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EVOLUTION OF APPALACHIAN-OUACHITA SALIENTS AND RECESSES FROM REENTRANTS AND PROMONTORIES IN THE CONTINENTAL MARGIN

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ABSTRACT. Distributions of thickness and facies of Late Precambrian and Paleozoic rocks bear specific relation to the curvature of salients (convex toward craton) and recesses (concave toward craton) along the length of the Appalachian–Ouachita structural system. Thus, the shape of salients and recesses was inherited from a structural framework that persisted from the inception of Appalachian–Ouachita geosynclinal sedimentation. Both the distribution of sedimentary units and the outlines of later compressional structures conform to a shape that suggests an orthogonally zigzag continental margin established in Late Precambrian time. The original shape of the continental margin may be the expression of several transform faults along a Late Precambrian rift. Reentrants (concave oceanward) and promontories (convex oceanward) in the continental margin are the sites of later structural salients and recesses respectively. The Appalachian–Ouachita stratigraphic succession may be divided into three genetic rock sequences: (1) Late Precambrian and Early Cambrian clastic-volcanic sequence, (2) Cambrian-Ordovician carbonate bank and associated marginal-basin shale facies, and (3) Ordovician through Pennsylvanian clastic wedges. The basal clastic-volcanic sequence, limited to interior structural belts, laps farthest cratonward across structural strike in salients. The facies boundary at the edge of the carbonate bank generally follows the sinuous trace of the salients and recesses; promontories of the bank facies extend farthest oceanward across structural strike in recesses. Clastic wedges are thickest in salients; thickness decreases across strike cratonward and also along strike toward recesses.

INTRODUCTION

The Appalachian–Ouachita structural system includes six regional salients (fig. 1, facing p. 1264). Salients are curves convex toward the craton; intervening recesses are concave toward the craton. Some interpretations depict salients and recesses as the result of post-orogenic bending or strike-slip offset of the deformed belt. However, recent papers present the conclusion that curves in mountain belts reflect irregularities in older continental margins (for example, Dewey and Burke, 1974; King, 1975; Rankin, 1975a; Rodgers, 1975; Thomas, 1975, 1976; Cebull and others, 1976); thus, a pre-orogenic framework of salient-recess curvature is implied.

The purpose of this paper is to show that the approximate outlines of Appalachian–Ouachita salients and recesses reflect the distributions of facies and thickness of Late Precambrian and Paleozoic sedimentary rocks. Stratigraphic data indicate that the shape of the large-scale structural framework was established at the inception of the Appalachian–

Ouachita geosyncline, and curves in the orogenic belt conform to that shape. Thus, the shape of salients and recesses may be interpreted to be a reflection of an orthogonally zigzag continental margin which may have been framed by several transform faults along a rift system initiated in Late Precambrian time (fig. 1, facing p. 1264).

REGIONAL STRUCTURAL GEOLOGY OF SALIENTS AND RECESSES

Salients and recesses are expressed in the belt of folded and thrust-faulted sedimentary rocks along the northwest side of the system (Valley and Ridge and part of the Appalachian Plateau) as well as in the belt of metamorphic rocks (Piedmont) adjacent on the southeast. A discontinuous belt of pre-Appalachian (Grenville-age) basement rocks in the cores of elongate anticlines and thrust blocks (Blue Ridge in North Carolina and Virginia, Reading Prong and Hudson Highlands from Pennsylvania to New York, Berkshire and Green Mountains of New England, northern Long Range of Newfoundland) follows the southeast side of the folded and thrust-faulted belt (figs. 1, 2). Grenville-age basement rocks are also exposed in isolated uplifts within the Piedmont (Chester dome, Avondale anticline, Baltimore domes, Sauratown Mountains, and Pine Mountain of Georgia). As used here, "interior structures" are adjacent to and within the metamorphic terrain, "frontal structures" are adjacent to the undeformed rocks of the craton. Because of curves in strike, the frontal structures in salients extend farther toward the center of the craton than those in recesses. That effect is enhanced because the salients commonly include frontal structures that end laterally toward the recesses.

