The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—Texas

By R. S. KIER, L. F. BROWN, JR., *and* E. F. MCBRIDE

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Historical review and summary of
areal, stratigraphic, structural,
and economic geology of Mississippian
and Pennsylvanian rocks in Texas

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By R. S. KIER,[1] L. F. BROWN, JR.,[2] and E. F. MCBRIDE[3]

ABSTRACT

Carboniferous rocks in Texas crop out in the Colorado, Brazos, and Trinity River valleys in central and north-central Texas and in the Trans-Pecos region of west Texas. In central and north-central Texas, Mississippian and Pennsylvanian strata are mainly shale, sandstone, and limestone deposited in fluvial-deltaic and interdeltaic environments, on open shelves and carbonate platforms, and in shelf-edge, slope, and basin environments. In west Texas, Carboniferous rocks are exposed in the Marathon and Solitario uplifts, and in the Sierra Diablo, Hueco, and Franklin Mountains. In the Marathon region, the principal site of Carboniferous deposition in west Texas, deepwater sandstone and shale (flysch) are capped by shallow-water shale, limestone, and conglomerate. Mississippian and Pennsylvanian rocks in the Franklin and Hueco Mountains are chiefly limestone, marl, and shale.

Carboniferous geology of central and north-central Texas is closely tied to the tectonic development of the Fort Worth (foreland) basin, the eastern shelf of the Midland basin, and the Red River uplift–southern Oklahoma mountains. In response to Late Mississippian and Early Pennsylvanian structural activity in the Ouachita geosyncline, the Fort Worth basin became well defined. Thick, westward-prograding terrigenous clastic wedges (Atoka Group) of probable fan-delta and related slope origin entered the basin along high-gradient paleoslopes from the Ouachita foldbelt. Fan deltas shifted westward over thin, relatively starved basinal Smithwick facies. Platform and shelf-edge carbonate environments (Marble Falls, Big Saline, Comyn, and Caddo) contemporaneously dominated the Concho platform. Eastern shelf edges of the Concho platform retreated periodically westward in response to westward-shifting fan-delta environments. Fan deltas reached the western flank of the Fort Worth basin late in the waning stages of Atoka deposition.

Decreased subsidence of the Fort Worth basin and diminished Atoka sediment supply marked the deceleration of Ouachita orogenic activity. Consequently, Strawn (Desmoines Series) deposition was dominated by fluvial-deltaic systems that overlapped shelf-edge carbonate facies (Marble Falls, Caddo, Big Saline) and prograded repeatedly across the shallow Concho platform. Youngest Smithwick prodelta-basinal facies were deposited in the path of the initial delta system to prograde over the Concho platform.

As the source areas were lowered by erosion and as paleogradients were diminished, less terrigenous sediment reached the Pennsylvanian coastline. Extensive and long-lived carbonate-bank and reef systems of the Canyon Group (Missouri Series) began to form on the stable platforms provided by abandoned Strawn deltas. Some of the carbonate banks, growing on the western edge of the structurally positive Eastern shelf of the rapidly subsiding Midland basin, extended upslope and intertongued with Canyon delta systems. Rejuvenation in the Ouachita foldbelt and eastern Fort Worth basin significantly increased sediment supply and initiated extensive lower Cisco (Virgil Series) delta-fluvial deposition. Cisco deltas prograded westward across the relatively stable Eastern shelf, overlapping Canyon carbonate facies and supplying sediment to thick, basinward-prograding slope and submarine-fan environments in the Midland basin. During deposition of the upper Cisco Group (Permian, Wolfcamp Series), sediment supplied from the east again diminished, and thick, low-relief, shelf-edge limestone banks became increasingly prominent.

The complex history of the Red River uplift-Oklahoma mountains structural elements is recorded by thick clastic wedges extending southward and southwestward into the subsurface of north-central Texas. These arkosic sediments represent fluvial and fan-delta deposition along steep paleoslopes adjacent to fault blocks in north Texas and southern Oklahoma. Fan-delta deposition was contemporaneous with limestone deposition on adjacent, structurally positive blocks.

The Marathon region of west Texas was the site of slope and deep basinal sedimentation during most of the Paleozoic. Radiolarian chert and shale in the upper part of the Caballos Novaculite (Lower Mississippian?) were deposited probably in water depths greater than 1,000 m. Black mud of the Tesnus Formation was followed by deposition of thin distal turbidities and shale that reflect progradation of a delta system from east to west. Siliciclastic detritus was derived almost entirely from Llanoria (Africa and South America).

Uplift of the western margin of the Ouachita geosyncline initiated an episode of calcareous flysch deposition (Dimple Formation). Sediment derived from carbonate banks on the shelf and from uplifted older rocks was transported into the basin as slides, debris flows, and turbidity currents. Renewed uplift of Llanoria brought a return to siliciclastic flysch deposition of the Haymond Formation beginning with black mud and followed by alternating turbidites and pelagites and by an olistostrome. Turbidite-pelagic deposition continued after formation of the olistostrome, but younger sandstone beds locally are burrowed, suggesting that the geosyncline was becoming shallower. The Haymond passes

[1] Bureau of Economic Geology, The University of Texas at Austin; presently with Camp Dresser and McKee, Inc., Austin, Tex.
[2] Bureau of Economic Geology, The University of Texas at Austin.
[3] Department of Geological Sciences, The University of Texas at Austin.

upward into shelf and slope deposits of the Gaptank Formation in the northern part of the outcrop belt.

Carboniferous rocks in central, north-central, and west-central Texas have contributed significantly to the economic development of Texas and to the Nation. Oil and gas production has dominated the economic picture, but industrial and ceramic clays, coal, and constructional limestone historically have been locally and periodically important. Potentially, uranium may be found within Carboniferous rocks in sufficient quantities to warrant further intensive exploration. Ground-water potential from Carboniferous rocks is poor and has not been significantly exploited. Resources of economic value have not been recognized in the Marathon basin or in the Franklin and Hueco Mountains.

INTRODUCTION

LOCATION AND EXTENT

Carboniferous rocks in Texas crop out in the Colorado, Brazos, and Trinity River valleys in central and north-central Texas and in Trans-Pecos Texas (figs. 1 and 2). Mississippian and lowermost Pennsylvanian rocks are exposed in isolated areas within and as relatively continuous exposures along the eastern, northern, and western margins of the central mineral region (Llano uplift; figs. 2 and 3). In north-central Texas, younger Pennsylvanian rocks are exposed in a north-northeast-trending belt from the central mineral region to the Red River (fig. 1). Carboniferous outcrops in central and north-central Texas cover approximately 15,675 km^2.

Within the Trans-Pecos region (fig. 1) of Texas, Mississippian and Pennsylvanian rocks are exposed in the Marathon uplift (400 km^2), Solitario uplift (4 km^2), Sierra Diablo Mountain (1 km^2), Hueco Mountains (20 km^2), and Franklin Mountains (10 km^2). Rocks that crop out in the Marathon and Solitario uplifts are similar, and only the Marathon uplift is discussed. Outcrops in the Sierra Diablo Mountain, which are poorly understood, are not included in this report.

In central and north-central Texas, a maximum of 1,333 m of Pennsylvanian rocks is exposed, but only 15 m of Mississippian rocks crops out locally in paleosinkholes. Approximately 2,800 m of Pennsylvanian rocks and 1,500 m of Mississippian rocks are exposed in the Marathon uplift of west Texas. In the Franklin Mountains, 965 m of Pennsylvanian rocks and 150 m of Mississippian rocks have been measured, and in the Hueco Mountains, approximately 400 m of Pennsylvanian rocks has been recognized. The basal part of the Mississippian section in the Hueco Mountains is not exposed.

This chapter was critically reviewed by Shirley J. Dutton, David K. Hobday, and Mark W. Presley of the Bureau of Economic Geology.

The stratigraphic nomenclature used in this paper has not been reviewed by the Geologic Names Committee of the U.S. Geological Survey. The nomenclature used here conforms with the current usage of the Bureau of Economic Geology, The University of Texas at Austin.

GENERAL GEOLOGY

In central and north-central Texas, Mississippian and Pennsylvanian strata consist of shale, sandstone, and limestone (fig. 4). Locally, conglomerate and coal deposits are found in the clastic sequences. Vertical distribution of rock types is generally predictable and provides the basis for stratigraphic classification. Pennsylvanian strata were deposited principally in deltas and interdeltaic embayments, on open shelves and carbonate platforms, and in slope and basin environments. Some Mississippian shale accumulated in relatively shallow starved basins.

Outcropping Mississippian and Pennsylvanian rocks in the Marathon uplift consist of about 3,600 m of deepwater (flysch) deposits capped by about 550 m of shallow-water shale, limestone, and conglomerate (fig. 4). Sandstone and shale make up 85 percent of the flysch units. One flysch sequence, however, the Dimple Formation, is characterized by calcarenite turbidites, chert, and shale. Olistostromes (boulder beds of submarine debris-flow origin) are found in all flysch deposits. Radiolarian chert beds in the upper part of the Caballos Novaculite may be of Mississippian age.

Mississippian and Pennsylvanian rocks exposed in the Franklin and Hueco Mountains (fig. 4) are chiefly limestone, cherty limestone, marl, and shale; chert conglomerate and sandstone are minor rock types. Mississippian sequences contain more shale and marl (15–20 percent) than Pennsylvanian sequences.

PREVIOUS WORK

The presence of Carboniferous rocks in Texas was first reported by Ferdinand Roemer (1848) on the basis of observations made during extensive travels in central Texas during German colonization. A few years later, Roemer (1852) published descriptions of Carboniferous fossils collected at two localities, probably in the Canyon Group and in the Marble Falls Formation (Gries, 1970). Other early explorers who reported Carboniferous-age rocks in Texas include Schumard (1854), Marcou (1856), Shumard (1860), Ashburner (1881), and Glenn (*in* Comstock, 1890).

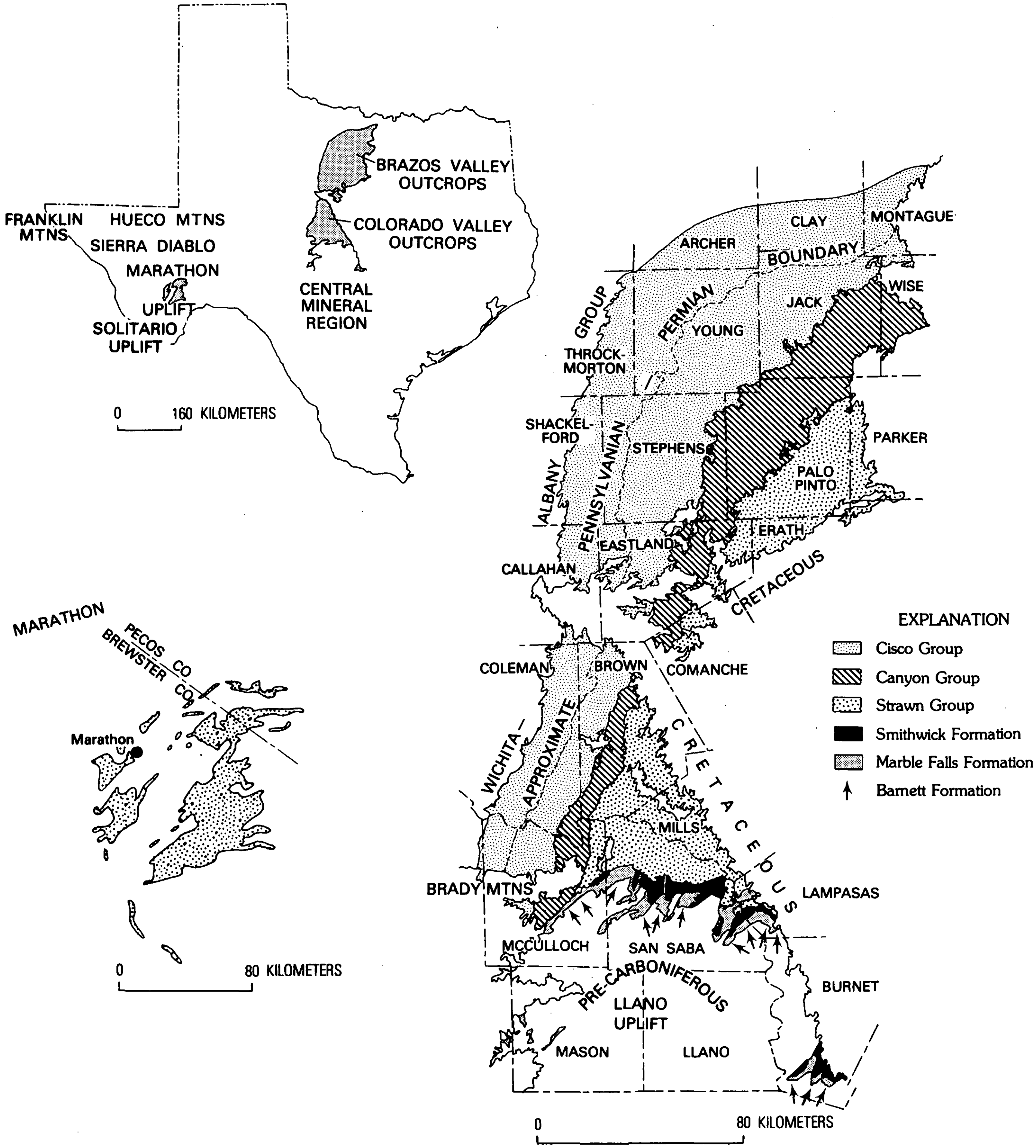

FIGURE 1.—Location of outcropping Mississippian and Pennsylvanian rocks in Texas.

Earliest systematic studies of Carboniferous rocks in Texas were published in annual reports of the third Geological Survey of Texas (1889–1901). In the First Annual Report, Dumble (1890) divided Carboniferous rocks of central Texas into "series," including the Bend series, Richland-Gordon Sandstones, Milburn-Strawn series, Brownwood-Ranger series, Wildrip-Cisco series, and Coleman-Albany

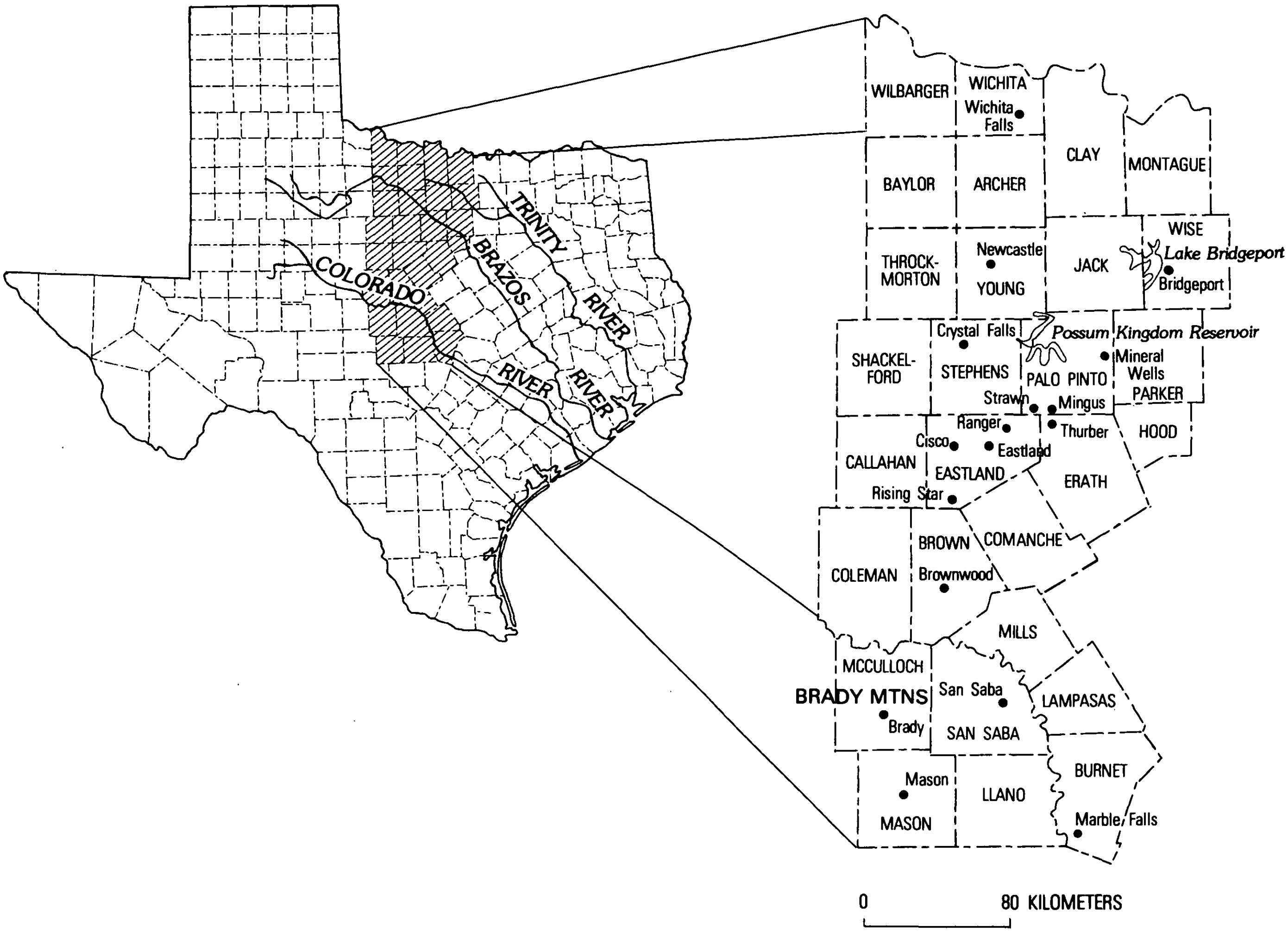

FIGURE 2.—Geographic index to central and north-central Texas.

series. Cummins (1890) and Tarr (1890) described the results of field explorations of the Colorado River valley outcrops. In the Second Annual Report, Cummins (1891) described the general geology of central and north-central Texas, defining five Carboniferous and four Permian "divisions." He also compared the stratigraphy of the central (Colorado River valley) coal fields with that of the northern (Brazos River valley) coal fields. In the Fourth (and last) Annual Report, Drake (1893) described the geology of the "Colorado Coal Field" and named or numbered numerous "beds," many of which subsequently became formal members and formations for central and north-central Texas.

Most early geologic investigations of Carboniferous strata in Texas were centered in central and north-central Texas (for example, Hill (1889, 1901), Paige (1911, 1912), Udden (*in* Udden and others, 1916), and Bridge and Girty (1937)). Plummer (1919) proposed a preliminary classification of Carboniferous rocks for the Brazos River valley. Later, Plummer and Moore (1921; see also Moore and Plummer, 1922) presented a more comprehensive lithostratigraphic classification of Pennsylvanian rocks in the Colorado and Brazos River valleys, including formations and groups. Sellards and others (1932) modified Plummer and Moore's classification, but continued to classify on the basis of lithologic characteristics.

Cheney (1940, 1947, 1948, 1949, and 1950; West Texas Geol. Soc., 1951; Cheney and others, 1945) initiated a major change in the approach to Pennsylvanian stratigraphic classification in Texas. Following concepts of Moore (1936) in Kansas, Cheney proposed a provincial time-stratigraphic classification of Pennsylvanian rocks in central and north-central Texas based on inferred faunal changes and unconformities which were thought to be regionally significant. The Strawn, Canyon, and Cisco were used as series names, and a new Lampassas series

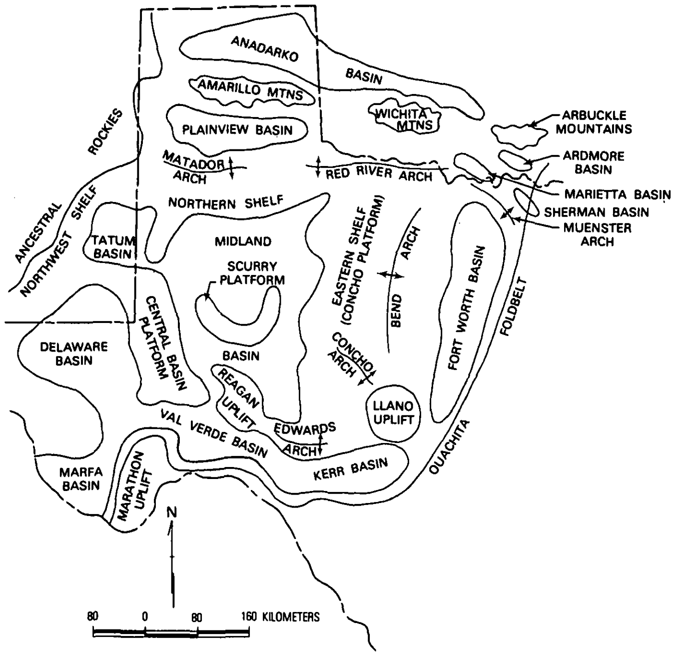

FIGURE 3.—Carboniferous tectonic elements in Texas and southern Oklahoma. Modified from Wermund and Jenkins (1970) and Galloway (1970).

was proposed to encompass strata of the Big Saline and Smithwick "Groups". Conceptually, this classification was intended to facilitate correlation between different basins or areas of exposure within a basin. Although several attempts have been made to apply Cheney's classification scheme in the field (Cheney and Eargle, 1951; Shelton, 1958), such application has proven difficult, if not completely inappropriate (see Brown, 1959; Brown and Goodson, 1972).

Other contributors to the stratigraphy of Mississippian and Pennsylvanian rocks in central and north-central Texas include Plummer (1945; 1947a, b; and 1950), Eargle (1960), Stafford (1960), Terriere (1960), and Myers (1965). Students at The University of Texas at Austin, Baylor University, Southern Methodist University, and Texas Christian University mapped Mississippian and Pennsylvanian rocks and described fossils in central and north-central Texas as part of thesis studies. Recent contributors to Pennsylvanian geology in the region include Bretsky (1966), Brooks and Bretsky (1966), Brown (1960a, b; 1962; 1969a, b, c, and d), Brown and others (1973), Feray and Brooks (1966), Galloway and Brown (1972 and 1973), Laury (1962), Wermund (1966, 1969, and 1975), Wermund and Jenkins (1964, 1969, and 1970), Erxleben (1975), and Cleaves (1975). Regional surface mapping of Mississippian and Pennsylvanian rocks was carried out by Brown and Goodson (1972) and by Kier and others (1976 and unpub. data).

Investigation of Mississippian and Pennsylvanian strata in Trans-Pecos has not been as extensive as in central and north-central Texas. Most studies have been concentrated in the Marathon uplift (figs. 2 and 3). Earliest studies of Pennsylvanian and Mississippian strata were by Baker (*in* Udden and others, 1916) and by Baker and Bowman (1917). Later, King (1931, 1934, and 1937) published a series of reports on Trans-Pecos geology including his landmark study of the Marathon uplift. More recent studies include those by Fan and Shaw (1956), Berry and Nielson (1958), Cotera (1962 and 1969), Johnson (1962), McBride (1966 and 1970), Ross (1962, 1963, 1965, 1967, and 1969), and McBride and Thompson (1970). Additional references on the Marathon region and the Franklin and Hueco Mountains are listed in table 1.

REGIONAL SETTING

CENTRAL AND NORTH-CENTRAL TEXAS

Mississippian and Pennsylvanian rocks exposed in central and north-central Texas were deposited (1) on the Llano uplift (fig. 3), a large structural dome (Cloud and Barnes, 1948) that has a core of Precambrian igneous and metamorphic rocks; (2) on the moderately stable Concho platform during Late Mississippian and Early Pennsylvanian (Cheney, 1929; Cheney and Goss, 1952); and (3) on the eastern shelf of the west Texas basin during Middle and Late Pennsylvanian. East of the Llano uplift and the Concho platform are the Fort Worth basin and the Ouachita foldbelt (Flawn and others, 1961); south of the uplift is the Kerr basin.

