft below top
6.3	1.0	2639		W	Bromide Formation Pooleville Limestone Member–Mountain Lake Member contact
		2639+80′	(32)	E	**Marker No. 32.**—Mountain Lake Member of Bromide Formation, 6 ft below top
		2640		C	**50 Mile Post**
6.4	0.9	2643		W	Mountain Lake Member of Bromide Formation, lower sandstone beds
		2643+50′	(33)	E	**Marker No. 33.**—Mountain Lake Member of Bromide Formation, lower sandstone, 53 ft above base (Fig. 19)
6.45	0.85	2644+50′		E	Fault zone, in valley, basal McLish sandstone to east, against Bromide and Tulip Creek Formations to west (Fig. 18)
		2649+68′	(34)	E	**Marker No. 34.**—McLish Formation, 2 ft below top
6.5	0.8	2649+70′		E	McLish Formation–Tulip Creek Formation contact (front cover, top)
6.6	0.7	2653		E	Tulip Creek Formation–Bromide Formation contact in valley
6.7	0.6	2657+10′		E	Bromide Formation Mountain Lake Member Pooleville Limestone Member contact
6.75	0.55	2658+60′		E	Bromide Formation Pooleville Limestone Member–Viola Group contact (front cover, bottom)
		2658+62′	(35)	E	**Marker No. 35.**—Viola Group, overturned, 2 ft above base
6.8	0.5	2664+60′		W	Viola Group—fault zone splinters, ending in sharp anticline
6.9	0.4	2668+30′		W	Viola Group–Sylvan Shale contact, covered; go west to U.S. Highway
			(13)	77	77 to see contact, where **Marker No. 13** is in the Viola, 24 ft below top
6.95	0.35	2672		W	Sylvan Shale–Hunton Group contact, covered; go west to U.S. High-
			(14)	77	way 77 to see contact, where **Marker No. 14** is in the Keel Limestone of the Chimneyhill Subgroup of the Hunton Group, 2 ft above the base (Fig. 17)
7.0	0.3	2675		C	Bridge over Highway 77-D; Haragan Limestone on west side of bridge; Woodford Shale on east side of bridge, north side 77-D
7.05	0.25	2677+80′		W	Hunton Group Haragan Limestone–Woodford Shale contact
7.15	0.15	2679+50′		E	Woodford Shale–Sycamore Limestone contact (Fig. 16)
		2681+55′	(36)	E	**Marker No. 36.**—Sycamore Limestone, 71 ft above base; on
			(15)	77	Highway 77 to west is **Marker No. 15**, about 110 ft above the base of the Sycamore Limestone
7.25	0.05	2685		W	Sycamore Limestone–Delaware Creek (Caney) Shale contact
7.3	0.00	2692+80′		C	**51 Mile Post**—West end of bridge across Honey Creek
		2703+33′		C	Bridge—U.S. Highway 77 overpass above U.S. I-35



Generalized geologic map of the Arbuckle Mountains showing field-trip stops 1–10. Modified from work by Ham published in Oklahoma Geological Survey Guidebook 17 (1969).