In the eastern part of the Pennsylvania salient, structural strike curves northeastward from N65°E to N35°E (fig. 1), and in the New York recess that trend intersects structures which strike N10°E in the southern part of the Quebec salient (Rodgers, 1970, p. 71). The frontal folds of the Pennsylvania salient flatten and end northeastward beneath essentially undeformed beds in the Pocono Plateau (Wood and Bergin, 1970, p. 147). The New York recess is the angular junction between structures of adjacent salients, but the folded and thrust-faulted belt in the recess is much more narrow than that in the Pennsylvania salient.

At the structural front of the Quebec salient, Logan's Line curves northeastward from the New York recess and extends beneath the St. Lawrence River. Strikes in Quebec and in Newfoundland point to a recess beneath the Gulf of St. Lawrence. The structures of western Newfoundland seem comparable to other Appalachian salients (King, 1951, p. 68; Rodgers, 1970, p. 148).

Southwestward from the Pennsylvania salient, the width of the deformed belt decreases toward the Virginia recess (fig. 1). Some of the frontal structures converge eastward at their southern ends (Rodgers, 1970, p. 17), but others end southwestward at northwest-trending zones of left-lateral strike-slip displacement (Rodgers, 1963, p. 1532; Gwinn, 1964, p. 890). An angular junction of about 35 degrees between fold