During Early Mississippian, limestone, shale, and chert breccia accumulated on a pre-Carboniferous karstic erosion surface on the Llano uplift. Middle and Late Mississippian shale accumulated in a starved basin west of the Ouachita geosyncline.

Beginning in Late Mississippian and during Early Pennsylvanian, orogenic activity in the Ouachita geosyncline produced a thrust-faulted foldbelt (fig. 3). The Ouachita Mountains served as a source of sediment during the rest of the Paleozoic. During the same time, the Fort Worth basin formed as a foreland trough between the rising Ouachita Mountains and the older Concho platform. The basin was initially filled by thick terrigenous clastic wedges of mudstone and sandstone deposited by prograding fan deltas and related slope systems. Clastic wedges grade westward into starved basinal shale of the

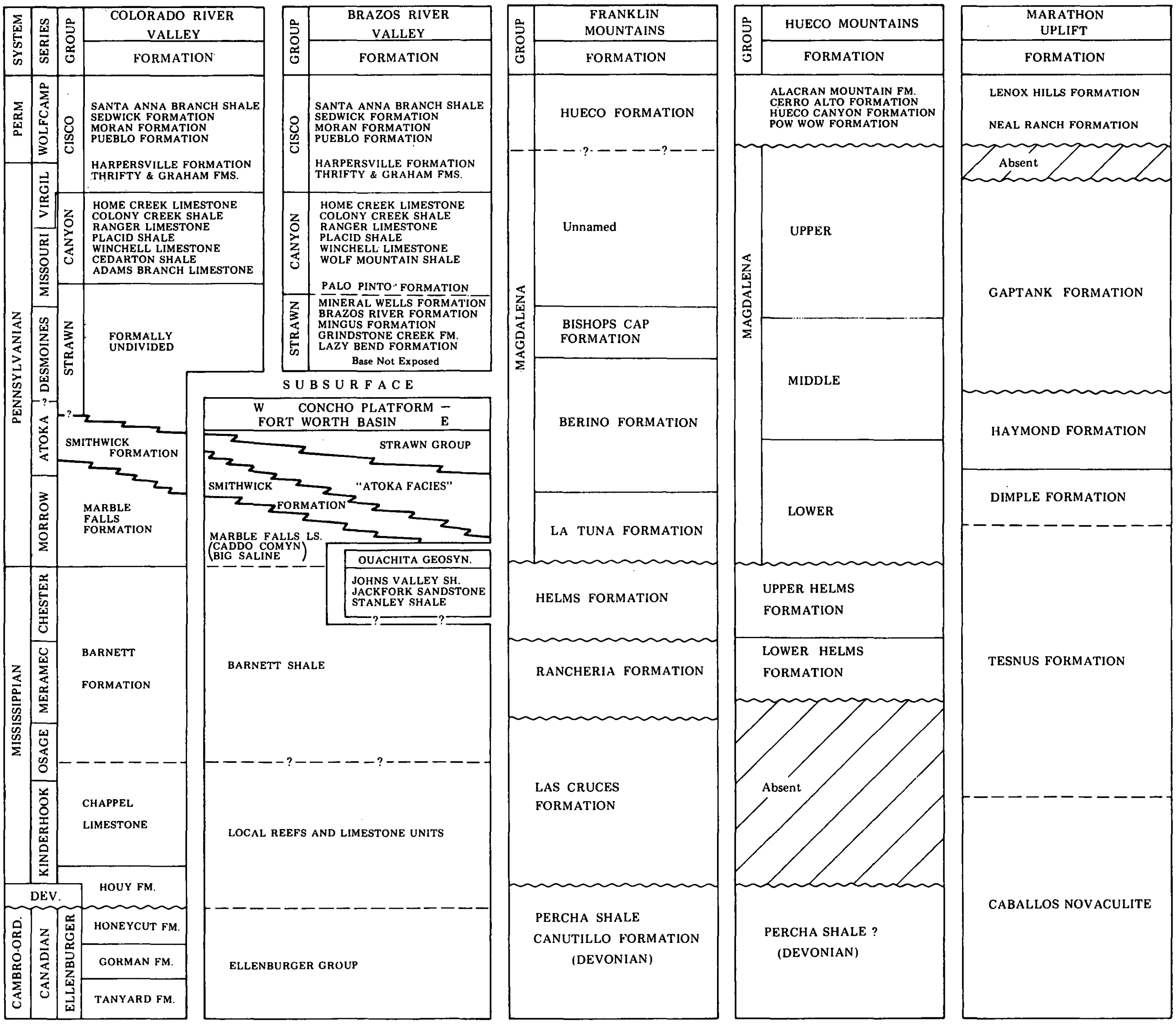

FIGURE 4.—Geologic units in the Carboniferous of central, north-central, and west Texas.

western Fort Worth basin and eastern Concho platform.

The Red River, Muenster, and Matador arches (fig. 3) are a discontinuous series of uplifted fault blocks across northwest Texas that served as foundations for major carbonate sequences and, from time to time, as a source of arkosic sediment. The Wichita, Arbuckle, and Amarillo Mountains of southern Oklahoma and the Texas Panhandle provided arkosic sediment to fluvial and fan-delta systems in north-central Texas.

The Midland basin, a moderate-sized interior basin, formed in west Texas during and after Ouachita deformation (Galloway and Brown, 1972). Fluvial and delta systems originating in the Ouachita Mountains and the uplifted eastern part of the Fort Worth basin prograded westward across the earlier Concho platform. Cyclic deposition by these clastic systems progressively constructed the eastern shelf of the Midland basin. Carbonate shelf and shelf-edge systems formed along the westward margin of the Eastern shelf as the depth of the basin increased and slope deposition was initiated.

Cyclic progradation of fluvial and delta systems westward across shelf and shelf-edge carbonate environments continued throughout the Middle and

TABLE 1.—*Contributors to stratigraphic data on the Marathon basin and the Franklin and Hueco Mountains*

Marathon basin	Franklin and Hueco Mountains
Aberdeen, 1940	King, 1934
Baker, 1963	King and others, 1945
Brooks, 1955	Laudon and Bowsher, 1949
Ellison, 1962	Nelson, 1940
Flawn, 1958	Seewald, 1968
Flawn and others, 1961	Stewart, 1958
Flores, 1972, 1977	Williams, 1963
Folk, 1973	
Goldstein and Hendricks, 1962	
King and King, 1929	
King and others, 1931	
Skinner and Wilde, 1954	
Thomson and Thomasson, 1969	
Waterschoot van der Gracht, 1931	

Late Pennsylvanian and Early Permian. Large volumes of sediment were transported across the Eastern shelf through fluvial and deltaic channels. Where deltas reached the shelf edge, shallow-marine sediments were redeposited in submarine fans on the floor of the Midland basin. Relief between the floor of the Midland basin and the Eastern shelf reached 455 m in the Late Pennsylvanian (Brown, 1973a).

As the Midland basin subsided, regional upwarping of the Ouachita foldbelt and the eastern flank of the Fort Worth basin took place. The hinge or axis of rotation between the rising Fort Worth basin and the downwarping Midland basin defines the Bend arch. Erosion of uplifted Lower Pennsylvanian sediments in the Fort Worth basin contributed considerable amounts of second-cycle sediment to Upper Pennsylvanian delta and slope environments.

Principal tectonic elements such as the Llano uplift, Ouachita foldbelt, Fort Worth basin, Concho platform, Eastern shelf, Bend arch, Red River arch trend, Oklahoma mountains, and Midland basin significantly determined the nature of depositional environments in which Mississippian and Pennsylvanian sediments accumulated. The interplay among orogenic pulses in the Ouachita and Oklahoma mountains, uplift of the eastern flank of the Fort Worth basin, subsidence of the Midland basin and westward tilting of the Concho platform affected sediment supply and water depth. Depositional processes operative in a myriad of fluvial, deltaic, embayment, shelf, platform, slope, and basin environments produced the many Mississippian and Pennsylvanian rock types.

WEST TEXAS

Pre-Permian Paleozoic rocks of the Marathon uplift in Trans-Pecos Texas were deposited in the Marathon trough, a segment of the Ouachita geosyncline that extended from Mexico to Arkansas (fig. 3). Although there are dissenting opinions (Folk, 1973; Flores, 1972, 1977), the Marathon trough apparently was a deep basin throughout pre-Permian-Paleozoic time. From Late Cambrian to Carboniferous time, the trough received about 1,000 m of slope and basinal sandstone, limestone, chert, shale, and olistostromes. This "early geosynclinal" phase of slow deposition culminated with deposition of the Caballos Novaculite.

Deposition of predominantly flysch rocks generally coincided with the beginning of the Carboniferous. During this "late or filling stage" of geosynclinal history, approximately 3,500 m of flysch deposits (Tesnus, Dimple, and Haymond Formations) and 550 m of shallower marine deposits (Gaptank Formation) accumulated in only 60 m.y. During the filling stage, most terrigenous detritus was derived from a continental mass east of the geosyncline, an element designated Llanoria by Dumble (1920) and other early workers but considered part of Africa or South America in newer plate-tectonic restorations (Rowett and Walper, 1973; Keller and Cebull, 1973). Locally, carbonate detritus and exotic blocks were derived from a positive cratonic element west of the depositional basin (McBride, 1970).

The Ouachita geosynclinal sequence underwent several pulses of deformation and mountain building from Desmoines to Middle Wolfcamp(?) time (King, 1937; Ross, 1962). The sequence was folded, faulted, and thrust northwestward at least 70 km along a major décollement surface; the deformed sequence is underlain by relatively undeformed foreland rocks. The Marathon region underwent broad domal uplift and normal faulting early in Tertiary time (King, 1937).

The Franklin Mountains and Hueco Mountains are north-trending Laramide fault blocks. About 1,700 m of Paleozoic rocks, chiefly limestone, is exposed in the Franklin Mountains. About 1,000 m of Silurian and younger Paleozoic rocks, chiefly limestone and shale, is exposed in the Hueco Mountains.

CARBONIFEROUS STRATIGRAPHY

CENTRAL AND NORTH-CENTRAL TEXAS

Major lithologic divisions of the Carboniferous of central and north-central Texas (fig. 4) are based on variations in the vertical succession of the stratigraphic sequence. These principal sequences record filling of the Fort Worth basin, cyclic sedimentation on the Eastern shelf, and partial filling of the Midland basin (fig. 3).

The Lower Mississippian (and Devonian) Houy Formation is a relict or lag deposit preserved in sinks and depressions in an erosion surface on the Llano uplift. Middle and Upper Mississippian Chappel Limestone and shale of the Barnett Formation were deposited during initial marine transgression onto the Llano uplift. In outcrop, the Lower Pennsylvanian Marble Falls Formation, predominantly limestone, represents establishment of carbonate platform and shelf environments on the Llano uplift adjacent to the Fort Worth basin. Similar Upper Mississippian and Lower to Middle Pennsylvanian shelf and shelf-edge facies were deposited west of the Fort Worth basin on the Concho platform.

The Smithwick Shale and the overlying Strawn Group record the final phase of filling of the Fort Worth basin and initial westward progradation of deltas onto the Concho platform (Cleaves, 1973, 1975). Diminished terrigenous influx and increased carbonate shelf and bank deposition distinguish the Canyon Group from the underlying Strawn and overlying Cisco Groups. Several medium to thick limestone units that were deposited in platform and open-shelf environments intertongue updip (eastward) with deltaic deposits and grade downdip into shelf-edge reef and bank deposits at the eastern margin of the Midland basin (Erxleben, 1973, 1975).

The Cisco Group records renewed deposition of terrigenous clastic materials and predominance of fluvial and deltaic environments. From 10 to 15 fluvial-deltaic progradational sequences can be recognized in the Cisco Group, each one terminated upward by transgressive sandstone and open-shelf limestone facies (Brown, 1973b). Downdip the clastic facies intertongue with extensive shelf and shelf-edge limestone deposits. Fluvial-deltaic deposits were the principal sources of sediment that ultimately was redeposited by density flows to produce thick, off-lapping wedges of deepwater deposits in the eastern part of the Midland basin (Galloway and Brown, 1972, 1973).

MISSISSIPPIAN SYSTEM

HOUY FORMATION

The Houy Formation established by Cloud and others (1957) includes strata transitional across the Devonian-Mississippian boundary. It is divided into two formal members, the Ives Breccia and the Doublehorn Shale, and several unnamed members.

The basal Ives Breccia Member is a poorly sorted, multicolored angular to subangular chert breccia with a matrix of medium to coarse, angular to subangular chert and clear quartz sand. Silicified crinoid fragments and conodonts are fairly common (Kier, 1972). Hematite, partly weathered to limonite, is very common. Maximum thickness of this member is about 1 m.

The overlying Doublehorn Shale Member is black, fissile, slightly radioactive shale that contains spores of unknown origin and silicified pieces of *Callixylon* (Cloud and others, 1957). It weathers light brown and is as much as 4.5 m thick. Other unnamed lithic units in the Houy Formation include siliceous limestone and silty calcareous shale below the Ives and phosphorite beds above the Doublehorn Shale.

The Ives Breccia Member crops out on the eastern, northern, and western sides of the Llano uplift (fig. 3). The Doublehorn Shale and other unnamed members in the Houy are found only on the northeastern side of the uplift. Poor exposure and isolation of the outcrops make interpretation difficult.

The Ives Breccia apparently is a lag deposit of locally derived chert weathered from limestone and dolomite beds of the underlying Ellenburger Group and deposited during one or more Upper Devonian and Lower Mississippian marine transgressions (Zachry, 1969; Kier, 1972). The Doublehorn Shale may be the offshore facies equivalent of the Ives (Kier, 1972). Cloud and others (1957) placed the Devonian-Mississippian boundary within the Ives, and Seddon (1970) suggested that the boundary is a disconformity and that the Doublehorn Shale is, in part, time-equivalent to the Ives. Also, at least part of the Ives Breccia may have been deposited more or less contemporaneously with the Chappel Limestone (Kier, 1972).

CHAPPEL LIMESTONE

The Chappel was named by Sellards (Sellards and others, 1932) for thin crinoidal limestone beds lying directly on the Ellenburger Group. It is predominantly a fine- to very coarse grained, poorly sorted, packed, ostracode-bearing, algal, crinoidal biomicrite and poorly washed biosparite (Kier, 1972). Most of the Chappel is light to medium greenish gray or dusky yellow. On the basis of conodonts, Hass (1959) concluded that the age of the Chappel is late Kinderhook to early Osage.

Chappel outcrops are scattered across the eastern, northern, and western sides of the Llano uplift. Commonly, the Chappel lies within or adjacent to sinks in the Ellenburger Group. Whether these sinks formed before, during, or after deposition of the Chappel Limestone is uncertain (Cloud and Barnes, 1948; Freeman, 1962; Turner, 1970). Thickness of the unit varies from about 10 cm to about 10 m.

The Chappel Limestone is apparently conformable with the underlying Doublehorn Shale Member and possibly the Ives Breccia Member of the Houy Formation. Seddon (1970) reported a continuous succession of conodonts from the Doublehorn Shale into the overlying Chappel. Locally, the Ives Breccia is adjacent to, within, or above the Chappel (Turner, 1970; Seddon, 1970; and Kier, 1972). Elsewhere, the Chappel Limestone unconformably overlies the Ellenburger Group.

Sellards, (*in* Sellards and others, 1932), Cloud and Barnes (1948), and Freeman (1962) stated that the overlying Barnett Formation was deposited unconformably on the Chappel Limestone. More recently, Zachry (1969), Turner (1970), and Kier (1972) concluded that the Barnett is probably conformable with the Chappel. Shale beds within the Chappel Limestone are similar to Doublehorn and Barnett shales (Rose, 1959; Winston, 1963; Turner, 1970; Kier, 1972). All investigators except Winston concluded that the Chappel is the shoreline or nearshore equivalent of the Barnett Formation. Deposition may have taken place during more than one marine transgression (Zachry, 1969; Kier, 1972). Whether the present distribution of Chappel outcrops approximates the original distribution of Chappel depositional environments or whether Chappel environments were much more extensive is uncertain.

BARNETT FORMATION

The Barnett Formation and subjacent Chappel Limestone lie between the Ordovician Ellenburger Group and the Pennsylvanian Marble Falls Formation and document major Mississippian marine transgressions across the Llano uplift. The Barnett was established by Plummer and Moore (1921). Previously, the shale unit was called the lower shale of the Bend Series, or simply the Lower Bend Shale (Girty *in* Paige, 1912; Udden *in* Udden and others, 1916; Moore, 1919).

Along the north side of the Llano uplift east of the town of San Saba (figs. 2 and 3), the Barnett is predominantly a black to olive-gray, very thinly laminated shale that weathers light to dark brown (Kier, 1972). Thin brown microsparite limestone beds, brachiopod and cephalopod coquinas, and large ellipsoidal microsparite concretions as much as 2.75 m in diameter are very common, especially in the upper half of the formation. Locally, the shale is very petroliferous, and freshly broken concretions yield a strong petroliferous odor.

On the northeast side of the Llano uplift, the upper 10–150 cm of the Barnett Formation is commonly a dusky to dark yellowish-brown, fine- to coarse-grained, poorly sorted, packed pelletiferous biomicrite and oomicrite. Cephalopods, brachiopods, conodonts, and ostracodes are abundant; glauconite is present in varying amounts, and most of the allochems are phosphatic. Thickness of the Barnett east of San Saba is 10.6–15.2 m.

West of the town of San Saba, the Barnett is divisible into two parts (Freeman, 1962; Turner, 1970). The lower part of the Barnett is a light-colored clay shale. Concretions are small, and phosphatic limestone is abundant at the top of the lower part of the formation. Farther west toward Brady (fig. 2), thin limestone beds are abundant in the lower part of the Barnett, and the section is progressively more phosphatic and glauconitic. The upper part of the Barnett Shale is grayish-black to yellowish-brown, fine- to coarse-grained, packed, glauconitic and phosphatic biomicrite and micrite. Micrite becomes predominant westward. Thickness of the Barnett west of San Saba is 7.6 m.

On the east side of the Llano uplift near the town of Marble Falls (fig. 2), the Barnett is light-colored shale containing concretions. The formation is capped by as much as 3.3 m of phosphatic limestone (Namy, 1969). Maximum thickness of the Barnett on the east side of the uplift is 6.4 m. On the west side of the Llano uplift near Mason (fig. 2), shale mapped as Barnett Formation (Winston, 1963) is probably Marble Falls (W.C. Bell, oral commun. 1970, *reported in* Kier, 1972).

On the north side of the Llano uplift, Gries (1970) and Schwarz (1975) recognized seven or eight species of ammonoids, four species of nautiloids, and several species of pelecypods and brachiopods. Algae, corals, bryozoans, brachiopods, conodonts, and echinoderm fragments occur in the Barnett on the east and northwest sides of the Llano uplift (Namy, 1969; Turner, 1970). On the basis of cephalopods, Schwarz (1975) assigned a late Osage to Chester age to the Barnett. Using conodonts, Hass (1953), and Defandorf (1960) assigned a late

Osage to early Morrow age to the Barnett, spanning the Mississippian-Pennsylvanian boundary.

The placement and nature of the upper contact of the Barnett Formation have been the subject of dispute. Most recent workers place the Barnett-Marble Falls contact between shale or phosphatic limestone and nonphosphatic limestone. Earlier workers did not include the phosphatic limestone in the Barnett (see Kier, 1972) and believed that the Barnett-Marble Falls contact is unconformable because (1) it generally coincides with the Mississippian-Pennsylvanian boundary; (2) in places the Barnett is absent, and Marble Falls limestone beds lie directly on Ellenburger limestone or dolomite beds; and (3) glauconite and phosphate are commonly concentrated at the contact. Zachry (1969), Turner (1970), and Kier (1972), however, noted gradational or interbedded Barnett and Marble Falls rock types at the contact. Only Namy (1969) presented good physical evidence for an unconformity between the Barnett and the Marble Falls. Nevertheless, on the basis of conodonts, Liner and others (1977, in press) inferred that the Barnett-Marble Falls contact on the northeast side of the Llano uplift represents a hiatus from middle Chester to Morrow. To the west, however, they found that the Mississippian-Pennsylvanian boundary and the inferred hiatus are within the Barnett Shale, below the sequence of phosphatic limestone.

The Barnett Formation probably accumulated in a sediment-starved basin under euxinic conditions. Evidence includes the lithic character and general absence of benthonic fossils, particularly an infauna, and the inferred length of time represented by the thin unit. The Barnett of the Llano uplift was probably deposited within an extension of the early, sediment-starved Fort Worth basin (Brown, 1973a). Maximum water depth was undoubtedly below wave base, but still relatively shallow.

Thin microsparite and coquina layers in the Barnett Formation probably reflect temporary cessation of euxinic conditions. Phosphatic beds at the top of the Barnett may record a gradual change from euxinic restricted conditions to open-marine shelf and platform environments characteristic of Marble Falls deposition (Kier, 1972). Yellowish-brown shale and limestone beds exposed on the eastern and northwestern sides of the Llano uplift were deposited under less reducing conditions at the basin margins. Where Barnett shale is thin or missing between the Marble Falls and the Ellenburger, it suggests lack of deposition rather than post-Mississippian erosion.

PENNSYLVANIAN SYSTEM

MARBLE FALLS FORMATION

The Marble Falls Formation records reestablishment of normal marine conditions and widespread limestone environments over the Llano uplift and the adjacent Concho platform. Platform, open-shelf, and shelf-edge carbonate deposition dominated the western margin of the rapidly subsiding Fort Worth basin. These carbonate environments ultimately shifted westward as they were progressively displaced by advancing Smithwick and Strawn deltaic environments.

Marble Falls was introduced by Hill (1889) for "Encrinoidal" limestone exposed along the Colorado River near the town of Marble Falls in Burnet County (fig. 2). Later, Hill (1901) concluded that the Marble Falls is correlative with limestone of the "Bend division" exposed along the north side of the Llano uplift, although the two outcrop areas are not physically connected. Various names and stratigraphic ranks have been proposed for the Marble Falls Formation at the surface and in the subsurface of central, north-central, and west-central Texas; Comyn, Big Saline (Cheney, 1940) and Sloan (Plummer, 1945, 1947a; see Kier, 1972). Most investigators have retained the name Marble Falls for surface exposures on and around the Llano uplift.

Except on the east side of the Llano uplift near the town of Marble Falls, the Marble Falls Formation can be subdivided into two outcropping units separated by an unconformity (Freeman, 1962; Namy, 1969; Zachry, 1969; Turner, 1970; Kier, 1972). The lower part of the Marble Falls is Morrow in age throughout its outcrop. The upper part of the formation, however, becomes progressively younger westward. On the east and northeast sides of the Llano uplift the upper Marble Falls is Morrow in age (Namy, 1969; Zachry, 1969; Kier, 1972); just west of the town of San Saba (fig. 2), on the north side of the uplift, the formation is Morrow and Atoka in age (Turner, 1970); and near Brady on the northwest side of the uplift, the formation is entirely of Atoka age (Freeman, 1962). The upper and lower parts of the Marble Falls accumulated under different depositional conditions. The subsurface extent of the unconformity within the Marble Falls is uncertain.

The lower part of the Marble Falls consists predominantly of light to dark cherty limestone and thin shale beds. Principal limestone types include algal biomicrite and biosparite, oosparite, spiculitic biomicrite, pelmicrite, micrite, and mixed skeletal fragment biomicrite and biosparite. Locally, coral and algal biolithite are found. Diagenetic alteration of limestones within the Marble Falls is limited generally to recrystallization of micrite to microspar or pseudospar and to inversion of aragonite to calcite.

Facies patterns in the lower part of the Marble Falls Formation are complex, and there is considerable local variation within individual exposures. Nevertheless, facies appear to have been arranged originally in a semicircular pattern around the Llano uplift. The structurally positive uplift apparently acted as the core for a major carbonate platform. Vertical accretion dominated over lateral accretion of carbonate facies, and high-energy facies tended to expand in areal extent at the expense of lower energy facies. Depositional relief was as much as 9 m and was a significant controlling factor in determining facies characteristics. Facies patterns at or near the end of deposition of the lower part of the Marble Falls demonstrate net regression. The lower Marble Falls is about 30 m thick but ranges in thickness from about 21 to 45 m. Near Mason, much of the lower Marble Falls Formation was apparently eroded prior to deposition of the upper part of the Marble Falls (W. C. Bell, oral commun., 1971).

The upper part of the Marble Falls Formation is predominantly light to dark algal biomicrite, calcarenite, siliceous spiculitic biomicrite, and shale. Although lithic types are similar to those in the lower part of the Formation, the depositional setting was distinctly different in several respects:

1. Facies patterns are oriented approximately north-south; a distinct semicircular carbonate platform on the Llano uplift cannot be recognized.
2. Shale and spiculitic limestones were deposited over the entire Llano uplift, in contrast to patterns in the lower part of the Marble Falls, where deposition of shale and spiculitic limestones was mostly restricted to off-platform environments.
3. Individual facies are thin and widespread; depositional relief was relatively low.
4. Although high- and low-energy facies were widespread, shifts in facies boundaries are common.
5. High-energy facies become more common upward within the formation, but they are not as common as they are within the lower part of the Marble Falls.
6. Facies patterns in the upper part of the Marble Falls record westward transgression.

The upper part of the Marble Falls Formation is 36–67 m thick in outcrop; average outcrop thickness of the entire Marble Falls Formation is 91 m.

Marble Falls deposition began with establishment of more open, less restrictive conditions than existed during deposition of the Barnett Shale. Incipient calcarenite shoals developed, apparently at some slight break in slope. A major carbonate platform centered about the Llano uplift rapidly formed over the shoals. Platform margins were approximately coincident with present Marble Falls outcrops. On the northeast side of the platform, depositional environments resembled the modern Bahamian platform, although platform/off-platform relief was not nearly so great (Kier and Zachry, 1973; Zachry and Kier, 1973). Ultimately the lower Marble Falls platform either built to sea level or was exposed by a drop in sea level.

The upper part of the Marble Falls Formation was deposited predominantly by algal buildups and calcarenite shoals and by shale and spiculitic biomicrite deposited in somewhat restricted depressions on the open-marine shelf marginal to the Fort Worth basin. When the lower Marble Falls carbonate platform was subaereally exposed, shale and spiculitic biomicrite continued to be deposited on the off-platform shelf between the Llano uplift and the Fort Worth basin (Namy, 1969). Subsidence of the eastern edge of the lower Marble Falls platform allowed shale and spiculitic limestone to onlap the erosion surface, followed by establishment of algal buildups and calcarenite shoals. Marine energy levels and depositional relief were less than during deposition of the lower Marble Falls.

Upper Marble Falls depocenters shifted progressively westward as the lower Marble Falls platform subsided. Lengthy erosion of the platform on the west side of the Llano uplift removed much, and locally all, of the lower Marble Falls Formation (Freeman, 1962) prior to deposition of the upper Marble Falls. Strawn deltas simultaneously prograded across the upper Marble Falls carbonate shelf from the east. The upper contact of the Marble Falls Formation with the overlying Smithwick Shale or Strawn sandstones is essentially conformable.

SMITHWICK FORMATION

The Smithwick Formation represents the initial terrigenous clastic deposits of Atoka and Strawn deltas which prograded across the Fort Worth basin and onto the Llano uplift. Regional subsurface correlations from Hill County to Brown County and from Dallas County to Stephens County (fig. 2) demonstrate the time-transgressive relationships of the Atoka and Strawn Groups and the Smithwick Formation. In outcrop, the Smithwick is the prodelta facies of Atoka and Strawn delta systems. Atoka facies are restricted to the Fort Worth basin; Strawn facies are found within the basin, but they also crop out around the Llano uplift and within the Colorado and Brazos River valleys.

The Smithwick Formation was named by Paige (1911) for shale and sandstone exposed near Old Smithwick in Burnet County. Girty (*in* Paige, 1912) inferred that the Smithwick is equivalent to the upper shale of the "Bend Series" north of the Llano uplift. Although several attempts have been made to restrict the Smithwick (Cheney and others, 1945; Plummer, 1950), application of the name Smithwick to shale overlying the Marble Falls in outcrop and in the subsurface of north and west-central Texas is generally accepted.

The Smithwick Formation consists of black, slightly calcareous, fissile clay-shale and lesser amounts of siltstone and sandstone. Minor amounts of dark limestone and conglomerate are composed of limestone, chert, and sandstone clasts. Sedimentary bed forms—ripple marks, flute casts, groove casts, and slump features—are common on sandstone bedding surfaces, particularly on the east side of the Llano uplift (McBride and Kimberly, 1963). Locally, hematitic concretions are found in the Smithwick.

Although macrofossils are rare in the Smithwick, they are concentrated in a few localities. Gries (1970) identified several species of rugose coral, particularly *Cumminsia aplata*, brachiopods, gastropods, pelecypods, and cephalopods. Turner (1970) observed spicules, arenaceous foraminifers, and plant fragments. The Smithwick is as much as 30 m thick or more on the north side of the Llano uplift (Kier, 1972) and as much as 121 m thick on the east side (McBride and Kimberly, 1963). In general, the Smithwick Formation thins westward; locally it is absent.

In outcrop, the contact between the Smithwick Formation and the Strawn Group is gradational. Upward within the Smithwick, the amount of sandstone increases, and shales become siltier (Kier, 1972). A faunal change that ostensibly represents a significant hiatus between the Smithwick Formation and the Strawn Group (Plummer, 1947a) is undocumented and may simply reflect environmental differences (Kier, 1972). Variations in thickness of Smithwick in outcrop probably reflect original depositional variations, differential compaction, and the effects of contemporaneous faulting. Differences in trends of Marble Falls-Smithwick outcrops and Strawn outcrops are due to contemporaneous structural relief on the Llano uplift, to postdepositional uplift and doming of the southern exposure of three contemporaneous yet time-transgressive facies, and perhaps to westward tilting of the Eastern shelf toward the Fort Worth basin (Kier, 1972).

Smithwick Shale exposed on the north side of the Llano uplift is the prodelta facies of Strawn deltas (Turner, 1970; Kier, 1972). McBride and Kimberly (1963) interpreted a "deep water" environment, periodically invaded by turbidity currents to explain the Smithwick east of the Llano uplift, but McBride (oral commun., 1977) no longer believes the water was particularly deep. Conglomerate in the Smithwick Formation on the east and northwestern sides of the Llano uplift is related to contemporaneous faulting and erosion of the Marble Falls Limestone (Freeman, 1962; Freeman and Wilde, 1964; McBride and Kimberly, 1963).

ATOKA GROUP

Very thick clastic strata were deposited within the Ouachita geosyncline during Late Mississippian and Early Pennsylvanian time (Stanley, Jackfork, and Johns Valley Formations). Near the end of the Morrow series, rejuvenation of structural activity in the Ouachita foldbelt provided an enormous volume of clastic sediment that was transported and deposited within the geosyncline and adjacent Fort Worth basin. The depocenter shifted westward during Atoka and Strawn deposition.

A clastic wedge nearly 2,000 m thick makes up the Atoka Group, which is restricted to the subsurface of the basin. Deposited by westward-prograding fan deltas, the group consists of thick shale and sandstone facies that probably range from shallow marine-deltaic to deeper basin facies. Westward, the Atoka is, in part, gradational with the older parts of the Smithwick Formation within the Fort Worth basin. Atokan clastic rocks may also be partly time-equivalent to upper Marble Falls facies on the Concho platform and Llano uplift. Because the group is restricted to the subsurface, its description is limited in this report.

STRAWN GROUP

The Strawn Group consists predominantly of cyclic terrigenous clastic facies deposited by Middle Pennsylvanian fluvial-deltaic systems that essentially filled and prograded westward across the Fort Worth basin, onto the Concho platform and into the incipient Midland basin. In the Fort Worth basin, the Strawn Group is gradational with the underlying Atoka Group. Only the upper part of the Strawn Group crops out, although Strawn facies on the Llano uplift may be equivalent to lower Strawn in the Fort Worth basin. Regional upwarping of the Ouachita foldbelt and the eastern margin of the Fort Worth basin continued to supply large volumes of eroded Atokan sediments to the Strawn rivers. Fluvial and deltaic deposition dominated the Concho platform and extended onto the Llano uplift. Near the end of Strawn deposition, erosion lowered the source area, and the supply of sediments diminished.

The Strawn Group crops out in both the Colorado River valley and Brazos River valley in central and north-central Texas, respectively (figs. 1, 2, 5, 6). In the Brazos River valley, Strawn rocks have been studied more extensively because of exploration for coal, oil, and gas; exposures are better, persistent limestone beds facilitate subdivision, and abundant marine faunas in the shale sections permitted time-stratigraphic interpretations. Five Strawn formations are recognized in the Brazos River valley (fig. 4) by Brown and Goodson (1972). Few formal subdivisions are recognized in the Colorado River

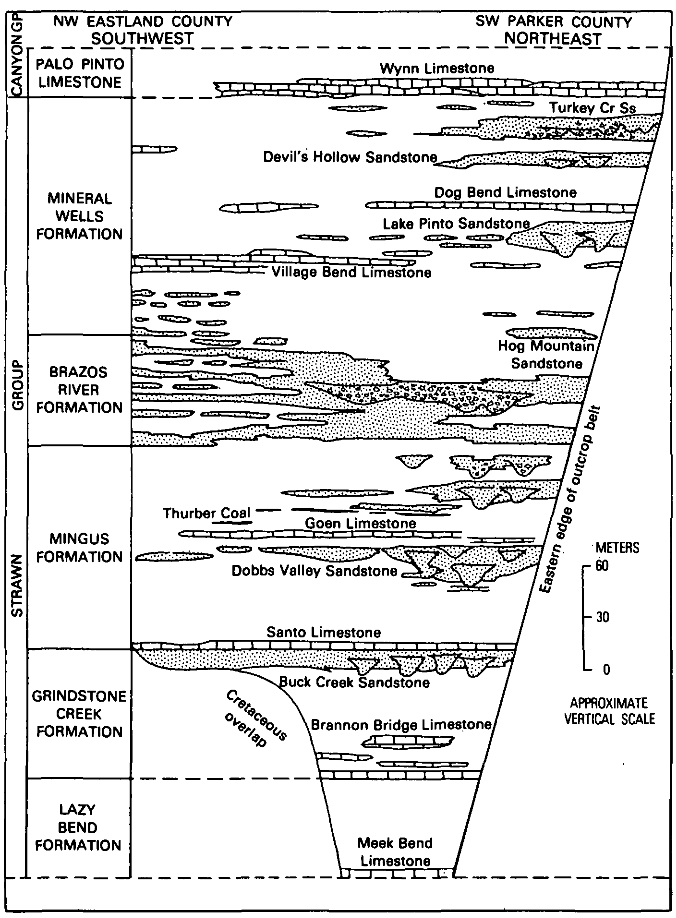

FIGURE 5.—Schematic cross section of outcropping Strawn Group in the Brazos River valley. Modified from Brown and Goodson (1972) and Cleaves (1975).

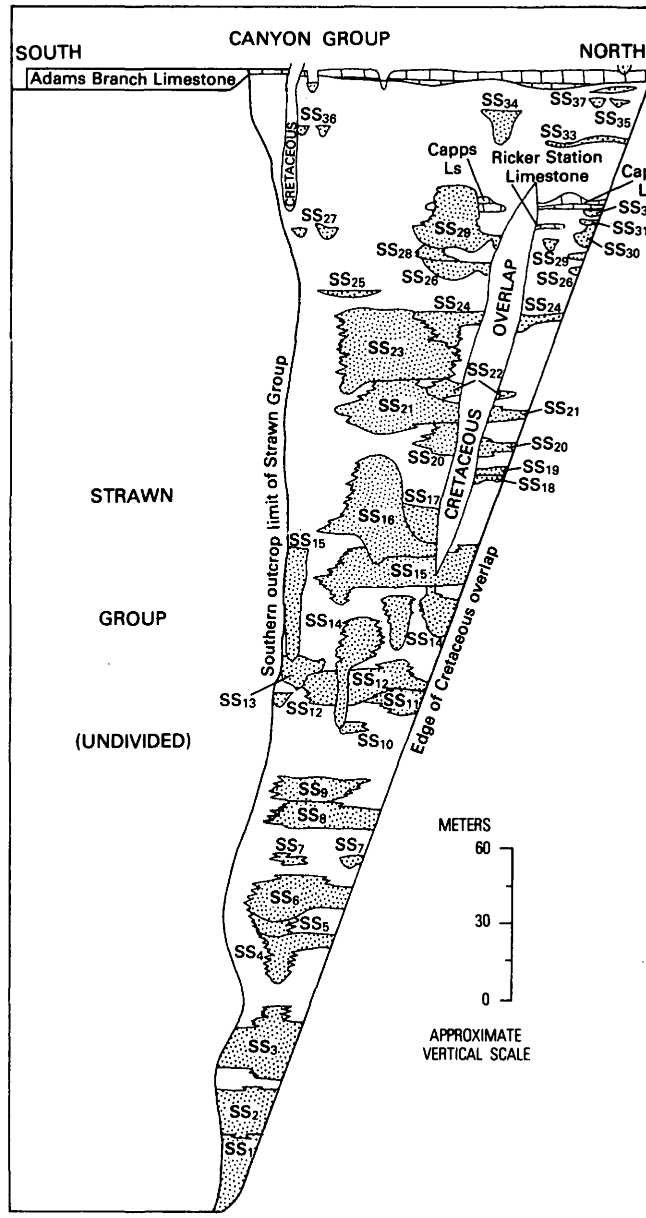

FIGURE 6.—Schematic cross section of outcropping Strawn Group in the Colorado River valley. Modified from Kier and others (1976). Unnamed sandstone units are numbered in approximate stratigraphic order from the base.

valley, and most lithic units have simply been numbered (Drake, 1893; Kier and others, 1976).

Dumble (1890) first applied the name "Strawn series" to coal- and limestone-bearing clay and shale exposed near the town of Strawn in Palo Pinto County (fig. 2). Equivalent strata in the Colorado River valley were apparently called "Milburn series," although in the same report, Tarr (1890) proposed the names "Richland Sandstone" and "Milburn Shales" for Strawn strata in the Colorado River valley. Cummins (1891) defined the "Strawn division" to include strata from the base of Canyon limestones to the base of Coal Seam #1 (Thurber Coal). Strata below Coal Seam #1 to the top of the underlying "Bend division" (Smithwick Formation) were called the "Milsap Division." The Milsap was not recognized in the Colorado River valley, however, where Cummins presumed that the middle part of the Strawn lay unconformably on the Bend. In the Colorado River valley, the top of the Strawn was placed at the base of the limestone (Capps) exposed near Brownwood (figs. 2, 4, 6).

Drake (1893) applied the name Strawn to all strata between the Bend and Canyon "divisions." In the course of his work, Drake mapped and numbered 20 Strawn "beds" cropping out in the Colorado River valley; this map was the only detailed map and subdivision of Strawn rocks in this area prior to publication of the Brownwood Sheet of the Geologic Atlas of Texas (Kier and others, 1976).

Application of the name Strawn underwent considerable evolution following Drake's (1893) work, first as a formation (Smith, 1903; Udden *in* Udden and others, 1916), then as a group (Plummer and Moore, 1921; Plummer, 1929; Scott and Armstrong, 1932; Plummer and Hornberger, 1935), and finally as a series (Cheney, 1940, 1947; Cheney and others, 1945; Spivey and Roberts, 1946; Quigley and Schweers, 1951).

In the 1950's, principal studies on the Strawn Group were by students working under S. P. Ellison at The University of Texas at Austin (Abilene Geological Society, 1954) and Leo Hendricks (1957) at Texas Christian University. The Abilene sheet of the Geologic Atlas of Texas (Brown and Goodson, 1972) used essentially the same nomenclature (figs. 4, 5, 6) as Plummer and Hornberger (1935). Cleaves (1973, 1975) studied the upper part of the Strawn and extended the lithic units mapped by Brown and Goodson (1972) westward into the subsurface. The lower part of the Strawn is not exposed in central or north-central Texas.

The Strawn Group consists predominantly of shale and sandstone and lesser amounts of limestone, coal, and conglomerate. For the group in the Brazos River valley, Cleaves (1973, 1975) made the following interpretations:

1. Shale and sandstone were deposited in delta, prodelta, and embayment environments.
2. Limestone was deposited in open-shelf environments or locally in interdeltaic-bay environments. Open-shelf carbonate rocks, mostly algal biomicrites, are regionally extensive in outcrop and in the subsurface, and they serve as marker beds for delineating formations within the Strawn Group.
3. Coal formed predominantly in marshes and swamps on delta-plain and along interdeltaic-embayment coastlines. Coal crops out principally in the Brazos River valley area, and only the Thurber coal of southern Palo Pinto County (figs. 2, 5) has been successfully mined.
4. Source areas for Strawn terrigenous clastic deposits were the Ouachita Mountains, the eastern part of the Fort Worth basin, and the Arbuckle Mountains of Oklahoma (fig. 3). Locally, the Wichita Mountains and the Criner Hills in southern Oklahoma supplied sediment to Strawn deltas and fan deltas.

Fluvial, deltaic, and related facies recognized in the Brazos River valley by Cleaves (1975) are presented in table 2.

Strawn rock types and facies exposed in the Colorado River valley differ from Strawn deposits in the Brazos River valley in several respects (figs. 5, 6). There is considerably more sandstone, locally conglomeratic, in the lower part of the outcropping Strawn Group in the Colorado Valley. Very little coal, none of it economic, is found in the Colorado Valley. Only two mappable limestone beds (Capps Limestone and Ricker Station Limestone) crop out within the Colorado Valley.

Kier (1972) summarized inferred depositional environments of the Strawn Group in the Colorado River valley and concluded that much of the outcropping Strawn is of fluvial or fluvial-deltaic origin. Source area for the terrigenous clastic sediment was to the east in the Ouachita Mountains. Conglomerate, composed of subangular to rounded pebbles, cobbles, and boulders of eroded Marble Falls Limestone, is found at or near the base of the Strawn Group in San Saba County (Kier, 1972)

TABLE 2.—*Fluvial and deltaic facies in Strawn Group, Brazos River valley*

[Data from Cleaves (1975)]

Deltaic facies	
Destructional	Distinguished on basis of stratigraphic position below open-shelf mudstones and limestone beds. Lenticular siltstone and silty sandstone interbedded with dark-gray to black bituminous mudstone; commonly burrowed, local long-crested symmetrical ripples; plant debris common; 2–30 cm thick.
Delta plain	Bituminous mudstone and siltstone; numerous root traces and tree stumps. Includes thin splay sandstone with trough crossbeds and meandering stream channel.
Distributary channel fill.	Fine- to very fine grained sandstone; sharp erosional base, abrupt gradational upper contact, lower beds commonly contoured where underlying mud is thick, no well-developed fining-upwards sequence, although base may be coarser grained than top, large-scale trough crossbeds in base grading upwards to small-scale trough beds and climbing ripple cross stratification; clay galls common near base; base of channel may contain abundant plant debris.
Interdistributary bay	Variable, distinguished on basis of stratigraphic position between channel-mouth bar and overlying channel sandstones. Commonly unlaminated brown mudstone with abundant ironstone nodules and thin muddy detrital coal zones; faunal content may include worthinid gastropods, pectinid and nuculanid bivalves, chonetid and spiriferid brachiopods, and crinoids; 0.3–3 m thick.
Channel-mouth bar	Massive beds of well-sorted fine- to very fine grained sandstone; plane beds and low-angle large-scale trough crossbeds dominant, high-angle trough-fill crossbeds common, soft-sediment deformation, particularly lower beds, growth faults occur; macrofossils rare, small plant fragments common; 6 m thick.
Proximal delta front	Thin to massive beds of well-sorted, fine- to very fine grained sandstone; oscillation ripple cross-stratification, plane beds, small- to medium-scale trough-fill crossbeds dominant, growth faults occur; macrofossils rare, small plant fragments common; 3–20 m thick (total delta front).
Marginal delta front	Massive, blocky beds of sandy, coarse siltstone and muddy very fine grained sandstone; extensively bioturbated and burrowed. Marine reworked sands transported along strike from channel-mouth bars and proximal delta front.
Deltaic facies—Continued	
Distal delta front	Thin beds of coarse-grained siltstone and fine sandstone interbedded with mudstone; graded beds, flow rolls, long-crested oscillation ripple cross-stratification, and small-scale trough crossbedding dominant; fossils and bioturbation rare; 1–10 m thick. Deposited in part by turbidity currents.
Prodelta	Dark-gray, brown, or black mudstone; fossils rare except in distal parts; 1.5–61 m thick.
Fluvial facies	
Confined valley fill	Fining-upward sequence; basal part composed of pebbly conglomerate with large-scale trough crossbeds grading upward to medium- to small-scale trough crossbeds, tabular crossbeds, and parallel bedding; found in shallow superimposed valleys commonly cut into delta-plain facies.
Fine-grained meanderbelt.	Fining-upward sequence; basal 1 m contains large-scale trough crossbedded chert-pebble conglomerate; overlain by small-scale trough crossbeds; upper part contains parallel-bedded silty clay partings and thin oscillation-rippled very fine grained sandstone.
Interdeltaic embayment facies	
Sheet sandstone	Thin-bedded siltstone and fine- to medium-grained sandstone; long-crested symmetrical ripples dominate, small- to medium-scale, low-angle crossbeds in lenses; burrows very common; macrofossils rare; thickness as much as 6 m. Derived from marine reworking of adjacent deltaic sediments, deposited in strand-plain and shoreface environments.
Mudstone	Massive, dark-gray to brown mudstone; thin discontinuous lenses of burrowed sandstone and siltstone; iron oxide and septarian nodules present; spiriferid brachiopods.
Coal and bituminous claystone.	Thurber coal: 0.3–1 m thick, jarosite partings; kaolinitic underclay with lycopod stigmaria and charcoal fragments; burrowed, silty sandstone below underclay and above sandstone. Bituminous claystone: platy, fissile, lenticular; contain finely divided unlaminated reedy plant debris; may or may not contain root traces.
Bayhead deltas	Similar to distributary channel-fill deposits, but small, and cut into or overlying bay facies.

and probably accumulated as fan and fan-delta deposits shed from local fault blocks.

In the Colorado River valley, the base of the Strawn Group is in contact either with the Smithwick or the Marble Falls Formation. The Strawn-Smithwick contact is conformable and gradational. Regionally, the Strawn-Marble Falls contact also is conformable (Kier, 1972). Principal evidence of conformity is the gradation from intraclastic Marble Falls limestones into Strawn sandstones or limestone conglomerates, which in turn grade laterally and vertically into Strawn sandstones. Paleontological evidence summarized by Turner (1957) also indicates a conformable contact. Observed unconformable relationships (relief on the Marble Falls (Turner, 1970) and channeling (Freeman, 1962)) are local and resulted primarily from erosion of Strawn delta distributaries into subjacent Marble Falls limestones. The Strawn Group pinches out between the Marble Falls Formation and Canyon limestones along the north side of the Llano uplift. Pinchout is apparently the result of nondeposition and perhaps local erosion associated with post-Marble Falls and pre-Canyon faulting (Cheney, 1940; Freeman, 1962; Freeman and Wilde, 1964).