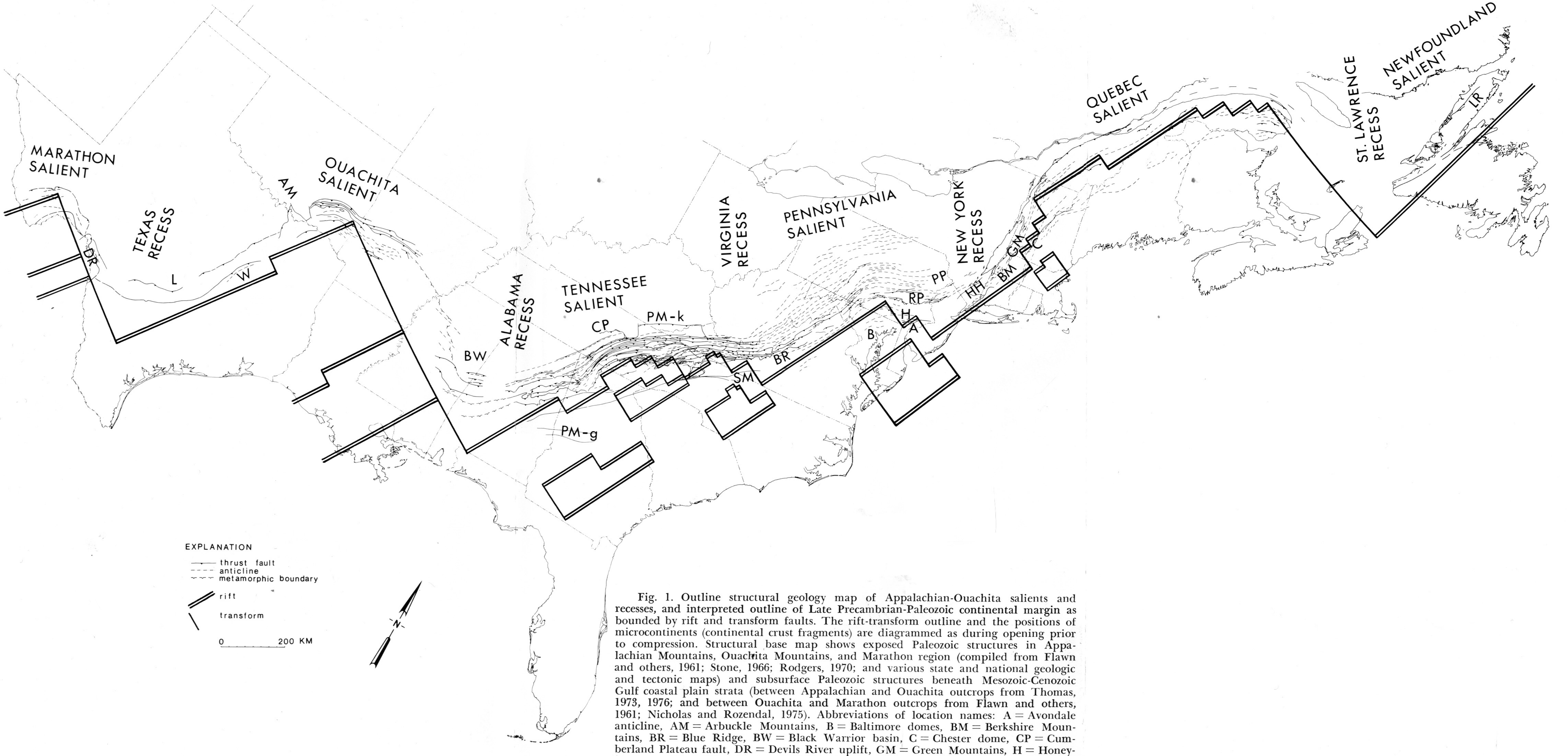


Fig. 1. Outline structural geology map of Appalachian-Ouachita salients and recesses, and interpreted outline of Late Precambrian-Paleozoic continental margin as bounded by rift and transform faults. The rift-transform outline and the positions of microcontinents (continental crust fragments) are diagrammed as during opening prior to compression. Structural base map shows exposed Paleozoic structures in Appalachian Mountains, Ouachita Mountains, and Marathon region (compiled from Flawn and others, 1961; Stone, 1966; Rodgers, 1970; and various state and national geologic and tectonic maps) and subsurface Paleozoic structures beneath Mesozoic-Cenozoic Gulf coastal plain strata (between Appalachian and Ouachita outcrops from Thomas, 1973, 1976; and between Ouachita and Marathon outcrops from Flawn and others, 1961; Nicholas and Rozendal, 1975). Abbreviations of location names: A = Avondale anticline, AM = Arbuckle Mountains, B = Baltimore domes, BM = Berkshire Mountains, BR = Blue Ridge, BW = Black Warrior basin, C = Chester dome, CP = Cumberland Plateau fault, DR = Devils River uplift, GM = Green Mountains, H = Honeybrook uplift, HH = Hudson Highlands, L = Llano uplift, LR = Long Range, PM-g = Pine Mountain of Georgia, PM-k = Pine Mountain of Kentucky and Tennessee, PP = Pocono Plateau, RP = Reading Prong, SM = Sauratown Mountains, W = Waco uplift.