CANYON GROUP

The Canyon Group is a sequence of Late Pennsylvanian (Missouri) carbonate and terrigenous clastic rocks that crop out in the Colorado, Brazos, and Trinity River valleys and are within the subsurface of west Texas. Reduction in terrigenous clastic sediment supplied from the eroded Ouachita Mountains and deposition of widespread open-marine limestone on the Eastern shelf (Concho platform) distinguish the Canyon Group from the underlying Strawn and overlying Cisco Groups. Individual limestone units as much as 26 m thick cap a series of prominent, east-facing cuestas in the Canyon outcrop belt. Steep cuesta slopes and grassy valleys are underlain by deltaic and marine shales. Lenticular deltaic sandstone bodies within the shales crop out in hummocky, post oak-covered ridges. Schematic cross sections of the Canyon Group are shown in figure 7.

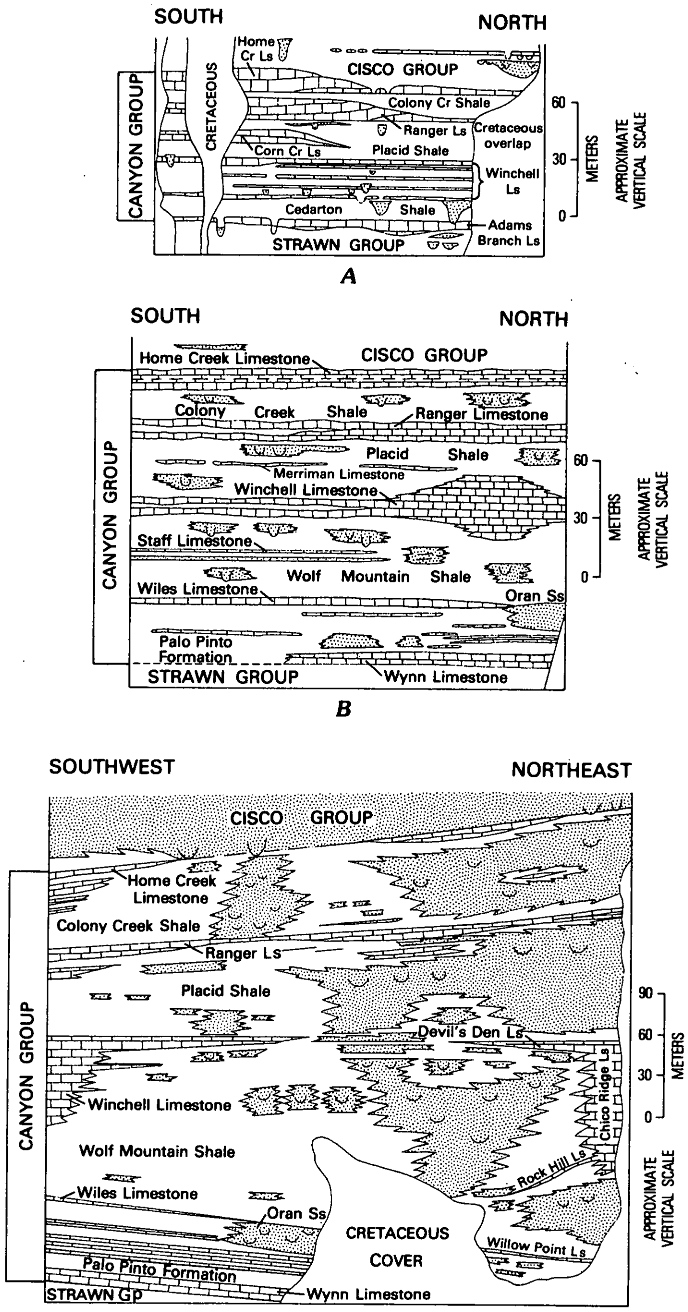

FIGURE 7.—Schematic cross sections of outcropping Canyon Group, Colorado, Brazos, and Trinity River valleys. *A*, Colorado River valley, McCulloch, Coleman, and Brown Counties (modified from Kier and others, 1976). *B*, Brazos River valley, Eastland, Stephens, and Palo Pinto Counties (modified from Brown and Goodson, 1972). *C*, Brazos and Trinity River valleys, Palo Pinto, Jack, and Wise Counties (modified from Erxleben, 1975).

The Canyon Group was named by Cummins (1891) for massive limestone and interbedded shale exposed near Canyon, a Texas and Pacific Railroad station 6.4 km west of Strawn in Palo Pinto County (fig. 2). In the Colorado River valley, Drake (1893) described and named many stratigraphic units within the Canyon Group. Plummer (1919) and Plummer and Moore (1921) named stratigraphic units in the Brazos River valley, and Scott and Armstrong (1932) named units in the Trinity River valley. Other early investigators include Dobbin

(1922), Reeves (1922), Plummer and Hornberger (1935), Bradish (1937), and Lee (1938).

Cheney (1940, 1947, 1949), who redefined the stratigraphic classification of Pennsylvanian and Permian strata of Texas, correlated Canyon rocks with the Missouri Series of the midcontinent. Lithofacies studies were carried out by Wermund (1966, 1969, 1975), Wermund and Jenkins (1964, 1969, 1970), Roepke (1970), and Erxleben (1973, 1975). Brown and Goodson (1972) and Kier and others (1976) mapped Canyon outcrops in the Brazos and Colorado River valleys, respectively. Other reports on the Canyon Group include: Abilene Geological Society (1954), North Texas Geological Society (1940, 1956, 1958), West Texas Geological Society (1951), Cheney and Eargle (1951), Jenkins (1952), Feray and Jenkins (1953), Shelton (1958), Eargle (1960), Terriere (1960), Laury (1962), Perkins (1964), Raish (1964), Bretsky (1966), Brooks and Bretsky (1966), Feray and Brooks (1966), Pollard (1970), and Heuer (1973). Wermund (1966) and Erxleben (1975) summarized previous investigations of the Canyon Group.

As presently defined (Brown and Goodson, 1972; Erxleben, 1973, 1975; Kier and others, 1976), the Canyon Group comprises seven formations (figs. 4, 7). In the Brazos River valley, the contact between the outcropping Canyon Group and the underlying Strawn Group is placed at the base of the Wynn Limestone Member, lowest limestone in the Palo Pinto Formation (Brown and Goodson, 1972). The Palo Pinto Formation (and Wynn Limestone Member) pinches out southward and is absent in the Colorado River valley outcrop area. In the Colorado River valley, the base of the Adams Branch Limestone defines the base of the Canyon Group (Kier and others, 1976). Although several other Strawn-Canyon contacts have been used in the past (see Shelton, 1958; Laury, 1962; and Roepke, 1970 for summaries), the base of the Palo Pinto Formation and the base of the Adams Branch Limestone best separate predominantly marine limestone and shale deposits (Canyon) and predominantly terrigenous clastic deposits (Strawn). The Adams Branch Limestone correlates with the Staff Limestone in the Brazos River valley (Cheney, 1929). Consequently, the base of the Canyon Group in the Colorado River valley is younger than the base of the Canyon in the Brazos River valley.

Drake (1893) placed the top of the Canyon Group at the top of the "*Campophyllum* bed" (Gunsight). Since the work of Plummer (1919) and Plummer and Moore (1921), however, the top of the Canyon has been recognized at the top of the Home Creek Limestone (figs. 4, 7), the uppermost thick limestone unit in the Pennsylvanian outcrop belt.

Shale and sandstone were deposited in terrigenous clastic delta, fan-delta, and shelf environments; limestone was deposited in carbonate shelf, bank, reef, and platform environments. Major influx of Canyon terrigenous clastic sediments into north-central Texas was concentrated in a high constructive delta system that crops out at the north end of the Canyon outcrop belt (Erxleben, 1973, 1975) in Jack and Wise Counties (figs. 2, 8). Thick shale and sandstone beds in the Wolf Mountain, Placid, and Colony Creek Formations (figs. 4, 7) were deposited in lobate and elongate deltas that prograded from Ouachita Mountains westward and northwestward across the Eastern shelf. Canyon delta facies resemble those in the Strawn Group (table 2). Additional terrigenous clastic sediments were deposited in north-central Texas by a fan-delta system (subsurface only) that prograded southward from the Arbuckle-Wichita Mountains (Erxleben, 1975). Minor amounts of terrigenous clastic sediments derived from the Ouachita Mountains were deposited in central Texas (Roepke, 1970).

Thick Canyon carbonate facies crop out near Possum Kingdom Reservoir in Palo Pinto County and Lake Bridgeport in Wise County (fig. 2) where the Winchell Limestone and its equivalent, the Devils Den Limestone, are composed of bank facies. Limestone banks are predominantly phylloid algal biomicrites including the genera *Eugonophyllum* and *Archaeolithophyllum* (Wermund, 1966, 1969, 1975), which acted as sediment traps for lime mud. A variety of other organisms lived in association with phylloid algae: encrusting algae; crinoids; fenestrate and encrusting bryozoans; fusulinids; echinoids; local rugose corals of the genera *Lophophyllidium* and *Caninia*, colonial syringoporid corals; sponge genus *Heliospongia;* brachiopods including the genera *Composita, Neospirifer, Echinoconchus,* and *Juresania;* gastropods of the genera *Bellerophon* and *Straparolus;* and pelecypods, including the genera *Aviculopinna, Myalina,* and *Culunana* (Wermund, 1969; Erxleben, 1975). Biosparites are uncommon but are found near the top of the bank deposits (Wermund, 1975). Both biohermal and biostromal banks are found.

Algal-bank facies are commonly found over paleobathymetric highs caused by differential compaction of subjacent deltaic sands or limestone banks and interlobe and interbank muds, respectively. The banks may also be associated with northeast

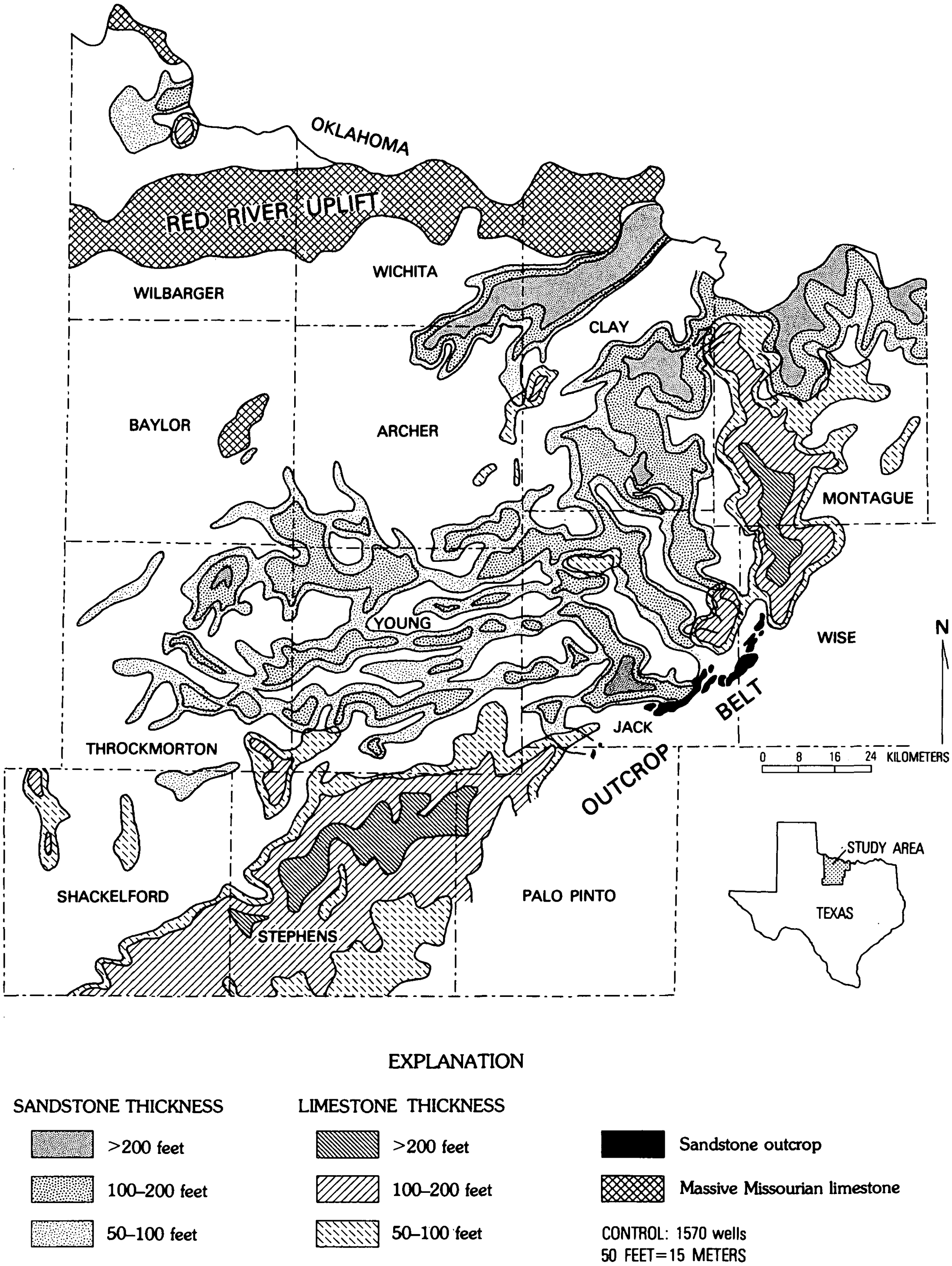

FIGURE 8.—Net thickness of sandstone and limestone of Winchell-Wolf Mountain Formations, north-central Texas. Area without pattern is shale or less than 50 feet of sandstone or limestone. Data on open-file, Texas Bureau of Economic Geology. Modified from Erxleben (1973).

structural trends (Brown, 1969c; Erxleben, 1975). Incipient paleorelief was probably a few centimeters (Wermund, 1966); maximum relief was probably 10 m (Wermund, 1975). Grain size, sorting, and crossbeds suggest limestone deposition above wave base. Analogy between Pennsylvanian phylloid algae and modern *Eudotia* algae suggests deposition in 1–3 m of water (Wermund, 1975).

Regionally extensive shelf-limestone deposits, such as the Palo Pinto, Adams Branch, Winchell, Ranger, and Home Creek Limestones (fig. 4), crop out in north-central Texas. After each major episode of delta progradation, shelf limestones onlapped (transgressed) subsiding delta lobes, providing widespread, relatively continuous marker beds that permit subdivision and correlation of Canyon strata in outcrop and subsurface. Shelf-limestone facies resemble bank facies but are irregularly bedded and contain thin marine shale beds. Individual shelf-limestone units are 1–15 m thick but may be thicker within interdeltaic embayments. An idealized Canyon depositional cycle (progradational deltaic sequence, destructional terrigenous clastic facies, and transgressive shelf limestone) is illustrated in figure 9; evolution of Canyon paleogeography is illustrated by figure 10. Platform and reef carbonates occur only in the subsurface along the Red River uplift and the eastern margin of the Midland basin, respectively (Erxleben, 1975; Wermund, 1975).

The Canyon Group is thickest in the Brazos River valley where outcrops are near the Canyon depocenter, the site of major terrigenous clastic deposition. As much as 545 m of Canyon rocks accumulated in Montague County (Erxleben, 1975; fig. 2); the Canyon Group thins southward to 273 m in Stephens County (fig. 2) and to only 120–135 m in the Colorado River valley. Although individual limestone units and interstratified terrigenous clastic facies thin southward in the Colorado River valley, limestone makes up a greater proportion of the section compared with the Canyon Group in the Brazos River valley. Near the Brady Mountains (fig. 2),

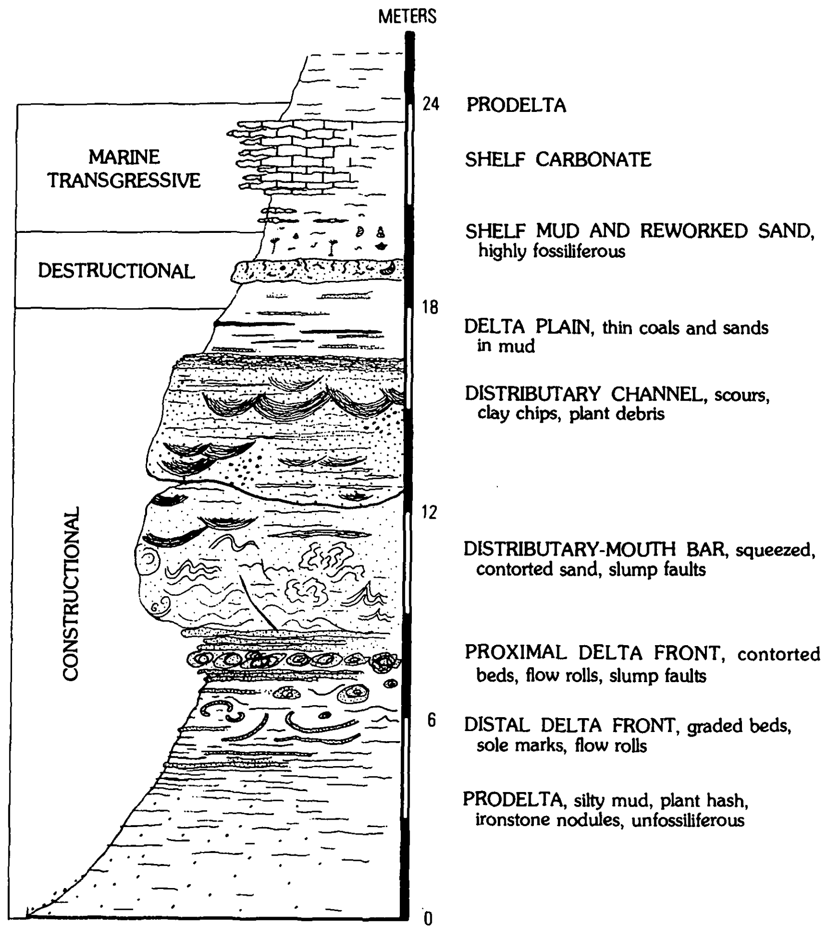

FIGURE 9.—Idealized delta sequence, Canyon Group, north-central Texas. From Erxleben (1975).

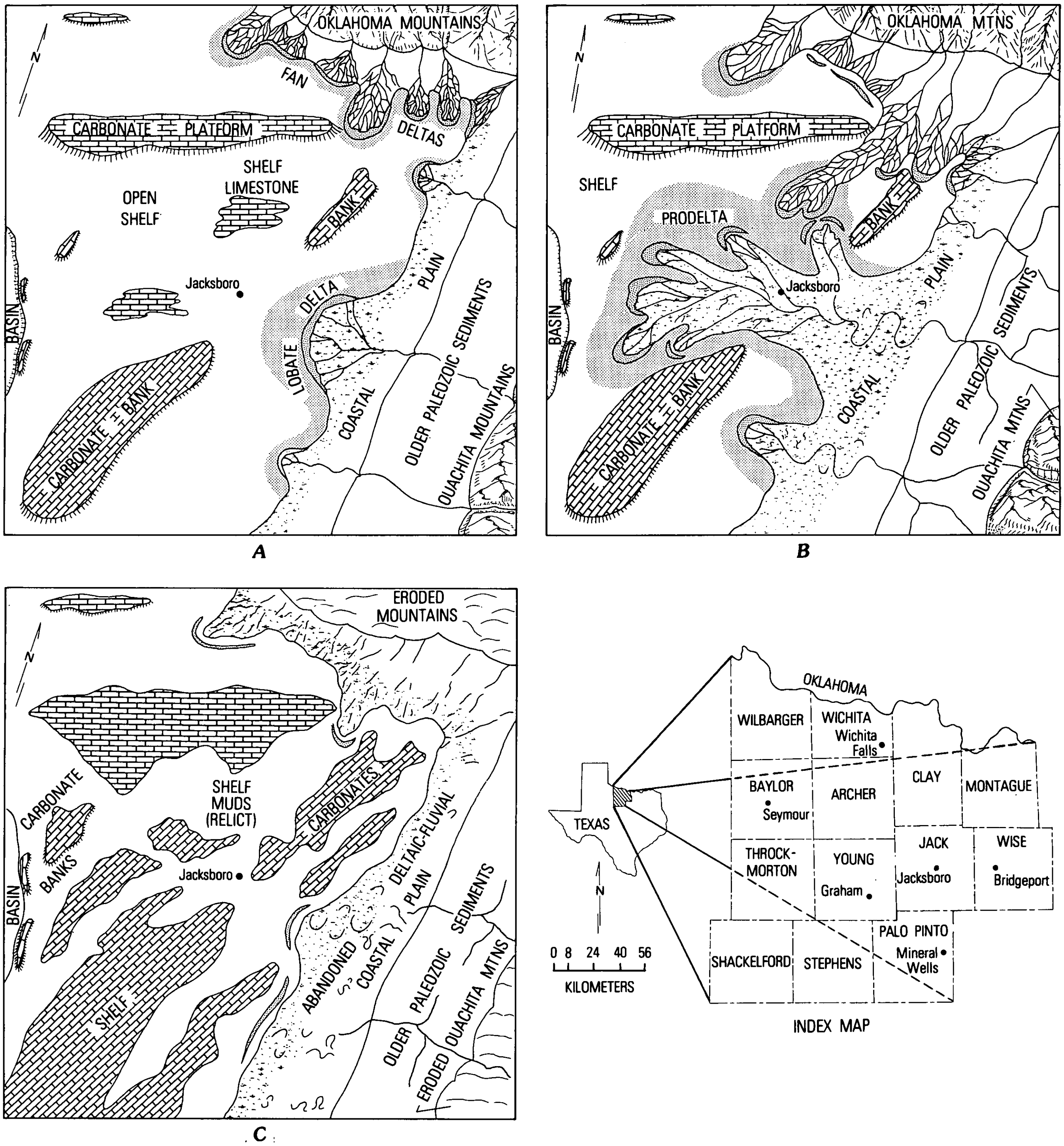

FIGURE 10.—Evolution of Canyon paleogeography in north-central Texas. *A*, Early progradation of delta systems. *B*, Maximum extent of delta development. *C*, Shelf transgression over abandoned deltaic facies. Based on three Canyon delta cycles. From Erxleben (1973).

shale and sandstone are essentially absent, and limestone makes up nearly all the Canyon Group.

CISCO GROUP

The Cisco Group, as originally defined by Cummins (1891), is a sequence of terrigenous clastic and carbonate facies that record rejuvenated uplift of the Ouachita foldbelt and Fort Worth basin. The increased sediment supply initiated extensive delta progradation across the Eastern shelf. Thick sandstone and conglomerate deposits distinguish the

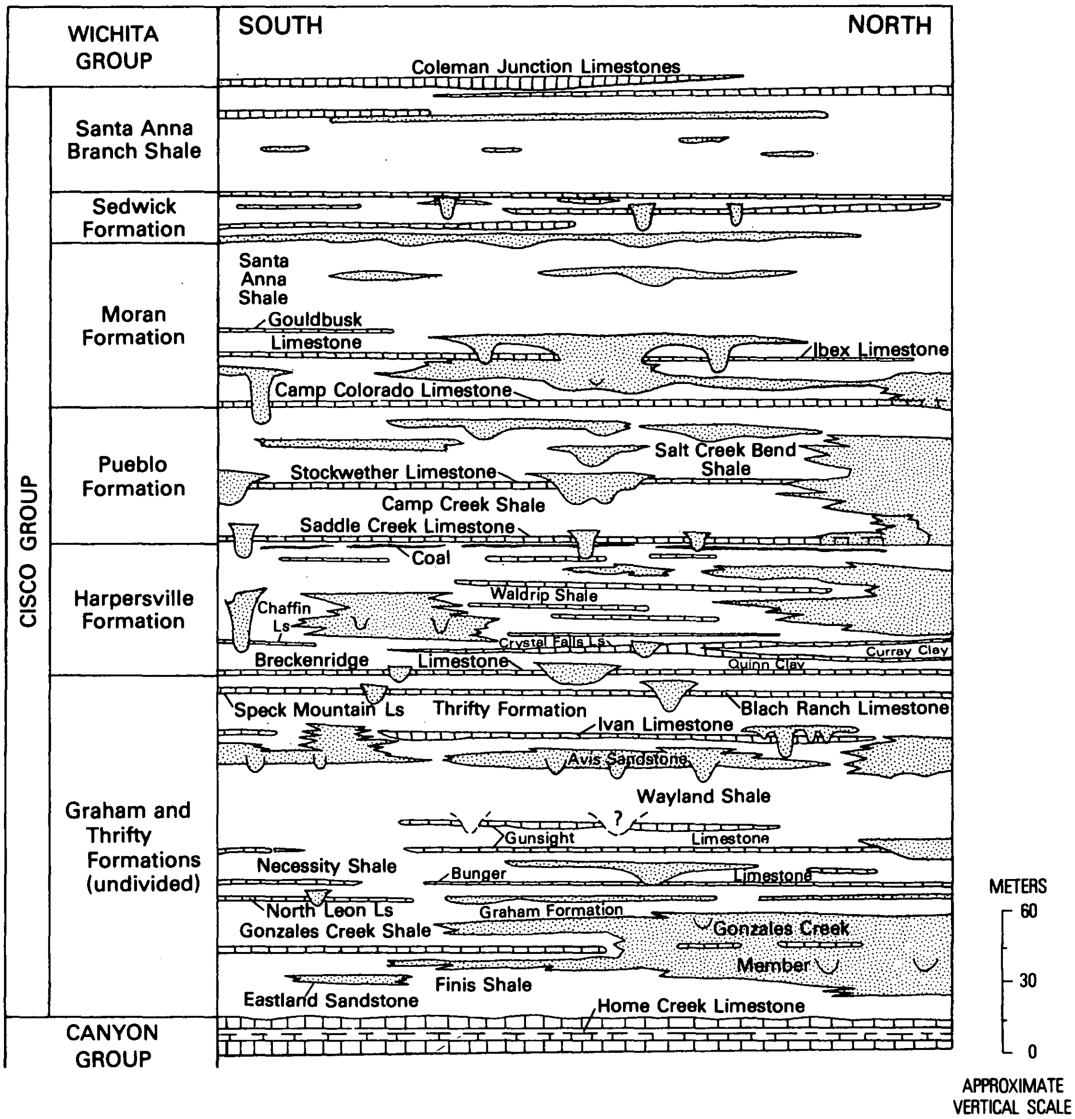

FIGURE 11.—Schematic cross section of outcropping Cisco Group in Brazos River valley. Modified from Brown and Goodson (1972).

Cisco Group from thick limestone and shale units within the underlying Canyon Group and overlying Wichita-Albany Group. Schematic cross sections of outcropping Cisco Group in the Brazos and Colorado River valleys are shown in figures 11 and 12. The Cisco Group as defined by the Bureau of Economic Geology (Brown and Goodson, 1972) is Virgil and, in part, Wolfcamp in age. Consequently, the Pennsylvanian-Permian boundary is within the group (fig. 4).

After Cummins' (1891) establishment of the Cisco, Drake (1893) divided it into "beds," many of which correspond to members in later classifications. Plummer (1919) proposed a preliminary classification of Pennsylvanian strata, including the Cisco, in which the tops of limestone beds were used as formational contacts. Lee (1938) first documented the complexity of Cisco facies and suggested some specific depositional conditions under which the group accumulated.

Cheney (1940) and Eargle (1960) redefined Cisco contacts to coincide with paleontologically inferred time boundaries, consequently elevating the Cisco Group (lithostratigraphic unit) to Cisco Series (time-stratigraphic unit). Brown (1959; 1960a, b; 1962; 1969a, b, c, d) and his students (McGowen, 1964; Seals, 1965; Waller, 1966; Galloway, 1970) studied the Cisco Group outcrop and subsurface in north-central Texas and devised a stratigraphic, sedimentologic, and structural framework for the

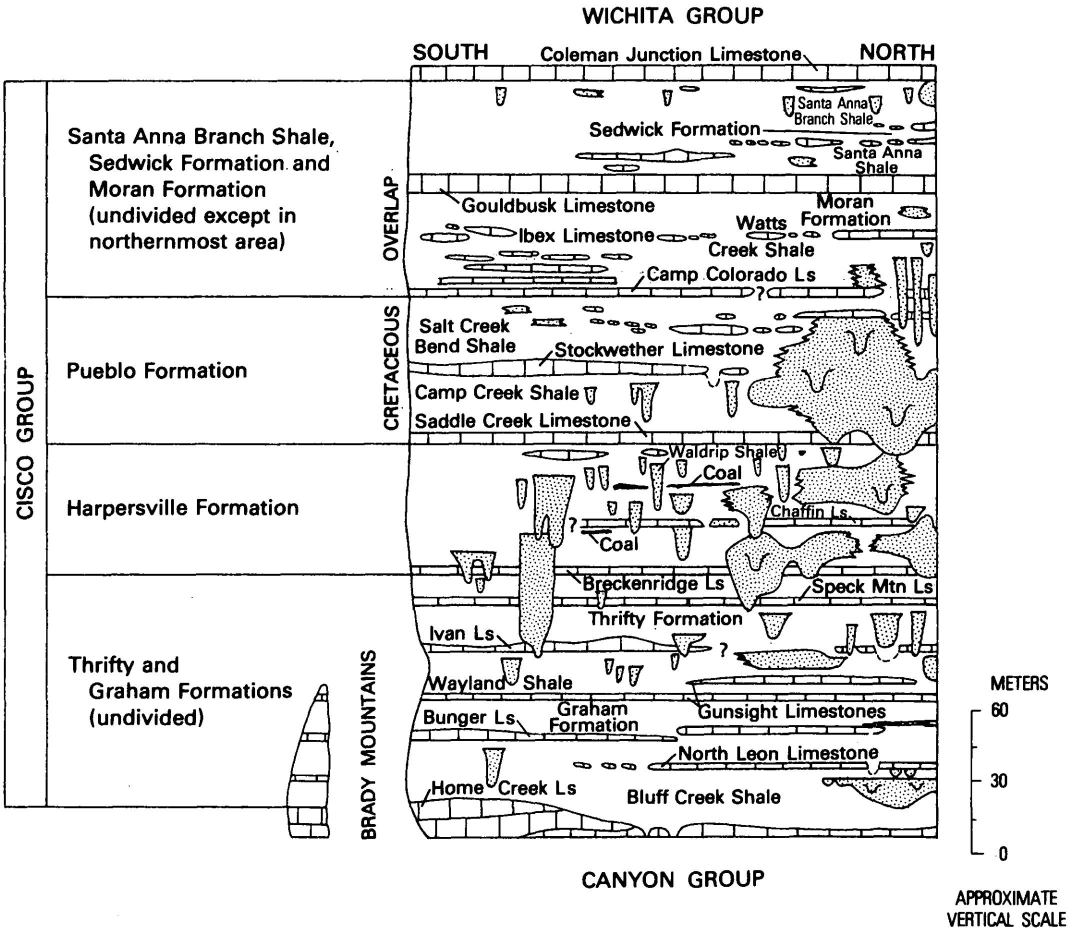

FIGURE 12.—Schematic cross section of outcropping Cisco Group in Colorado River valley. Modified from Kier and others (1976).

region. They recognized the original lithostratigraphic significance of the Cisco (Group) in surface and subsurface mapping programs aimed at lithogenetic interpretation of facies. Galloway and Brown (1972, 1973) presented an integrated depositional interpretation of middle Cisco rocks from outcrop, across the eastern shelf, into the Midland basin.

Brown and Goodson (1972) mapped outcrops of the Cisco Group throughout the Brazos River valley, and Kier and others (1976) extended this mapping into the Colorado River valley. Using the bases of widespread limestone beds as contacts, they divided the Cisco Group into six mappable formations in the Brazos River valley and four formations in the Colorado River valley (figs. 4, 11, 12). They placed the base of the Cisco Group at the top of the Home Creek Limestone (Canyon Group), and the top of the Cisco at the base of the Coleman Junction Limestone, following the original group definition of Plummer and Moore (1921).

The Pennsylvanian-Permian boundary, which is within the Cisco Group as defined by the Bureau of Economic Geology (Brown and Goodson, 1972), has been the subject of controversy for more than a century (San Angelo Geological Society, 1958). Moore (1940, fig. 4) illustrated various opinions about the placement of the time-stratigraphic boundary in Texas during the previous 60 years. The controversy is not yet resolved in North America (Wilde, 1975a, 1975b). The Pennsylvanian-Permian boundary in central and north-central Texas is difficult to determine because (1) no obvious regional physical or paleontological break provides a convenient boundary; (2) different boundaries have been selected using different faunal elements (fusulinids, brachiopods, or ammonites); (3) the Pennsylvanian-Permian boundary in the Glass Mountains of Texas, the reference area for the North American Permian, has not yet been settled (Cooper and Grant, 1972; Wilde, 1971, 1975a); and (4) few appropriate paleontological investigations have been carried out in the Cisco Group of central and north-central Texas, and none have been related to Cisco biofacies.

Nevertheless, fusulinids have been the basis for zonation in the Carboniferous and Permian throughout the world, and they have been used to recognize a Pennsylvanian-Permian boundary in North America for more than 30 years (Wilde, 1975b). The boundary in North America has been recognized by certain species of *Triticites* and the genus *Dunbarinella* in latest Pennsylvanian strata and by *Schwagerina, Pseudofusulina, Leptotriticites,* and other species of *Triticites* in earliest Permian beds (Wilde, 1975a). Roth (1931) found *Pseudofusulina* and Permian *Triticites* in Drake's (1893) Waldrip No. 2 limestone, which is within the shale (Waldrip Shale) between the Chaffin (Crystal Falls) and Saddle Creek Limestones (figs. 4, 11, 12). Cheney (1940) and Moore (1949) placed the Pennsylvanian-Permian boundary in a shale unit below Waldrip No. 2, suppressed the Harpersville Formation (Plummer and Moore, 1921), which included the boundary, and redefined and elevated the Cisco Group to Series and the Thrifty and Pueblo Formations to Groups. Their contact between the redefined Thrifty and Pueblo "Groups" was placed at the inferred time-stratigraphic boundary. Because this boundary is not mappable, Henbest (1958), Eargle (1960), Myers (1965), and others of the U.S. Geological Survey placed the Pennsylvanian-Permian boundary at the base of the Waldrip Shale (top of Crystal Falls Limestone, where present). Brown (1959) argued against suppression of the Harpersville as well as elevation of Cisco to series status. He preferred to apply the time-stratigraphic unit, the "Virgil Series" in Texas, rather than to redefine lithostratigraphy to "fit" inferred faunal-zone boundaries. Consequently, Brown and Goodson (1972) resurrected the Harpersville Formation as a regionally mappable formation and placed the highly subjective Pennsylvanian-Permian boundary in the upper one-third of the Harpersville Formation somewhere between the Chaffin (Crystal Falls) Limestone and the Saddle Creek Limestone.

The Cisco Group represents the last major episode of extensive fluvial-deltaic deposition on the Eastern shelf. After Cisco deposition, Middle and Late Permian carbonate, evaporite, and red-bed deposition dominated the shelf. Sandstone and shale are principal rock types in the Cisco Group; only thin transgressive shelf limestone beds crop out in central and north-central Texas. Downdip in the subsurface, thick shelf and shelf-edge limestone facies are common, but they commonly pinch out updip into nearshore clastic deposits. When each delta system was abandoned, it subsided, was reworked, and was transgressed by marine destructional facies (barrier and nearshore sands) and ultimately by marine shale and limestone (fig. 13). Cycles composed of regressive terrigenous clastic deposits and transgressive limestone deposits make up vertical sequences that show abrupt lateral facies changes (fig. 14). Brown (1973b) recognized 10 to 15 principal fluvial-deltaic progradational (regressive) episodes in the Cisco Group of north-central Texas.

In the Brazos River valley area during deposition of the Cisco Group, a westward or basinward shift in facies took place so that outcropping Cisco facies grade progressively upward (Brown, 1973b; fig. 14) from principally deltaic in the lower part (Virgil Series) to principally fluvial in the upper part (Wolfcamp Series). Fluvial facies recognized in outcrop are parts of braided, coarse-grained meanderbelt and fine-grained meanderbelt systems. Delta facies are components of thin high constructive elongate and lobate systems. Distribution of Cisco deltas was controlled by subjacent paleotopography induced by Canyon carbonate banks, differential sand/mud compaction, and differential rates of structural subsidence (Brown, 1969c). Interdeltaic-embayment facies include mudflats, chenierlike strandplains and coal, and brackish-bay mudstone and limestone (Galloway and Brown, 1972).

In the Colorado River valley, terrigenous clastic facies are restricted principally to the lower part of the Cisco Group (fig. 12) within the Graham, Thrifty, Harpersville, and Pueblo Formations. Scattered sand-filled fluvial channels are found in the upper part of the Cisco Group, and lenticular limestone beds are common.

Total thickness of the Cisco Group is 394 m in the Brazos River valley and 303 m in the Colorado River valley.

MARATHON UPLIFT

SYSTEMS BOUNDARIES

Neither the Devonian-Mississippian nor the Mississippian-Pennsylvanian boundary has been recognized in the Marathon uplift (fig. 4). The Mississippian-Pennsylvanian boundary clearly is within the Tesnus Formation; the Tesnus contains fossils as young as Early Pennsylvanian and as old as Late Mississippian (Ellison, 1962). The youngest (and only) fossils dated from the Caballos Novaculite, which apparently conformably underlies the Tesnus,

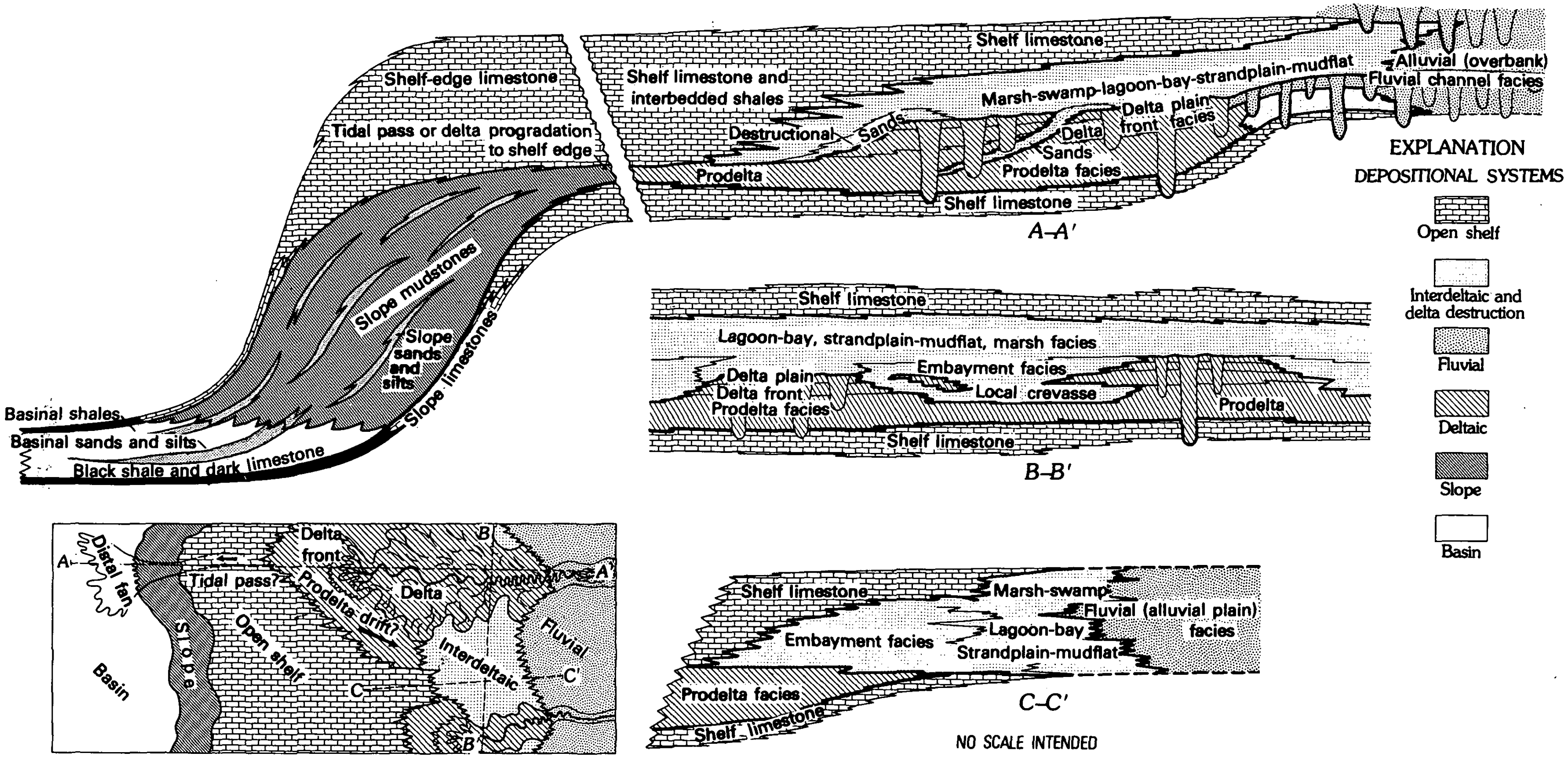

FIGURE 13.—Schematic cross section along Cisco paleoslope showing principal depositional systems. Based on 15 cross sections and 13 net-thickness maps. From Brown (1969d); reprinted with permission, Dallas Geological Society.

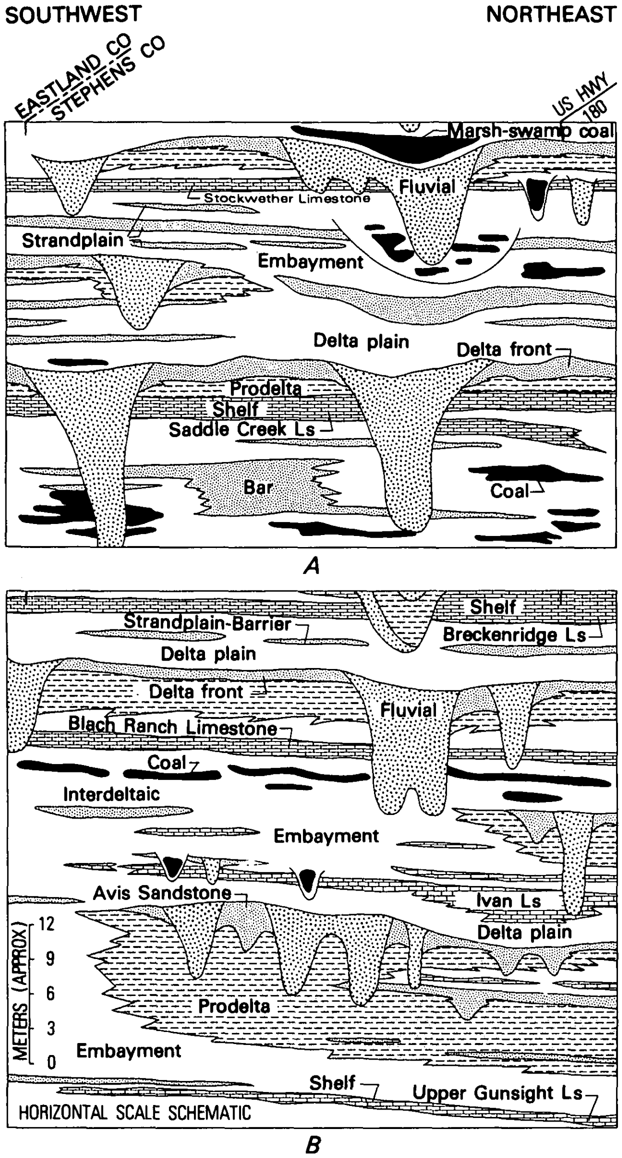

FIGURE 14.—Nature of outcropping cyclic facies, Cisco Group, Stephens County, Tex. *A*, Fluvially dominated marine-nonmarine cycles, upper part of Cisco Group. *B*, Delta-dominated marine-nonmarine cycles, lower part of Cisco Group. Based on detailed mapping and 350 measured sections. From Brown (1973).

are Late Devonian conodonts (Graves, 1952) some 50 m below the Caballos-Tesnus contact. Late Mississippian conodonts are found 5–10 m above the Caballos-Tesnus contact.

CABALLOS NOVACULITE

The Caballos Novaculite is a lens-shaped sequence 30–250 m thick containing equal amounts of massive novaculite (pure white chert) and thin-bedded green to gray radiolarite rhythmically interbedded with shale. The Devonian-Mississippian boundary probably is within the upper radiolarite. The upper boundary of the Caballos Novaculite is placed at the uppermost chert bed thicker than 5 cm. The formation was named by Udden, Baker, and Böse (1916).

TESNUS FORMATION

The Tesnus Formation (named Tesnus Shale by Baker and Bowman, 1917) is a wedge-shaped unit composed predominantly of repetitive beds of olive-drab to black shale and fine- to very fine grained sandstone. The formation is 2,000 m thick in the southeast outcrop area but thins westward to 100 m. A blanketlike shale about 100 m thick forms the base of the Tesnus in the east and composes nearly the entire formation in the west (fig. 15). An olistostrome 10 m thick containing exotic blocks of novaculite and older Paleozoic shelf carbonates crops out at the base of the Tesnus in the easternmost exposures (McBride, 1978). The olistostrome conformably overlies the Caballos Novaculite and is composed of siliceous shale in the lowermost part. Overlying beds of siltstone and sandstone gradually increase in thickness, grain size, and abundance and form a transition into the monotonous clastic Tesnus Formation. Sandstone layers range from 1 mm to 3 m thick, but most are 30–150 cm thick. Sandstone-shale ratios range from about 1:1 to 5:1, with an apparent, but undocumented, increase to the east.

Carbonaceous plant fragments and spores are locally abundant, and casts of wood as much as 30 cm long (chiefly calamarians, pteridosperms, and *Lepidodendron* (King, 1937, p. 61)) are found in shale. Sparse conodonts (Ellison, 1962), sponge spicules, radiolarians (Baker, 1963; Cotera, 1969), inarticulate brachiopods (J. Sprinkle, oral commun., 1976), and a single crustacean (Brooks, 1955) also have been found.

Most sandstones are olive-drab to light-brown, nonporous beds that contain 5–15 percent clay matrix and less than 10 percent quartz cement. Major framework components other than quartz are 5–10 percent each of feldspar and rock fragments. Several well-sorted, quartz-cemented quartzarenite beds, 10–15 m thick, are found only in the southeastern part of the basin.