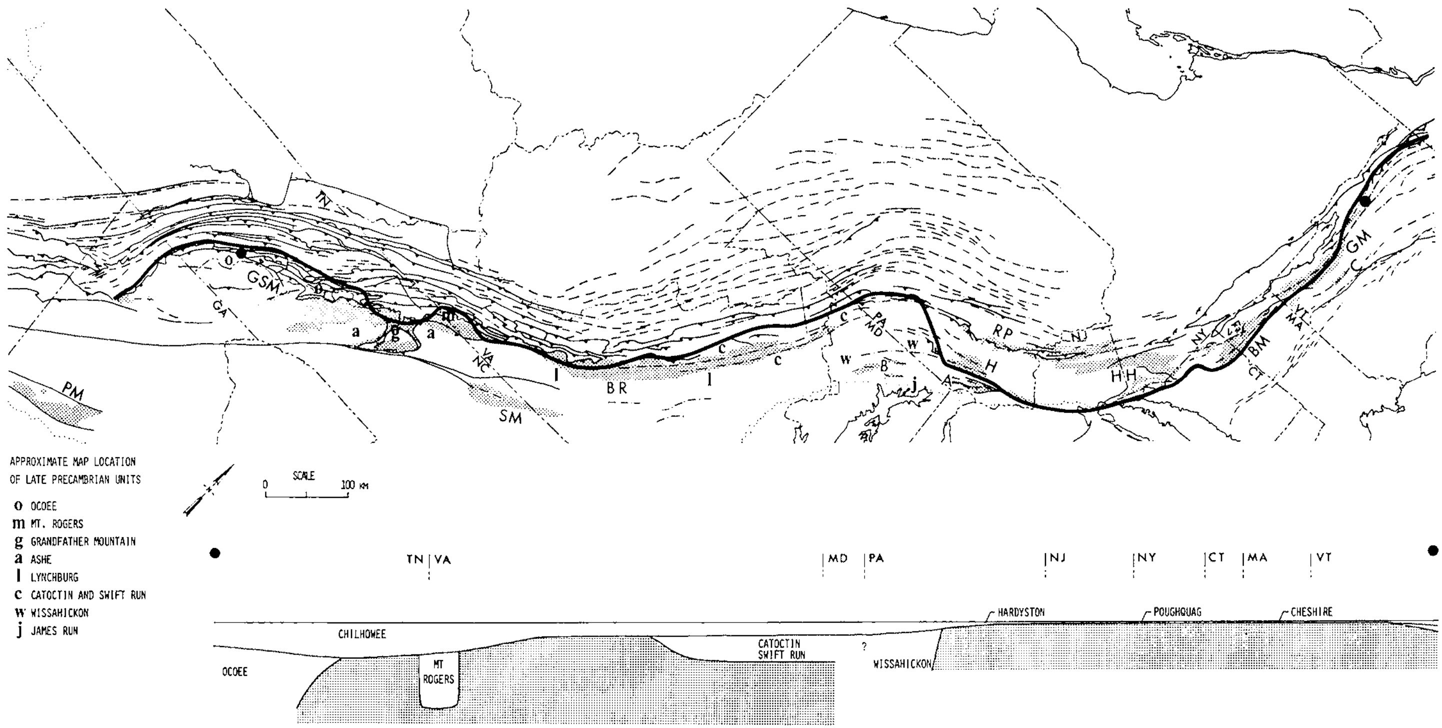


Fig. 2. Distribution of Late Precambrian-Early Cambrian clastic-volcanic sequence. Heavy line on map shows cratonward limit of Late Precambrian clastic-volcanic rocks. Diagrammatic cross section (not to scale vertically) shows relation of stratigraphic units along basement uplifts from Blue Ridge to Green Mountains; datum of cross section is top of Lower Cambrian clastic sequence. Stipple on map shows area of outcrop of Grenville-age basement rocks; stipple on cross section shows Grenville-age basement. Data from Doll and others (1961), Brown (1970), King (1970), Rankin (1970, 1975b), Palmer (1971), Higgins (1972), Rodgers (1972), United States Geological Survey (1974). Abbreviations of location names: A = Avondale anticline, B = Baltimore domes, BM = Berkshire Mountains, BR = Blue Ridge, C = Chester dome, GM = Green Mountains, GSM = Great Smoky Mountains, H = Honeybrook uplift, HH = Hudson Highlands, PM = Pine Mountain, RP = Reading Prong, SM = Sauratown Mountains. Structural base map from figure 1; limit of Mesozoic-Cenozoic coastal plain strata shown by light dotted line.

axes marks the Virginia recess in the interior part of the folded and thrust-faulted belt (fig. 1).

Strike-slip faults define both ends of the Pine Mountain thrust block at the front of the Tennessee salient (fig. 1). Southwest of the end of the Pine Mountain fault, the Cumberland Plateau fault extends southwestward from a right-lateral fault zone into the Cumberland Plateau where it ends (Wilson and Stearns, 1958, p. 1295; Tennessee Div. Geology, 1966; Milici, 1970, fig. 4).

Thrust faults diminish southward from the Tennessee salient, and southward-plunging folds are common in northwestern Georgia (Butts and Gildersleeve, 1948; Rodgers, 1970, p. 56). South-trending structures in northwestern Georgia intersect southwest-trending structures at an angular junction that defines the northeastern part of the Alabama recess (fig. 1). The southwest-trending structures extend farther southwest and continue beneath Mesozoic and younger beds of the Gulf coastal plain in western Alabama. In the subsurface of eastern Mississippi, strike curves westerly (Thomas, 1973, fig. 4) and defines the southwestern part of the Alabama recess. Like other recesses, the Alabama recess is relatively narrow. Unlike other recesses, it extends for some distance along strike, and perhaps this part of the structural system should be considered as a complex of two recesses (in northwestern Georgia and in eastern Mississippi) separated by a small-magnitude salient in Alabama.