Probably the most striking aspect of the Tesnus Formation in outcrop is the abundance of slump structures. In places, it is impossible to find 10 m of undisturbed (unslumped) section. Slump features include warped, folded, disrupted, and broken sandstone beds and contorted shale beds. Sandstone dikes of uneven thickness, but generally less than 5 cm

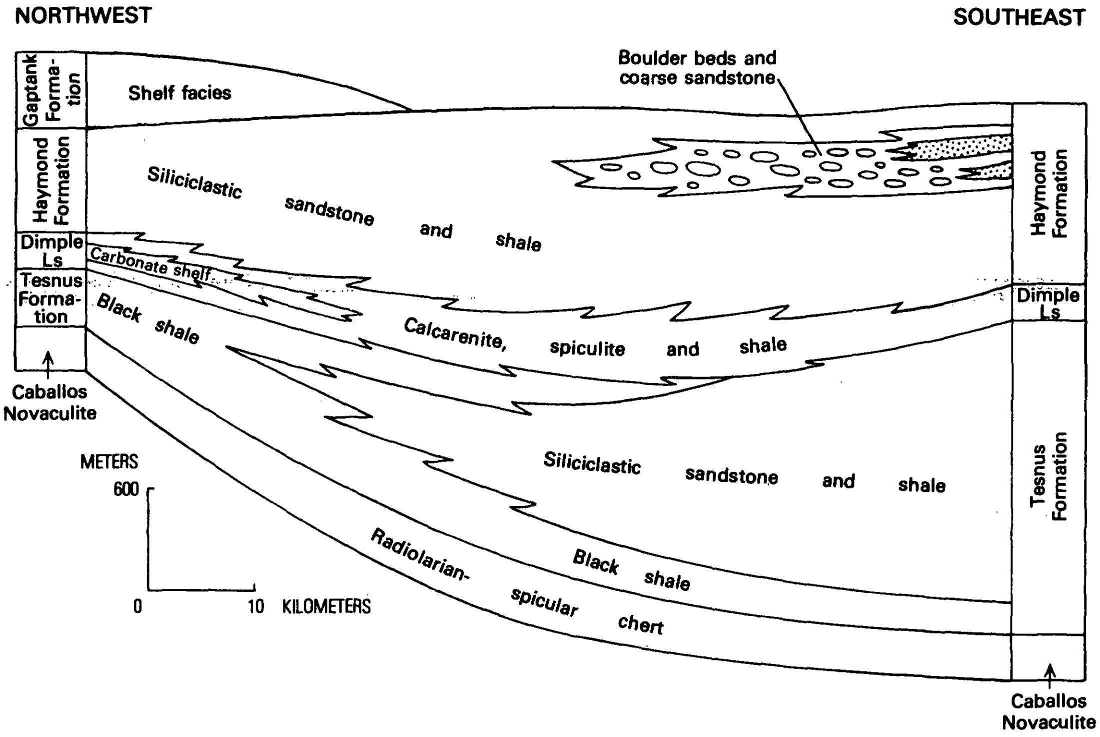

FIGURE 15.—Schematic cross section of flysch units approximately perpendicular to axis of Marathon geosyncline. From Thomson and McBride (1964).

thick, are associated features. Soft-sediment faults that transect sandstone soles are commonly regularly oriented and can be used to determine paleoslope direction (Thomson, 1973).

DIMPLE FORMATION

The Dimple Formation (named the Dimple Limestone by Udden and others, 1916, p. 46) has a gradational basal contact with the Tesnus. Over a 5-m interval, limestone beds and intercalated shale beds increase in thickness until they predominate, marking the basal Dimple contact. Maximum thickness of the formation is 300 m; the formation thins to 150 m at the margins of the Marathon uplift.

From northwest to southeast over a distance of 35–45 km, the Dimple Formation grades from rocks interpreted as "shelf" to slope and then to basin deposits (Thomson and Thomasson, 1969). The "shelf" facies are characterized by intercalated, crossbedded biosparite and oosparite (grainstones) in beds, most of which are 30–50 cm thick, lesser amounts of chert and limestone pebble conglomerate, and partings of shale that are 75 percent carbonate mud and 25 percent illite. Skeletal grains include crinoids, bryozoans, brachiopods, echinoids, fusulinids, algae, and conondonts. Trace amounts of quartz, chert, glauconite, phosphate, and heavy minerals are present. Grain size within calcarenite beds ranges from coarse to fine; sorting is good, and porosity may be as much as 10 percent.

Rocks of the basin facies of the Dimple Formation are about 70 percent calcarenite and cherty calcarenite, 20 percent mudstone, and 10 percent chert. Calcarenite and some mudstone beds are interpreted to be turbidites, whereas mudstone and spicular and radiolarian chert beds are pelagic deposits. Calcarenite beds are chiefly less than 30 cm thick, but some are 150 cm thick. Most calcarenite beds are uniform in thickness across an outcrop. Also, most beds (96 percent) are graded, but many contain other bedding types.

Rocks of the slope facies of the Dimple Formation are intermediate between shelf and basin facies. The most conspicuous feature of this facies is the presence of chert and limestone pebble conglomerate in beds 10–50 cm thick that fill erosional surfaces with as much as 30 cm relief and of calcarenite beds 60–100 cm thick that contain a single crossbed set. A 10-m-thick olistostrome containing exotic blocks crops out in easternmost exposures.

Most of the Dimple Formation is Atoka in age, but fusulinids suggest that the oldest part may be Morrow (Sanderson and King, 1964).

HAYMOND FORMATION

The Haymond Formation has a maximum preserved thickness of 1,400 m in outcrop. The basal contact with the Dimple Formation is gradational over a thickness of 5 m. The placement of the upper contact with the Gaptank Formation is disputed.

About three-fourths of the Haymond is composed of thin rhythmic beds of shale and fine-grained quartzose sandstones, which comprise more than 15,000 shale/sandstone couplets. The formation contains an olistostrome and associated lenticular coarse sandstone bodies that are exposed in strike sections along the eastern margin of the uplift. The olistostrome, or wildflysch unit, is 330 m thick and contains indigenous and exotic debris from pebbles to blocks 40 m long.

Sandstone beds have 3–10 percent clay matrix and less than 10 percent calcite or quartz cement. Cementation and compaction have destroyed all porosity. Framework composition averages 71 percent quartz, 15 percent feldspar, and 14 percent rock fragments. In the rhythmic sequences, sandstone beds are graded, laminated, and current rippled; many show convolute lamination. Current-formed sole marks are abundant.

The age of the Haymond is uncertain. Although an abundant Pennsylvanian fauna has been collected from exotic blocks in the boulder beds (King, 1937, p. 72), the only abundant indigenous fossils in the Haymond are plant remains and trace fossils. Fragmental remains of fusulinids, echinoids, brachiopods, and mollusks are found in the area of Dugout Mountain in beds transitional between Haymond and the overlying Gaptank Formation. In addition, fusulinids and echinoid fragments are found in two calcarenite beds in the lower part of the formation. An Atoka age (King, 1937, p. 72; Skinner and Wilde, 1954; Wilde, oral commun., 1962) was assigned to the fusulinids. These beds are turbidites of "Dimple aspect" that are several hundred feet above the base of the Haymond Formation.

DEPOSITIONAL ENVIRONMENTS OF THE CABALLOS NOVACULITE AND OVERLYING FLYSCH

Environments and processes of deposition of the flysch units and the underlying Caballos Novaculite remain controversial. Most early workers interpreted all formations to be of shallow-water origin (Waterschoot van der Gracht, 1931; King, 1937), whereas most later workers (King *in* Flawn and others, 1961, p. 84; Johnson, 1962; Cotera, 1969; McBride and Thomson, 1970; McBride, 1970) infer a deepwater origin. McBride (1970) interprets the Tesnus to be chiefly submarine fan deposits, and the Haymond to include submarine fan, basin-plain, and large-scale submarine slide deposits (wildflysch unit). Folk (1973) and Flores (1972, 1977), however, have argued for shallow-water origin of facies in the Caballos, Haymond, and Tesnus.

GAPTANK FORMATION

The Gaptank Formation conformably overlies the Haymond Formation. Although exposures are relatively small, stratigraphic subdivision of the Gaptank Formation has been the subject of controversy. Figure 16 charts the history of classification and shows various contacts within the Gaptank Formation exposed in the northern part of the Marathon uplift. A Chaetetes-bearing limestone 15 m thick is interbedded with limestone and sandstone of early Desmoinesian age. An overlying lenticular conglomerate member as much as 200 m thick is of early Missouri age (Ross, 1963, p. 17); five resedimented debris beds grading from limestone conglomerate to calcarenite characterize this unit. A superposed sandstone and shale member 200 m thick is poorly exposed. The uppermost ledge-forming limestone member about 150 m thick is composed of limestone bodies that rise southward in the section. Ross (1967) recognized shallow-shelf, shelf-edge, and deepwater facies in the uppermost limestone member. Conspicuous grainstones, packstones, and wackestones are present, and local bioherms are rich in dasycladacean algae.

According to Ross (1967) the outcrop belt of the Gaptank trends at a low angle to the northeast to north-northeast depositional strike of the formation. He inferred that the Gaptank is composed of several cyclic carbonate facies, which include (upward) slope, shelf-edge, and shelf deposits. Gaptank carbonate facies separate shallow-water clastic deposits to the southeast from deep-water clastic deposits to the northwest.

FRANKLIN AND HUECO MOUNTAINS

In the Hueco Mountains, the Helms Formation is composed of interbedded limestone and shale 150 m thick and is divided into a lower cherty member and chert-poor upper member. King and others (1945) reported that the Helms is sparingly fossiliferous but listed no fossils. The Magdalena Group, which rests unconformably on the Helms, was divided by King and others (1945) into three informal members. The lower member consists of 150 m of dark-gray, thick-bedded biomicrite; the

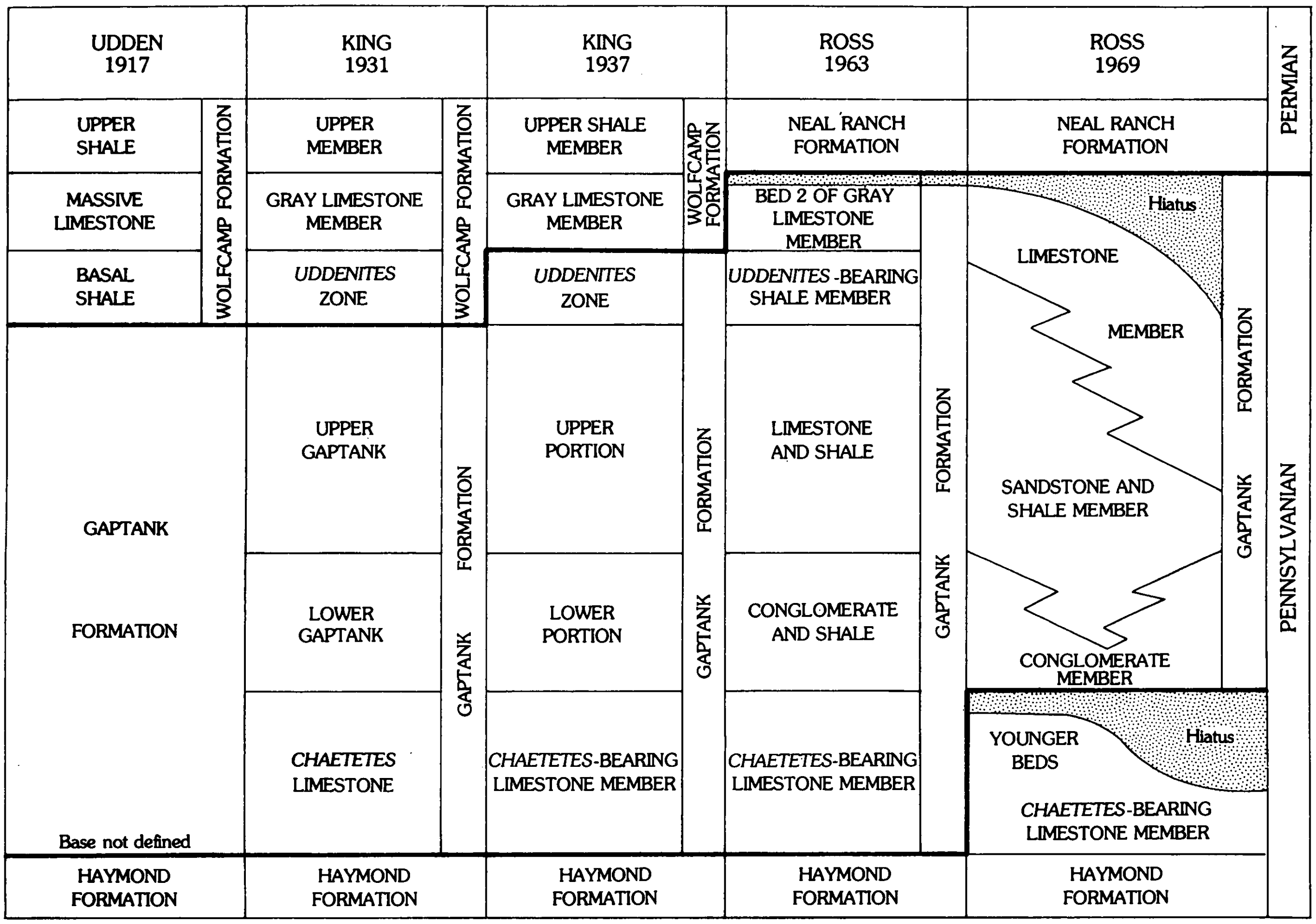

FIGURE 16.—Stratigraphic nomenclature of the Gaptank and adjacent formations. From Ross (1969).

middle member is composed of 100 m of marl, shale, and limestone; and the upper member is composed of 150 m of light-gray coral and algal biomicrite, biosparite, and conglomerate (Seewald, 1968).

In the Franklin Mountains, Mississippian strata include (upward) the Las Cruces Limestone, an even-bedded (beds 40–60 cm thick) gray micrite 20 m thick; the Rancheria Formation, a black argillaceous and cherty limestone 80 m thick; and the Helms Shale, a gray and green shale 50 m thick containing interbedded sandstone and limestone (Laudon and Bowsher, 1949). Unconformities bound the Las Cruces, Rancheria, and Helms Formations. The overlying Magdalena Group consists of more than 500 m of thin-bedded, dark-gray micrite and biomicrite containing some minor shale beds. The Berino Member is a shaly unit about 155 m thick, which separates the overlying Bishops Cap Member from the basal La Tuna Member. The uppermost, and unnamed, unit is poorly exposed and presumably is composed mostly of shale. The Magdalena-Helms contact is inferred to be unconformable (King and others, 1945). Environmental interpretations have not yet been reported for the Franklin Mountain sequence.

BOUNDING UNITS

CENTRAL AND NORTH-CENTRAL TEXAS

LOWER BOUNDARY

In outcrop, basal Mississippian rocks (fig. 4) generally unconformably overlie the lower part of the Ordovician Ellenburger Group, a thick group of carbonate strata that crop out around the Llano uplift (fig. 3) and extend throughout the subsurface of west Texas. The pre-Mississippian erosion surface appears to have little or no topographic relief. Regionally, however, the pre-Mississippian unconformity is angular (Barnes and Cloud, 1972).

The Ellenburger Group is composed predominantly of limestone and dolomite that accumulated in warm, shallow-marine environments "sedimentologically and ecologically" similar to the Bahamian Banks (Barnes and Cloud, 1972, p. 32). Numerous sinkholes formed in the Ellenburger during one or more periods of exposure, and post-Ellenburger strata commonly collapsed into or were originally deposited in them. Very locally, Upper Ordovician, Silurian, and Devonian limestone, shale and chert breccia (assigned to the Burnam, Starke, Piller Bluff, Stribling, Bear Spring, Zeson, and Houy Formations and several unnamed stratigraphic units) are preserved in collapse structures, cracks, and fissures in the top of the Ellenburger (Barnes and Cloud, 1972; Barnes and others, 1946, 1947, 1953, 1966; Cloud and others, 1957; Seddon, 1970). Aggregate thickness of the sinkhole deposits is less than 5 m.

Stratigraphic relationships among post-Ellenburger and pre-Mississippian deposits, as well as relationships with overlying Mississippian strata, are poorly understood. Stratigraphic and paleontological information about the remnant formations is limited. The Houy Formation apparently is, in part, Mississippian (Cloud and others, 1957; Seddon, 1970) and may record continuous deposition across the Devonian-Mississippian boundary. (See Seddon, 1970; Barnes and Cloud, 1972; and Kier, 1972.) Distribution of post-Ellenburger and pre-Mississippian rocks is so limited on the Llano uplift that the basal Mississippian boundary is essentially unconformable.

UPPER BOUNDARY

Permian rocks conformably overlie Pennsylvanian rocks in central and north-central Texas and record continuous sedimentation across the Pennsylvanian-Permian boundary. The boundary, as defined by fusilinids, is in the upper part of the Harpersville Formation (Plummer and Moore, 1921; Brown and Goodson, 1972) in the middle of the Cisco Group. No regionally significant hiatus has been recognized. Deltaic sedimentation, which dominated Cisco deposition, continued uninterrupted from Late Pennsylvanian into Early Permian time.

Lower Permian (Wolfcamp) deposition was marked by a gradual reduction in the influx of Cisco terrigenous clastic materials. During this time, shelf carbonates and marine muds onlapped the foundering Cisco deltas, and shelf environments dominated the Eastern shelf. Middle Permian carbonate and marine clastic shelf environments expanded, especially in central Texas, and were replaced upward by Middle and Upper Permian tidal-flat and sabkha environments (Smith, 1974).

WEST TEXAS

In the Marathon region of west Texas, the lower part of the Tesnus Formation, which is of Mississippian age, gradationally overlies the Caballos Novaculite, which is inferred to be principally Devonian in age. Late Devonian conodonts are found about 50 m below the top of the Caballos (Graves, 1952), and Late Mississippian conodonts are found 5–10 m above the base of the Tesnus (Ellison, 1962).

An angular unconformity separates the youngest Carboniferous unit, the Pennsylvanian Gaptank Formation, from Permian strata, either the Neal Ranch Formation or the Lenox Hills Formation of early and late Wolfcamp age, respectively (Ross, 1963). The Neal Ranch Formation is composed of shale, limestone, and siltstone cyclothems, and the Lenox Hills Formation consists of chert and limestone conglomerate, shale, and siltstone. Both formations were deposited in shallow water on the northern flank of the Marathon orogenic belt (Ross, 1963).

In the Franklin Mountains, limestone of Kinderhook age (Las Cruces Formation) rests unconformably on Upper Devonian shale (Percha Shale) according to Nelson (1940). In the Hueco Mountains, limestone beds of Meramec age (Helms Formation) rest on shale inferred on lithic character to be of Late Devonian age (King and others, 1945).

At the outcrop in the Franklin Mountains the Pennsylvanian-Permian boundary is inferred to be in the covered interval between limestones in the upper part of the Magdalena Group (Missouri age) and the overlying Hueco Formation (Wolfcamp age). In the Hueco Mountains, the boundary is placed within the Hueco Group (Seewald, 1968, p 47).

SUBSURFACE GEOLOGY

The subsurface geology of Carboniferous strata on the eastern shelf of north- and west-central Texas and within the Midland basin of west Texas is as well known as any sequence of rocks in the world (fig. 3). Tens of thousands of oil wells have penetrated fluvial and deltaic, shelf and shelf-edge, carbonate platform, and slope and basin facies throughout the region. Less intensively drilled but still reasonably understood are Carboniferous rocks in the Palo Duro and Dalhart basins of the Texas Panhandle and the Fort Worth, Kerr, and Val Verde Foreland basins that separate the Ouachita geosyncline on the east and southeast from the Texas cratonic region on the west and northwest.

Carboniferous strata in the Delaware basin and beneath thick Mesozoic and Cenozoic rocks along the Rio Grande in Trans-Pecos Texas are poorly understood because of structural complications and limited well information. Carboniferous rocks of the Marathon region are allochthonous, and little is known about their subsurface extent. Subsurface geology of Carboniferous strata within the basins and on the shelves of the west Texas oil province necessarily must be generalized in this report.

MISSISSIPPIAN ROCKS

Mississippian strata are of two principal types within the subsurface of Texas: (1) thick shale beds and interbedded sandstone within the Ouachita geosyncline and (2) limestone and marine shale facies on positive elements of the Texas craton (Texas peninsula, Adams, 1954; Concho arch, Cheney and Goss, 1952) and within shallow flanking basins, respectively. Within the frontal zones of the buried Ouachita foldbelt (geosyncline) are thick sequences of Mississippian fan-delta and slope shale and interbedded sandstone (Stanley, Jackfork, lower Johns Valley, and equivalent formations; Flawn and others, 1961). Upper Mississippian and Lower Pennsylvanian terrigenous clastic facies were deposited within the rapidly subsiding geosyncline and were highly folded and faulted during Early Pennsylvanian time in central and north-central Texas. Farther southwest, in the Val Verde basin (Young, 1960) and in the Marathon region, orogenic deformation took place as late as Early Permian.

Mississippian cratonic facies crop out in small areas of the Llano uplift and in some faulted mountains of southwest Texas and southeastern New Mexico. These shallow-marine facies grade into or intertongue with shale deposited in shallow flanking and intracratonic Mississippian basins (Barnett Formation). Pre-Pennsylvanian erosion resulted in the removal of Mississippian strata over significant areas of the Texas craton. Locally, as in Eastland, Stephens, Young, and Jack Counties (fig. 5), Mississippian biohermal reefs(?) formed along the eastern flank of the Concho arch or platform. Geosynclinal Mississippian deposits are restricted to the subsurface in central Texas, although the sequences crop out in southern Oklahoma and in the Marathon area of southwest Texas.

PENNSYLVANIAN ROCKS

In central and north-central Texas, Lower Pennsylvanian subsurface strata are inferred to have been deposited by two depositional systems (fig. 17). Fan-delta and slope clastic deposition (Atoka Group and part of Smithwick Formation) dominated in the Forth Worth foreland basin, supplied by the orogenically active Ouachita foldbelt. Contemporaneous deposition of limestone and marine shale (Marble Falls Formation and associated units) took place on the Concho platform and eastern platform margin. In west Texas and the Texas Panhandle, Lower Pennsylvanian platform carbonate facies grade laterally into arkosic fan-delta and slope facies near positive structural elements and grade basinally into deeper marine limestone and shale within the incipient Midland basin.

Middle Pennsylvanian subsurface strata (Strawn and Canyon Groups) are composed dominantly of proximal fluvial-delta facies from a diminishing sediment supply eroded from the Ouachita Mountains and from older Atoka foreland deposits. The deltas repeatedly prograded westward across the Concho platform into the deepening Midland basin (figs. 17, 18, 19). Extensive, contemporaneous, carbonate shelf-edge bank systems intertongue updip (sourceward) with nearshore clastic deposits and basinward with thin basinal shale and limestone beds. Middle Pennsylvanian carbonate shelves define the eastern and western flanks of the Midland and Palo Duro basins. As the Midland basin deepened during Middle Pennsylvanian time, the carbonate shelves were successively superposed or, in many places, they retreated landward in a recessive manner. Many isolated carbonate platforms formed at this time, such as the Scurry County "Horseshoe Atoll" and many en echelon banks formed along the edge of the relict Eastern shelf (fig. 3). Carbonate deposition continued on the Red River fault blocks.

Upper Pennsylvanian subsurface strata in north-central and west Texas reflect rejuvenation of the eastern source area and accelerated subsidence of the Midland basin. Extensive, cyclic fluvial-deltaic sandstone and shale sequences of the Cisco Group are deposited on the eastern shelf of the Midland basin (figs. 20, 21). The delta deposits grade westward into shelf and shelf-edge limestone facies, which in turn grade basinward into deep-water shale and sandstone. Most of the Midland basin contains only very thin Cisco shale and siliceous limestone beds ("starved basin" deposits (Adams and others, 1951)). Basinward progradation of delta, shelf, and slope deposits filled the Palo Duro and northern Midland basins by Early to Middle Permian (fig. 3). Along the margins of the Amarillo uplift and other similar structural elements in

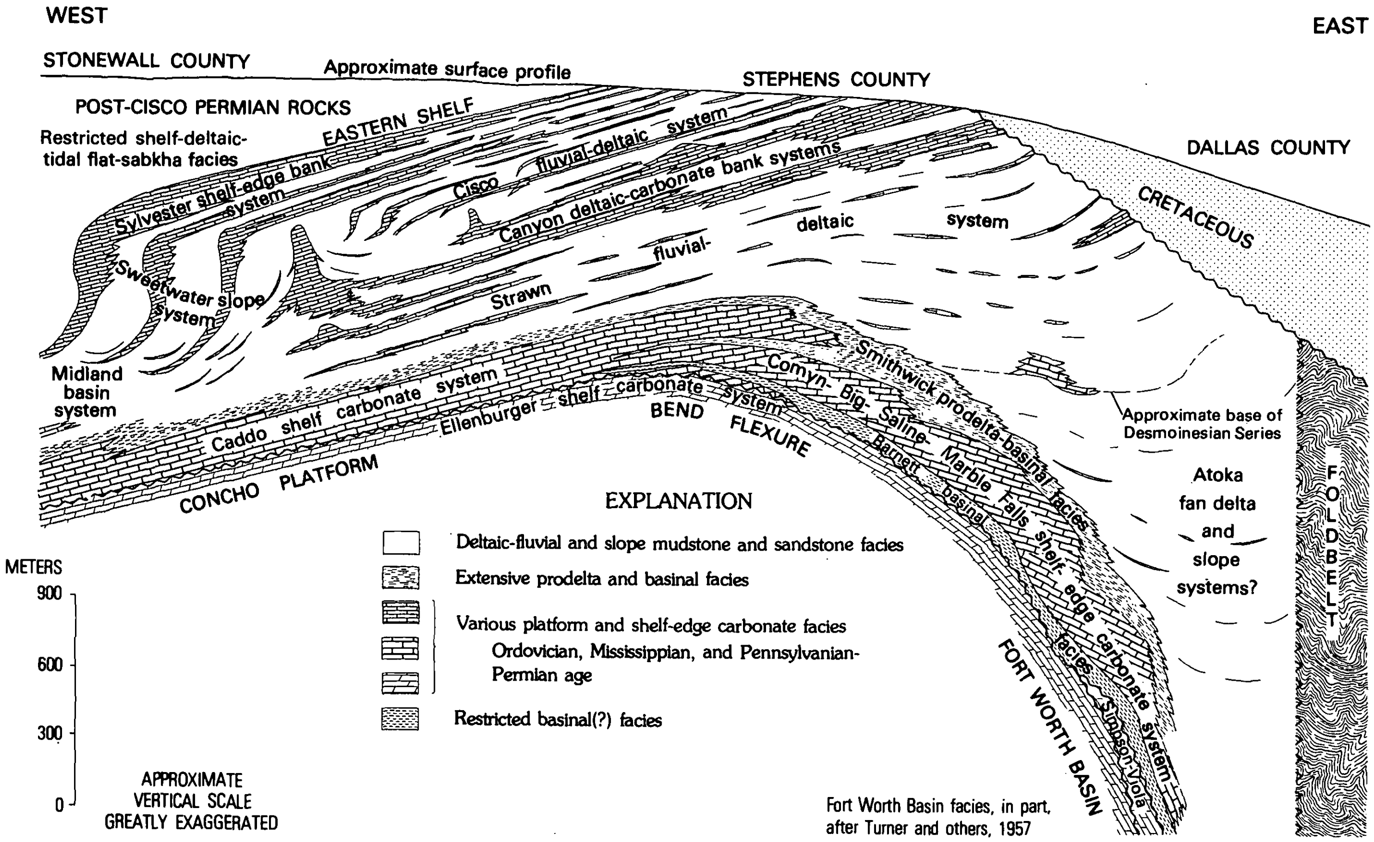

FIGURE 17.—Evolution of depositional systems, north-central Texas: Fort Worth basin, Concho platform, and Eastern shelf.

southern Oklahoma and eastern New Mexico are thick arkosic fan-delta deposits of Late Pennsylvanian age. Platform carbonates were deposited on the central basin platform.

Strata on the northwestern shelf of the Midland basin in eastern New Mexico and the northern shelf of the Anadarko basin in Oklahoma generally resemble sequences that compose the eastern shelf of the Midland basin. The southern flank of the Anadarko basin in Texas is filled by thick arkosic fan-delta deposits.

GEOLOGIC HISTORY

CARBONIFEROUS EVENTS

CENTRAL AND NORTH-CENTRAL TEXAS

The geologic history of central and north-central Texas is closely tied to the tectonic development of the Fort Worth (foreland) basin, the eastern shelf of the Midland basin, and the Red River uplift and southern Oklahoma mountains (fig. 3). Structural evolution of these basins and associated tectonic elements determined to a great extent the nature and distribution of the principal basin-filling depositional systems.

Beginning with Late Mississippian and Early Pennsylvanian structural activity in the Ouachita geosyncline, the Fort Worth foreland basin became well defined (figs. 3, 17). Platform and shelf-edge carbonate environments (Marble Falls, Big Saline, Comyn, and Caddo) contemporaneously dominated the Concho platform. Late Mississippian and Early Pennsylvanian shelf edges faced generally eastward toward the rapidly subsiding, but not necessarily deep, Fort Worth basin. Generally equivalent westward-prograding terrigenous clastic wedges (Atoka Group) entered the basin along a high-gradient paleoslope from the Ouachita foldbelt to the east. Thousands of feet of Atoka mudstone and sandstone of probable fan-delta and related slope origin graded westward and basinward into the thin, relatively starved basinal Smithwick facies. Basinal Smithwick shale and siltstone intertongued westward with shelf-edge carbonates of the Concho platform.

As Atoka clastic wedges built westward under gradually diminished but westward-shifting basinal subsidence, the shelf edges of the Concho platform carbonates retreated westward in a series of

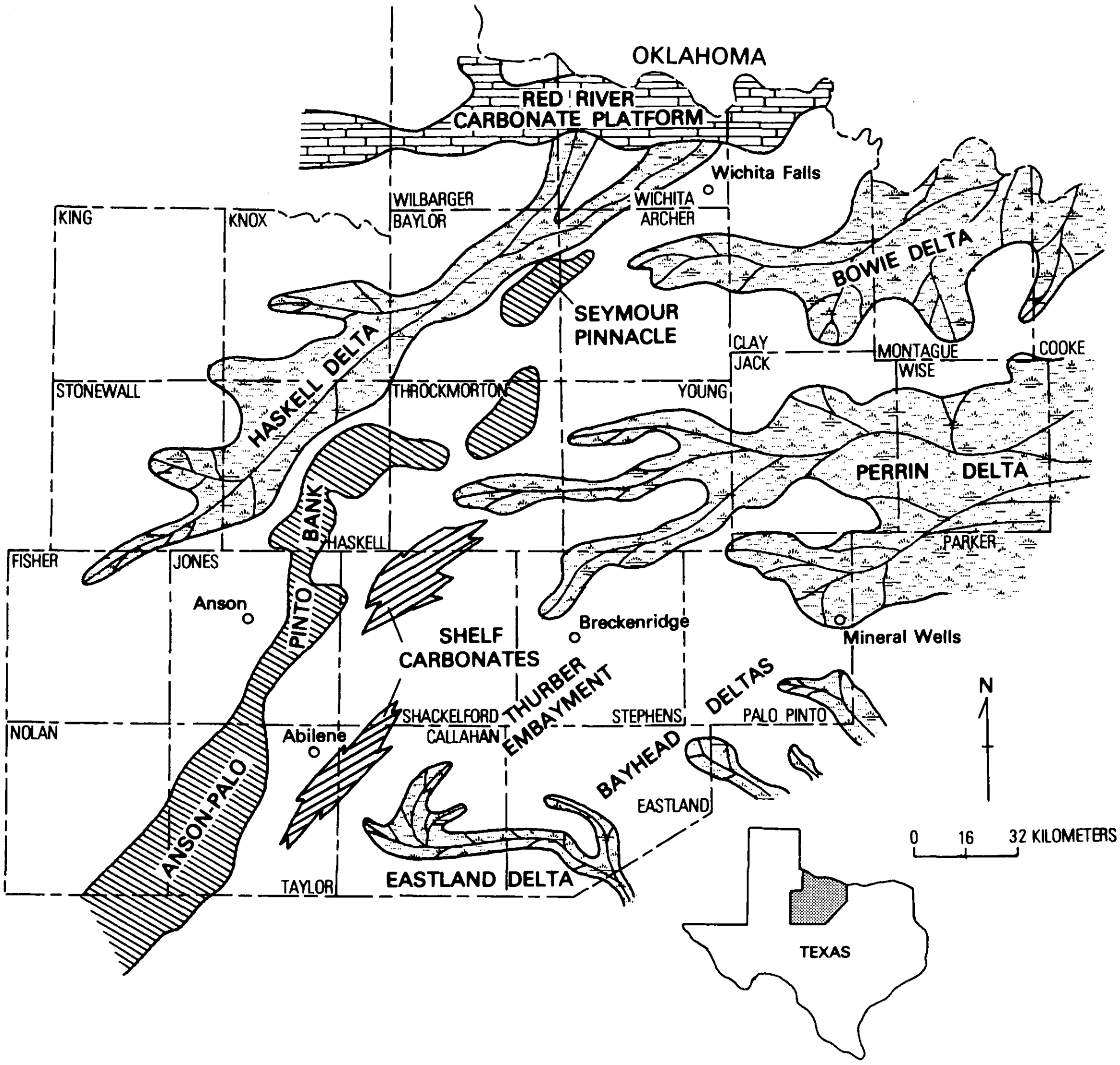

FIGURE 18.—Distribution of Strawn depositional systems, north-central Texas. From Cleaves (1975).

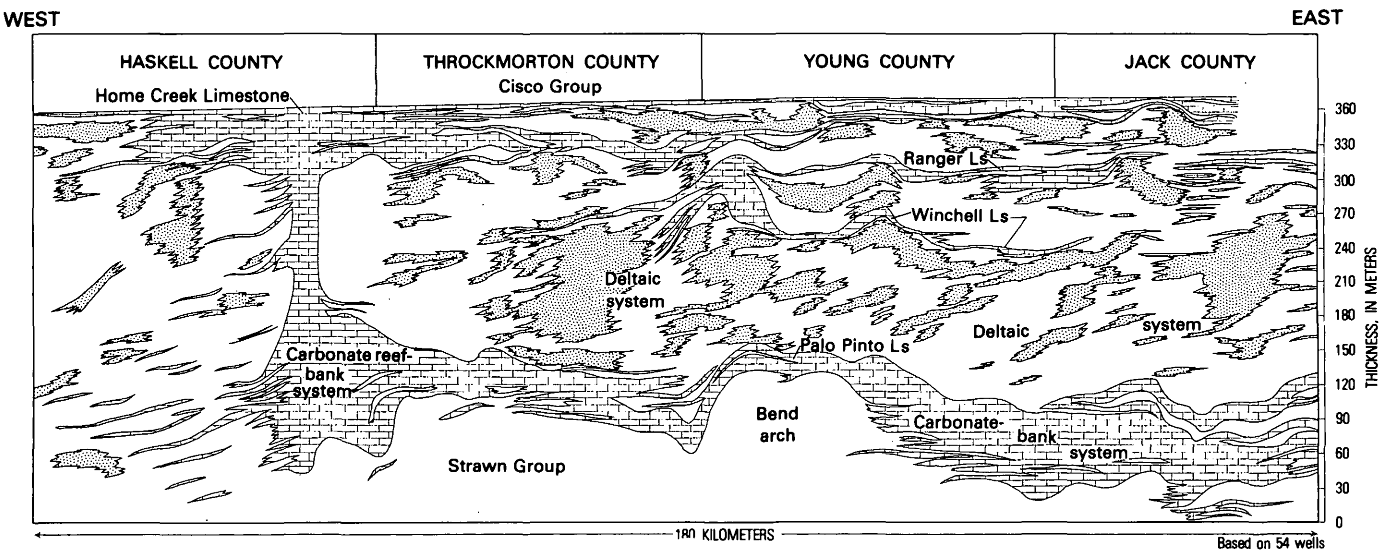

FIGURE 19.—Subsurface cross section of Canyon Group from Jack County to western Haskell County, Tex., showing deltaic and carbonate facies; based on 54 electric and sample logs. From Erxleben (1975).

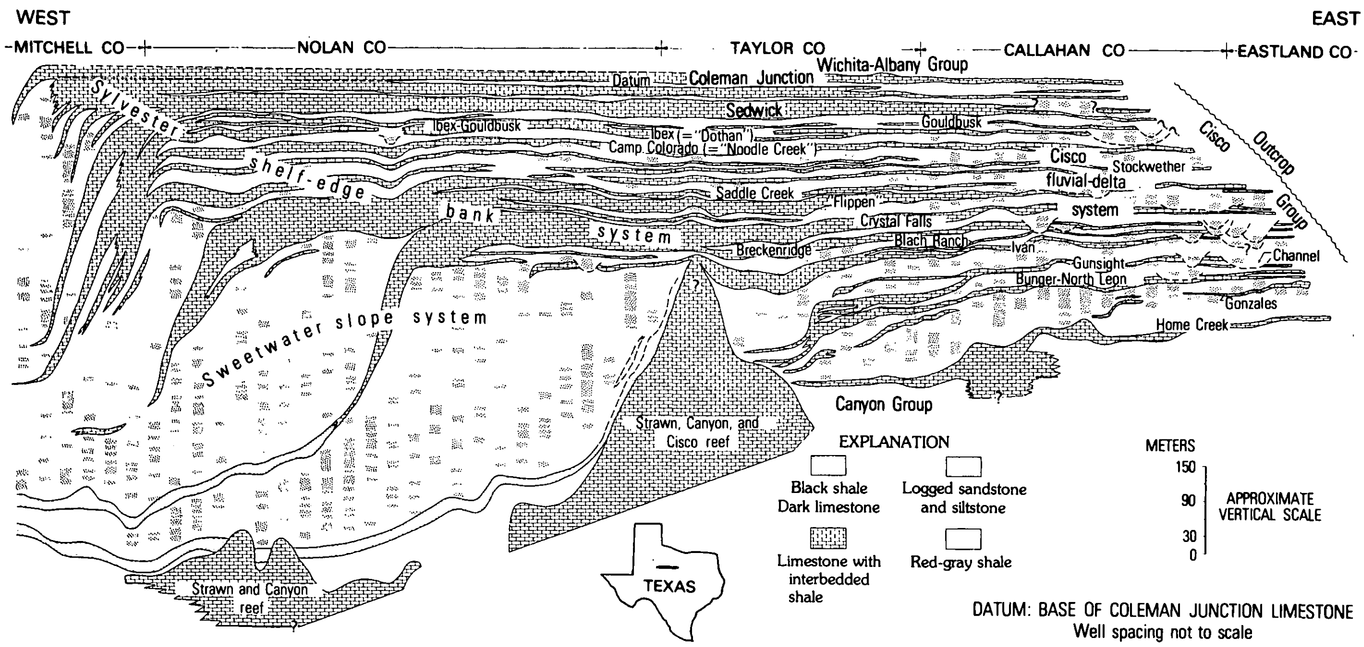

FIGURE 20.—Subsurface cross section showing facies of the Cisco Group, from Mitchell County to Eastland County, Tex. Based on 60 wells. From Brown (1969d). Data on open-file, Texas Bureau of Economic Geology.

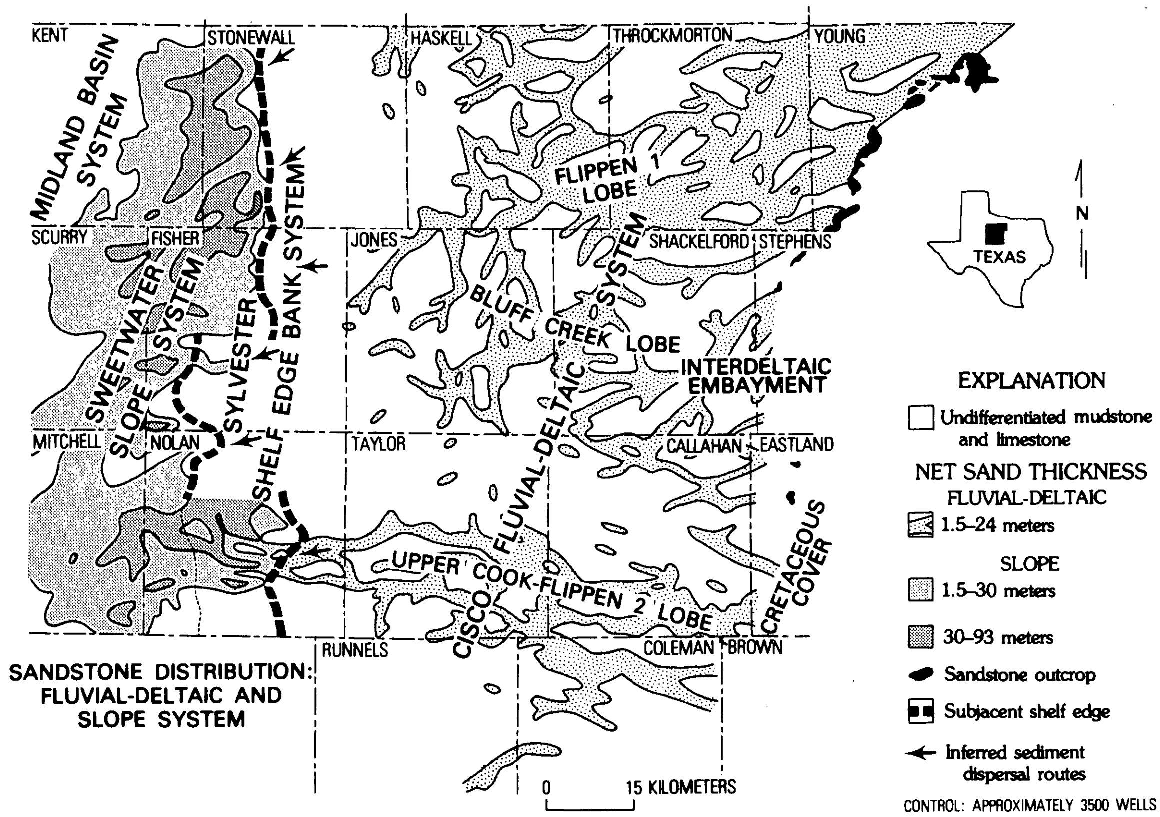

FIGURE 21.—Net-thickness map, Upper Cook-Flippen Sandstone, Cisco Group, north-central Texas. Based on 3,500 wells; data on open-file, Texas Bureau of Economic Geology. From Galloway and Brown (1972, 1973); reprinted with permission, American Association of Petroleum Geologists.

progressive "back steps," overlapped by the advancing Smithwick facies (fig. 17). Coarse clastic facies of Atoka fan deltas ("Bend Conglomerate") reached the western flank of the Fort Worth basin late in the waning stages of Atoka deposition. Facies within the Fort Worth basin, both terrigenous clastic and carbonate, indicate an uncommonly high degree of time transgression. Depositional environments shifted westward as a result of a shift in basin axis. Variable rates of subsidence and sediment supply also affected this shift.

EASTERN SHELF AND MIDLAND BASIN

Decreased subsidence in the Fort Worth basin and diminishing Atoka clastic input marked deceleration of Ouachita orogenic activity. During early Strawn (Desmoines Series) deposition, terrigenous clastic deposits gradually assumed a deltaic character, having lower paleogradients and a very shallow basin. During middle Strawn deposition, fluvial-deltaic systems overlapped the shelf-edge carbonate facies (Marble Falls, Caddo) and began several cycles of extensive progradation westward across the Concho platform (figs. 17, 18). Youngest Smithwick prodelta-basinal facies were deposited in the path of the delta systems. During late Strawn deposition, delta-fluvial sedimentation continued as the Concho platform underwent a gradual westward tilting and increased subsidence in response to accelerated subsiding of the Midland basin. Even though the stability of the Concho platform decreased near the end of Strawn deposition, this structurally positive element provided support for many upper Strawn and Canyon reefs and limestone banks (figs. 17, 19). Deposition of these carbonate systems initiated the high-relief shelf edges that later characterized the Eastern shelf during deposition of the Cisco Group.

Regional upwarping in the Ouachita foldbelt and the eastern flank of the Fort Worth basin was coincident with Midland basin subsidence and provided a significant supply of second-cycle sediments to Strawn deltas. The hinge or axis of rotation between the subsiding Midland basin and the gradually rising eastern Fort Worth basin defines the present Bend flexure or arch (fig. 3).

As the source areas were lowered by erosion and as paleogradients diminished, less terrigenous sediment reached the Pennsylvanian coastline. Extensive and long-lived carbonate-bank and reef systems began to form on the stable platforms provided by abandoned Strawn deltaic clastic materials (fig. 19). As the Midland basin continued to subside, many reefs or banks grew vertically to maintain necessary water depth. Trends of atoll-like limestone bodies that parallel the basin margin grew throughout much of the uppermost Pennsylvanian, but they are most common in the Canyon Group. Some of the carbonate banks, growing on structurally positive trends of the Eastern shelf, extended upslope to the present outcrop area where they intertongued with Canyon deltaic systems. Although Canyon deltas prograded extensively during three principal deltaic cycles, terrigenous clastic deposition was about equally balanced with limestone deposition.

Near the end of Canyon (Missouri Series) deposition, rejuvenation in the Ouachita foldbelt and eastern Fort Worth basin slightly increased paleogradients and significantly increased sediment supply, much of which was second-cycle detritus from earlier Atoka fan-delta facies and easternmost Strawn fluvial facies (fig. 17). As the supply of terrigenous clastic materials increased, extensive lower Cisco (Virgil Series) delta-fluvial systems began building westward across the eastern shelf, overlapping Canyon carbonate facies (fig. 20). Cisco delta systems prograded 10 to 15 times across the relatively stable eastern shelf of the Midland basin. Accelerated subsidence of the Midland basin provided as much as 455 m of relief between Cisco shelf edges and the floor of the Midland basin. Fluvial-delta systems built across the shelf and supplied sediment to thick, basinward-prograding slope-fan facies (figs. 17, 20). During deposition of the upper part of the Cisco Group (Permian, Wolfcamp Series), sediment supplied from the east again diminished, and thick, low-relief limestone shelf-edge banks became increasingly prominent.

After Cisco deposition, extensive carbonate-shelf and shelf-edge facies gradually restricted circulation on the landward parts of the eastern shelf. Minor deltaic and fluvial systems supplied fine-grained sediment that prograded the coastline locally and provided sediment to extensive tidal-flat systems. These tidal-flat systems accreted basinward and were overlapped by broad supratidal-flat (sabkha) evaporite systems.

RED RIVER ARCH AND OKLAHOMA MOUNTAINS

The complex history of the Wichita, Arbuckle, and Red River structural elements (Tomlinson and McBee, 1959) is recorded in thick clastic wedges extending southward and southwestward into the northernmost part of north-central Texas. These arkosic or "granite wash" deposits represent fluvial and fan-delta deposition along steep paleoslopes adjacent to fault blocks near the Red River in north Texas and southern Oklahoma. The fan deltas prograded basinward as a braided complex; prodelta facies and reworked fan-fringe deposits are of marine origin. Fan-delta deposition was contemporaneous with limestone deposition on adjacent, structurally positive blocks.

WEST TEXAS

Reconnaissance work in the Hueco Mountains indicates that the Magdalena Group records deposition on a shallow carbonate platform and adjacent slope. Other stratigraphic units have not been interpreted, and no environmental interpretation has yet been published for the Franklin Mountains.

The Marathon region was the site of slope and deep basinal sedimentation for most of the Paleozoic, including most of the Carboniferous. Radiolarian chert and shale of the upper part of the Caballos Novaculite probably were deposited in water depths greater than 1,000 m. Initial black mud of the Tesnus was followed by deposition of thin distal turbidites and shale that reflect progradation of a clastic wedge from east to west. Siliciclastic detritus was derived almost entirely from Llanoria (Africa and South America). Which deltaic environments are represented in addition to distal prodelta environments remains controversial.

Uplift of the western margin of the geosyncline initiated an episode of calcareous flysch deposition recorded in the Dimple Formation. Sediments derived from carbonate banks on the shelf and from uplifted older rocks were transported down a slope and into the basin as slides, debris flows, and turbidity currents. Pelagic calcilutite, shale, and spiculites are minor deposits. Unlike the Tesnus, the Dimple is thickest in the center of the geosyncline, rather than toward its source.