A curve in the structural front from central Mississippi northwestward to the exposed Ouachita structures in central Arkansas defines the eastern part of the Ouachita salient (Thomas, 1973, p. 386, 1976, p. 326). The scale of the curve from the Alabama Appalachians to the Ouachita salient is similar to that from the New York recess to the Pennsylvania salient (fig. 1).

Structural strike curves around the Ouachita salient, and Paleozoic structures extend southward beneath Mesozoic coastal plain beds in southern Oklahoma. Subsurface data in Texas define the tracing of the structural system from the Ouachita salient into a recess around the Llano uplift and farther west into the Marathon salient (fig. 1). The frontal zone of the deformed belt is narrower in the Texas recess than in either of the salients (Flawn, *in* Flawn and others, 1961, p. 166). Southeast of the Marathon salient, low-grade metamorphic rocks of the Ouachita interior zone have been thrust onto foreland facies rocks on the Devils River uplift (Flawn, *in* Flawn and others, 1961, p. 172; Nicholas and Rozendal, 1975, p. 199).

STRATIGRAPHY

Classically the stratigraphic succession in the Appalachians has been recognized as being thicker and more dominantly clastic than the equivalent section on the craton (Hall, 1859, p. 66); similar generalizations apply to stratigraphic sequences in the Ouachita Mountains (Miser, 1921) and Marathon region (King, 1937, p. 20). However, Appalachian–Ouachita geosynclinal sequences are not uniform along strike but are marked by significant lateral variations in thickness and composition. The oldest beds rest unconformably on Grenville-age metamorphic rocks which constitute the basement of Appalachian geosynclinal deposits. Strata that accumulated within the framework of the Appalachian–Ouachita geosyncline may be classed into three major genetic types: (1) Late Precambrian-Early Cambrian clastic-volcanic sequence, (2) Cambrian-Ordovician (and younger) carbonate bank and associated marginal-basin shale facies, and (3) Ordovician through Pennsylvanian clastic wedges.

BASAL CLASTIC-VOLCANIC SEQUENCE

The Late Precambrian-Early Cambrian clastic-volcanic sequence overlies Grenville-age basement and continues upward into *Olenellus*-bearing beds (Rodgers, 1956, 1972, p. 507). The Late Precambrian part of the sequence locally is very thick, but it is limited to interior structural belts and pinches out toward the craton (fig. 2). From the Tennessee salient to the Pennsylvania salient, the cratonward limit of Late Precambrian rocks nearly coincides with the Blue Ridge. Late Precambrian rocks extend cratonward across the axis of the Blue Ridge in both salients but are limited to the southeast limb of the Blue Ridge in the Virginia recess (fig. 2). Where Late Precambrian rocks are absent, Lower Cambrian sandstones unconformably overlie basement along the northwest side of the Blue Ridge in the Virginia recess and similarly in the highlands in the New York recess. Limited exposures in the Quebec and Newfoundland salients indicate similarity to these distribution patterns, but Precambrian rocks are not exposed in the Ouachita and Marathon salients.

The relatively thin, Early Cambrian part of the sequence oversteps Late Precambrian strata and extends onto the craton as the basal transgressive sandstone (Brown, 1970, p. 340; Schwab, 1972, p. 80) of the Sauk Sequence (as defined by Sloss, 1963, p. 95). Sources of Late Precambrian sediment appear diverse; however, the quartzose sand of the Early Cambrian was transported eastward from the craton (Whitaker, 1955, p. 765; Brown, 1970, p. 343; Schwab, 1970, p. 362, 1972, p. 73; Whisonant, 1970, p. 2785).