Renewed uplift of Llanoria brought a return to siliciclastic flysch deposition of the Haymond Formation. Initial deposits of the Haymond, like those of the Tesnus, consisted of black mud, which was followed by alternating turbidites and pelagites that make up the bulk of the formation. Turbidites increased slightly in thickness prior to the deposition of the slide, debris-flow slump, and turbidite beds of the chaotic boulder-bed member, but the wildflysch deposition apparently began abruptly because of the sharp base of the boulder-bed unit. Turbidite-pelagic deposition continued after formation of the wildflysch unit, but the sandstone beds locally are burrowed internally, suggesting that the geosyncline was getting shallower. At the northernmost outcrops, the Haymond passes upward into shelf and slope deposits of the Gaptank Formation.

POST-CARBONIFEROUS EVENTS

CENTRAL AND NORTH-CENTRAL TEXAS

Progressively restricted depositional environments existed along the margins of the Midland and Palo Duro basins (fig. 3) throughout the Permian. Tidal-flat, sabkha, and evaporite deposits predominated. Locally, fluvial and deltaic systems prograded across the marginal-marine deposits to supply fine-grained terrigenous clastic materials to the evaporite basin. As much as 2,350 m of Permian strata accumulated in the Midland basin; 1,200 m crops out on the eastern side of the basin.

Triassic rocks of the Dockum Group are inferred to overlie unconformably Permian rocks exposed in the vicinity of Palo Duro Canyon in the Texas High Plains, but to the south, the Triassic strata apparently are conformable with the Permian (J. H. McGowen, oral commun., 1977). The Dockum is inferred to be Late Triassic on the basis of vertebrate remains, but underlying strata assumed to be Permian are unfossiliferous, and no lithologic break has been observed in outcrop. Dockum deposition was principally lacustrine, centered within the relict Midland and Palo Duro basins. Locally, fluvial and deltaic systems prograded into the Triassic lakes. As much as 667 m of Triassic sediments was deposited; 330 m of Triassic rocks is exposed in outcrop.

After deposition of the Triassic Dockum Group, all of central, north-central, and west Texas underwent Jurassic and Early Cretaceous erosion that coincided with subsidence and seaward tilting of the Gulf Coast basin. Geomorphic patterns—shale valleys, sandstone hills, and limestone cuestas—and erosional relief that formed in central and north-central Texas at that time were similar to modern topography in the region. Headward eroding, eastward-flowing rivers supplied sediment derived from the exposed Triassic and upper Paleozoic rocks to delta and barrier-island systems along a transgressive Cretaceous shoreline.

Basal Cretaceous deposits in central and north-central Texas are quartzose and calcareous sand and gravel, principally of fluvial and shoreline origin. Continued westward onlap of Cretaceous marine environments out of the east Texas basin resulted in deposition of calcareous shale and limestone.

Eventually, all of central, north-central, and west Texas was submergent, and a major reef-lagoon system formed on the structurally high Llano uplift and Concho arch (fig. 3). Shallow-marine environments existed east and north of the reef tract.

Gradual uplift and erosion followed Cretaceous deposition in central and north-central Texas. No major Tertiary tectonic events took place in the region. Drainage similar to present-day patterns was established initially about Eocene time (O. T. Hayward, oral commun., 1977).

Major alluviation in central, north-central, and west Texas during Late Miocene and Pliocene time resulted in deposition of the Ogallala Formation. Aggrading fluvial systems originating in the Rocky Mountains deposited sand and gravel in a thick coalescing alluvial plain that ultimately stretched from Texas to South Dakota. In Texas, the alluvial deposits extended eastward over Triassic and upper Paleozoic rocks to about 144 km east of the present caprock escarpment (O. T. Hayward, oral commun., 1977). At the same time, valley-fill fluvial systems (Uvalde gravels) extended westward across lower Tertiary coastal deposits to upper Miocene and Pliocene Gulf shorelines.

During the Pleistocene, much of the distal part of the Ogallala alluvial plain was eroded, producing the Caprock escarpment. Headward erosion of the Pecos River system isolated the Ogallala alluvial plain in Texas from its source area in New Mexico (Thomas, 1972). Several episodes of Pleistocene alluviation followed, during which the Seymour Formation and other high gravels, derived predominantly from the Ogallala Formation and nearby Cretaceous exposures, were deposited by gulfward-flowing streams. Subsequent regional uplift, erosion, and stream piracy have isolated the Seymour and the other gravel deposits along drainage divides that are as much as 52 m higher than present-day drainage (Epps, 1973).

WEST TEXAS

In the Marathon region, the Paleozoic sequence underwent an episode of mountain building by compressional deformation during Late Pennsylvanian and early Wolfcamp (Ross, 1962) that caused the formation of folds, thrust faults, and strike-slip faults. A thick sequence of late Wolfcampian and Leonardian clastic and carbonate strata and Guadalupian and Ochoan carbonate strata, which were deposited over parts of the area, were subsequently tilted northeastward and eroded. Shelf carbonate rocks of Cretaceous age but of unknown thickness blanketed the area, and they, in turn, were arched and partly eroded during and following domal uplift in Tertiary time.

The Franklin Mountains and Hueco Mountains apparently underwent faulting during Laramide tectonism, but the main uplift of the Franklin Mountains began in late Cenozoic during formation of the Rio Grande rift. Some faults are still active. Quaternary alluvial fans flank both the Franklin and Hueco Mountains today.

ECONOMIC PRODUCTS

Carboniferous rocks in central, north-central, and west-central Texas have contributed significantly to the economic development of Texas and of the Nation. Oil and gas production has been predominant but industrial and ceramic clay, coal, and constructional limestone historically have been locally and periodically important. Uranium may occur within Carboniferous rocks in sufficient quantities to warrant further intensive exploration. Ground-water potential from Carboniferous rocks is poor, and ground water has not been significantly exploited. No resources of economic value have been recognized in the Marathon basin or in the Franklin and Hueco Mountains.

OIL AND GAS

Carboniferous rocks in Texas have undergone 60 years of intensive exploration for and production of petroleum. Earliest discoveries and principal production from Carboniferous rocks took place in north- and west-central Texas on the eastern shelf of the Midland basin, along the Red River, Muenster-Electra uplifts, and in the "Horseshoe Atoll" centered in Scurry County (fig. 3). Between 1916 and 1921, earliest discoveries were in sandstone beds of the Strawn Group and in the Marble Falls Limestone (and associated facies) in fields such as Ranger, Desdemona, and Breckenridge. Farther west, Cisco sandstone reservoirs were soon discovered (fig. 22).

Carboniferous sandstone reservoirs are generally shallow (less than 1,000 m) fluvial and deltaic facies located on subtle structural closures and in stratigraphic traps. Shelf-edge Marble Falls limestone facies provide slightly deeper targets directly beneath the shallow clastic reservoir rocks. Arkosic and feldspathic reservoirs (granite wash) of fan-delta origin are productive along the uplifted fault blocks that constitute the Red River, Wichita, and

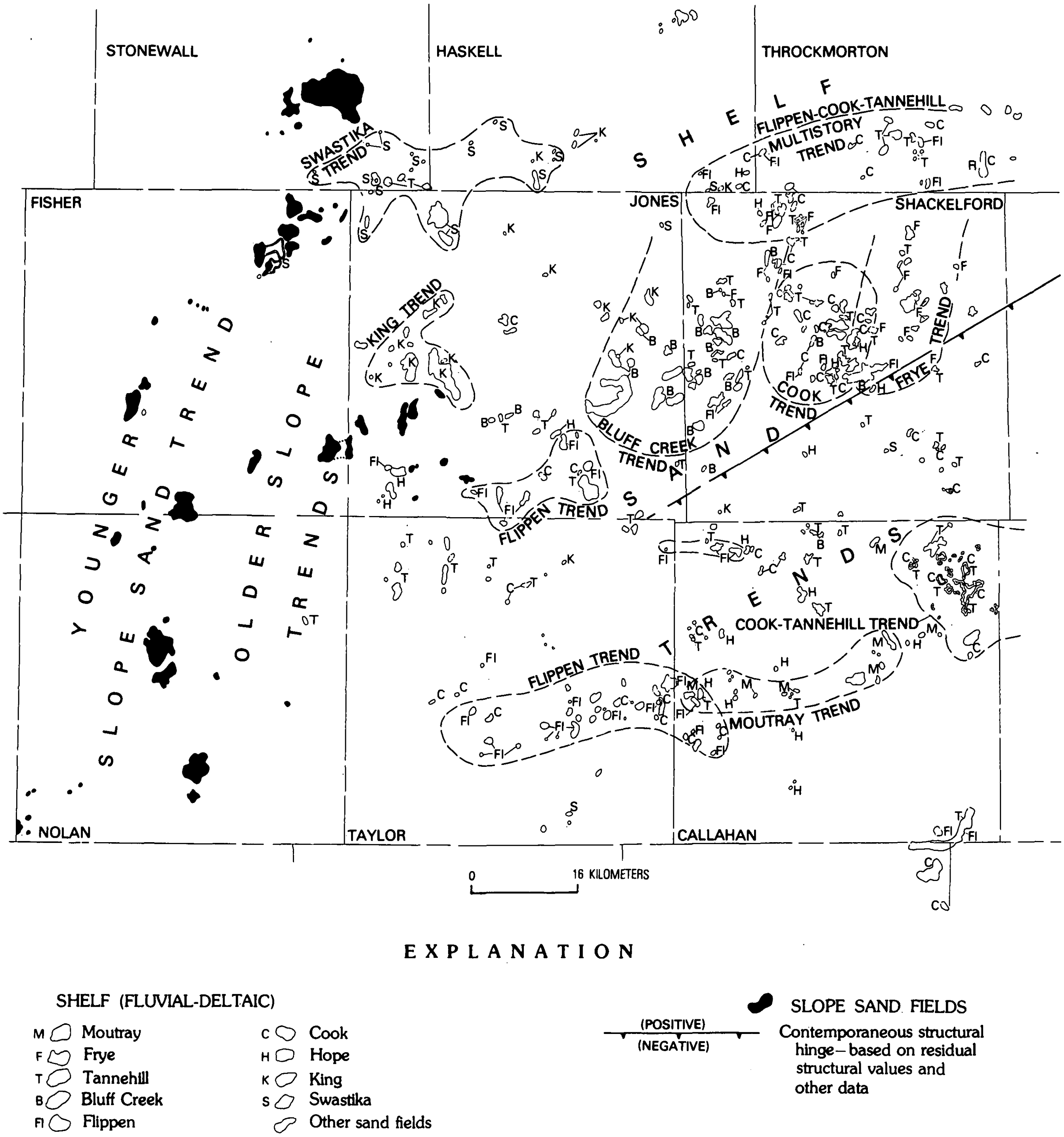

FIGURE 22.—Distribution of oil production from Cisco Group on central part of the eastern shelf, north-central Texas. Based on Abilene Geological Society data, by L. F. Brown, Jr., and W. E. Galloway. Modified from Abilene Geological Society (1960).

Amarillo uplifts (fig. 3). Most shallow structures were located by surface mapping. Shallow structures supported more than 40 years of intensive wildcatting, and today the oil province is again the site of active exploration for marginal oil and gas left behind in the small lenticular reservoirs.

After World War II, the single greatest discovery in Texas was the Horseshoe Atoll, an extensive complex carbonate system west of the Eastern shelf within the Midland basin (fig. 3). This discovery of Strawn, Canyon, and Cisco limestone reservoirs triggered extensive exploration of Pennsylvanian and Permian shelf edges during the 1950's. Many similar Strawn and Canyon limestone bank (or reef?) reservoirs were discovered trending approximately northeast along the eastern margin of the Midland basin. Some sandstone reservoirs of deep-water, submarine-fan origin were also discovered within the Midland basin adjacent to the edge of the relict Upper Pennsylvanian eastern shelf.

Principally gas has been discovered in the thick Lower to Middle Pennsylvanian fan-delta and deltaic sandstones and conglomerates in the Fort Worth and Sherman basins (fig. 3). Intensive exploration is currently underway in the Kerr and Val Verde basins (fig. 3), which are southwestern extensions of the foreland basin system separating the Texas craton and the Ouachita foldbelt.

Production from Pennsylvanian carbonate rocks takes place on the central basin platform. Deltaic sandstone beds of Early Pennsylvanian age were discovered recently on the northwest shelf in nearby eastern New Mexico. Localized oil fields produce from small isolated Mississippian reefs(?) found along the eastern flank of the broad Concho platform, the precursor of the later Eastern shelf.

COAL

In the last two decades of the 19th century, coal production from cyclic Pennsylvanian fluvial and deltaic facies in north-central Texas was a regional industry. Coal was used principally within the region for heating, although much of the Strawn coal was used until about 1920 as boiler fuel on locomotives of the Texas and Pacific Railroad. Mining in the region terminated in 1943.

Pennsylvanian coal in north-central Texas (fig. 23) is principally in the Mingus Formation, Strawn Group, and in the Harpersville Formation, Cisco Group (Mapel, 1967; Evans, 1974). Coal is found also in the Canyon Group, but it is very limited in distribution. Atoka coal deposits are only in the subsurface near the Red River arch.

The Thurber coal within the Strawn Group was the most economic deposit in the region, although several other coals were mined. The Thurber coal crops out beneath thick Brazos River Sandstone for about 19–24 km in southwestern Palo Pinto and Erath Counties (fig. 2). Shallow mining was concentrated on dip slopes near Thurber and Mingus along the Erath-Palo Pinto County line. The coal is about 1 m thick and was deposited within marshes and swamps bounding the Thurber embayment (Cleaves, 1975). Strawn coals rank as high-volatile bituminous with 2–8 percent moisture, 10–25 percent ash, 1.5–4 percent sulfur, and Btu (dry basis) from 10,390 to 13,755 per pound.

Several coal beds in the Harpersville Formation, Cisco Group, are found near the Pennsylvanian-Permian boundary (on the basis of fusulinids). Several lenticular coal beds as much as 1 m thick were deposited along shorelines bounding large inter-deltaic embayments (Brown, 1973b). Harpersville coal production was concentrated near Newcastle in Young County, near Crystal Falls in Stephens County, near Cisco in Eastland County, and in McCulloch and Coleman Counties (fig. 2). Various local names have been applied; the Newcastle coal was the most productive in the Cisco Group. Cisco coals are variable and have a moisture content of 8–18 percent, ash content about 15 percent, sulfur from 1.1 to 8.9 percent, and heating value (dry basis) from 5,669 to 7,054 Kcal per kilogram.

Limited possibilities exist for future stripping and for possible in situ gasification of Texas subbituminous coals. Metal byproducts and special uses in cement offer other possibilities for use of Pennsylvanian coal resources.

CLAY PRODUCTS

Common industrial clays mined from Middle and Upper Pennsylvanian strata are used extensively in the brick, tile, clay pipe, and expanded aggregate industry in north-central Texas (Brown, 1958). The industry is concentrated west of the Dallas-Fort Worth metroplex near Mineral Wells, Palo Pinto County. Five plants produce industrial clay products from prodelta facies that are associated with several Strawn delta systems. Principally illitic in composition, the clays are not of ceramic quality.

Locally near Cisco, Eastland County, and Cross Cut, Brown County (fig. 2), Cisco deposits such as the Quinn Clay (Plummer and Bradley, 1949) contain sufficient kaolinite to be used for cast items and glazed tile. Large volumes of expanded aggregate are produced from Canyon and Cisco prodelta-marine clays near Ranger and Eastland in Eastland County, respectively (fig. 2).

CONSTRUCTIONAL LIMESTONE

Limestone deposits of the Canyon Group are extensively quarried for use as aggregate in concrete and as base material for highways and airport runways. The largest operation is at Bridgeport in Wise

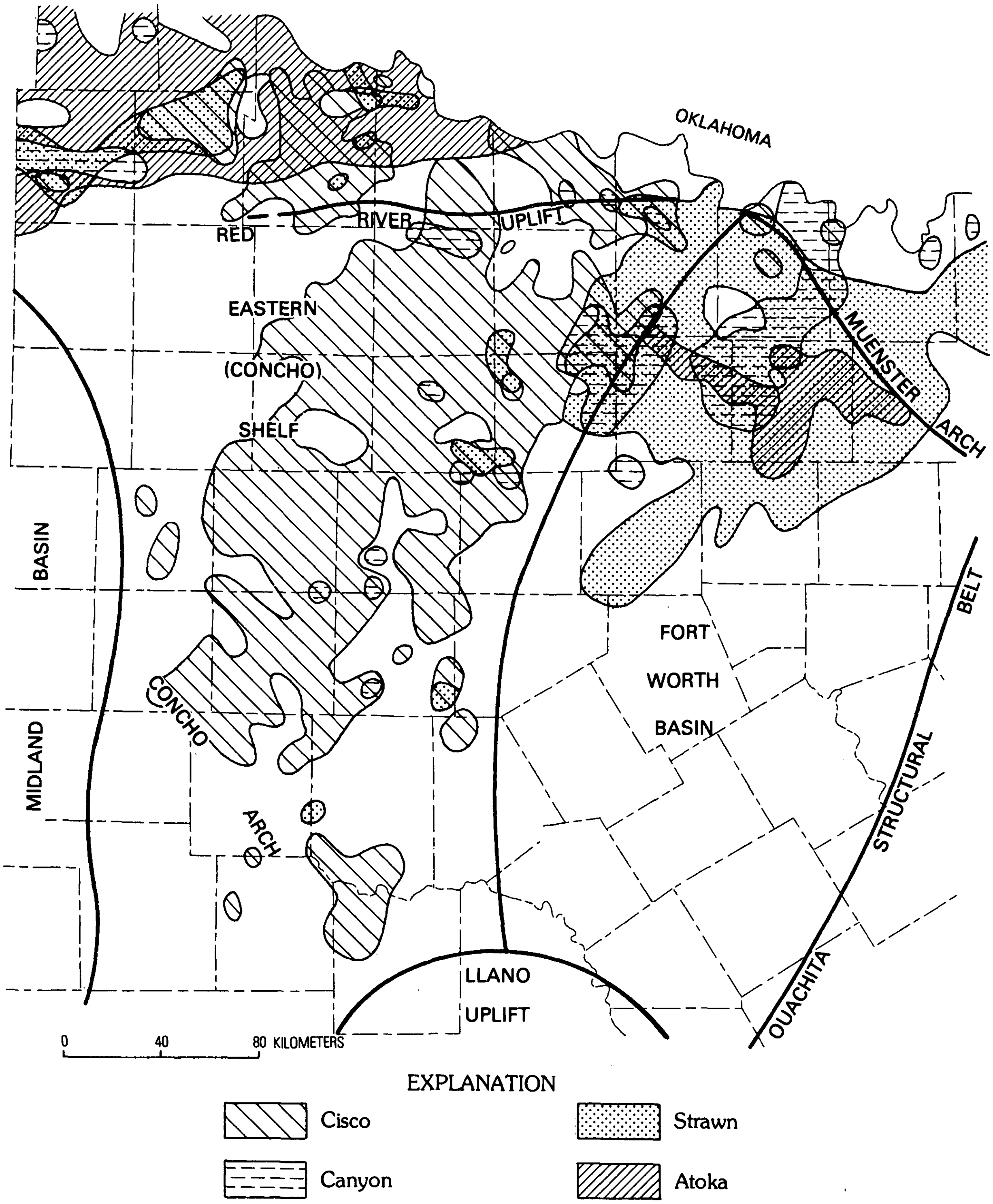

FIGURE 23.—Distribution of Pennsylvanian coal deposits, north-central Texas. From Mapel (1967).

County (fig. 2), where 200 feet of Canyon limestone bank facies, called the "Chico ridge or bank," supplies most industrial limestone for the Dallas-Fort Worth region. Other large quarries in both Strawn and Canyon limestones operate at Brownwood in Brown County, along Interstate Highway 20 in Parker, Palo Pinto, and Eastland Counties, and along other Texas highways.

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2001

The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States

GEOLOGICAL SURVEY PROFESSIONAL PAPER 1110–M–DD

ON THE COVER

Swamp-forest landscape at time of coal formation: lepidodendrons (left), sigillarias (in the center), calamites, and cordaites (right), in addition to tree ferns and other ferns. Near the base of the largest *Lepidodendron* (left) is a large dragonfly (70-cm wingspread). (Reproduced from frontispiece in Kukuk, Paul (1938), "Geologie des Niederrheinisch-Westfälischen Steinkohlengebietes" by permission of Springer-Verlag, New York, Inc.)

The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States—

M. Iowa, by Matthew J. Avcin and Donald L. Koch
N. Missouri, by Thomas L. Thompson
O. Arkansas, by Boyd R. Haley, Ernest E. Glick, William M. Caplan, Drew F. Holbrook, and Charles G. Stone
P. Nebraska, by R. R. Burchett
Q. Kansas, by William J. Ebanks, Jr., Lawrence L. Brady, Philip H. Heckel, Howard G. O'Connor, George A. Sanderson, Ronald R. West, and Frank W. Wilson
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CECIL D. ANDRUS, *Secretary*

GEOLOGICAL SURVEY

H. William Menard, *Director*

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FOREWORD

The year 1979 is not only the Centennial of the U.S. Geological Survey—it is also the year for the quadrennial meeting of the International Congress on Carboniferous Stratigraphy and Geology, which meets in the United States for its ninth session. This session is the first time that the major international congress, first organized in 1927, has met outside Europe. For this reason it is particularly appropriate that the Carboniferous Congress closely consider the Mississippian and Pennsylvanian Systems; American usage of these terms does not conform with the more traditional European usage of the term "Carboniferous."

In the spring of 1976, shortly after accepting the invitation to meet in the United States, the Permanent Committee for the Congress requested that a summary of American Carboniferous geology be prepared. The Geological Survey had already prepared Professional Paper 853, "Paleotectonic Investigations of the Pennsylvanian System in the United States," and was preparing Professional Paper 1010, "Paleotectonic Investigations of the Mississippian System in the United States." These major works emphasize geologic structures and draw heavily on subsurface data. The Permanent Committee also hoped for a report that would emphasize surface outcrops and provide more information on historical development, economic products, and other matters not considered in detail in Professional Papers 853 and 1010.

Because the U.S. Geological Survey did not possess all the information necessary to prepare such a work, the Chief Geologist turned to the Association of American State Geologists. An enthusiastic agreement was reached that those States in which Mississippian or Pennsylvanian rocks are exposed would provide the requested summaries; each State Geologist would be responsible for the preparation of the chapter on his State. In some States, the State Geologist himself became the sole author or wrote in conjunction with his colleagues; in others, the work was done by those in academic or commercial fields. A few State Geologists invited individuals within the U.S. Geological Survey to prepare the summaries for their States.

Although the authors followed guidelines closely, a diversity in outlook and approach may be found among these papers, for each has its own unique geographic view. In general, the papers conform to U.S. Geological Survey format. Most geologists have given measurements in metric units, following current practice; several authors, however, have used both metric and inch-pound measurements in indicating thickness of strata, isopach intervals, and similar data.

FOREWORD

This series of contributions differs from typical U.S. Geological Survey stratigraphic studies in that these manuscripts have not been examined by the Geologic Names Committee of the Survey. This committee is charged with insuring consistent usage of formational and other stratigraphic names in U.S. Geological Survey publications. Because the names in these papers on the Carboniferous are those used by the State agencies, it would have been inappropriate for the Geologic Names Committee to take any action.

The Geological Survey has had a long tradition of warm cooperation with the State geological agencies. Cooperative projects are well known and mutually appreciated. The Carboniferous Congress has provided yet another opportunity for State and Federal scientific cooperation. This series of reports has incorporated much new geologic information and for many years will aid man's wise utilization of the resources of the Earth.

H William Menard

H. William Menard
Director, U.S. Geological Survey

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