In the Tennessee salient, the Late Precambrian Ocoee Supergroup of clastic sedimentary rocks is at least 12 km thick in the central Great Smoky Mountains (Hadley, 1970, p. 247), but northeastward along strike, the Ocoee wedges out and is overstepped by the Lower Cambrian Chilhowee Group (King, 1970, p. 39). Much of the Ocoee (especially Great Smoky Group) clastic sediment probably was derived from basement rocks on the northeast, but the lower Ocoee (Snowbird Group) had a provenance of basement rocks on the east and southeast (Hadley and Goldsmith, 1963, p. B47; King, 1964, p. 69, 1970, p. 43; Hadley, 1970, p. 249). Northeast of the limit of the Ocoee, the Chilhowee Group rests directly on Grenville-age basement; however, locally on the northwest side of the Blue Ridge, the Late Precambrian Mt. Rogers Volcanic Group intervenes (fig. 2). Maximum thickness of the Mt. Rogers is about 3 km (Rankin, 1970, p. 231). To the southeast, Mt. Rogers equivalents are included in thick successions of sedimentary and volcanic rocks of the Grandfather Mountain Formation (King, 1970, p. 38) and Ashe Formation (Rankin, 1970, p. 235, 1975b, p. 306). The Ashe Formation is evidently continuous northeastward with the Lynchburg Formation (Rankin, 1975b, p. 309), and the Lynchburg is restricted to the southeast side of the Blue Ridge in the Virginia recess.

Along the northeastern Blue Ridge, arkosic conglomerate of the basal Lynchburg continues into the thinner Swift Run Formation which extends to the northwest limb of the Blue Ridge (fig. 2). The Swift Run is overlain by the Late Precambrian Catoctin volcanics, which, along with the Swift Run, pinch out southwestward between Grenville-age basement and Lower Cambrian sandstone along the northwest limb of the Blue Ridge (Brown, 1970, p. 340). Along the southeast side of the Blue Ridge, the Catoctin either overlies or interfingers with the Lynchburg (Rankin, 1975b, p. 309).

In the southeastern part of the Pennsylvania salient, the Glenarm "Series" (Setters Quartzite, Cockeysville Marble, Wissahickon Formation, and other units described by Higgins, 1972) overlies Grenville-age basement (Baltimore domes and Avondale anticline, fig. 2). Glenarm rocks include a probable age span of latest Precambrian (600 to 650 m.y.) to Late Ordovician (Higgins, 1972, p. 1017). The Wissahickon is a flysch succession which is probably more than 6 km thick and was derived primarily from the east (Hopson, 1964, p. 129-130). Part of the Wissahickon grades eastward into volcanic and vocaniclastic rocks of the James Run Formation (Higgins, 1972, p. 1021). Higgins (p. 1021) considers the James Run to be part of the "Atlantic Seaboard volcanic province" along the Appalachians from Canada to Georgia, although he also recognizes the importance of "crystalline basement" rocks in the eastern provenance of the Wissahickon.

Farther north and west in the Pennsylvania salient (Honeybrook uplift), Lower Cambrian Chilhowee sandstone rests on basement; the Chilhowee reflects beach and shallow-marine environments north and west of the deeper Wissahickon basin (fig. 2). Previously, the Wissahickon was thought to be equivalent to the Late Precambrian Lynchburg (Hopson, 1964, p. 207), and absence of Wissahickon below Chilhowee on the northwest was inferred to reflect an unconformity. More recent work suggests that the Wissahickon spans the age of the Chilhowee (Higgins, 1972, fig. 21), although the lower Wissahickon may be older. Thus, part of the Wissahickon and related rocks accumulated in a contemporaneous basin southeast of the shelf on which Chilhowee sandstone was deposited.

On the Baltimore Gneiss domes, the Wissahickon overlies a quartzite-carbonate sequence (Setters-Cockeysville) which rests on basement rocks. The Setters-Cockeysville is interpreted either as in normal sequence below Wissahickon (Hopson, 1964, fig. 31; Wise, 1970, fig. 7; Higgins, 1972, fig. 21) or as lower Paleozoic over which the Wissahickon has been thrust (Rodgers, 1970, p. 191, 1972, p. 515; Rankin, 1975b, p. 322). Regardless of the structural-stratigraphic setting, the Setters-Cockeysville is some distance southeast of the apparent southeast limit of Chilhowee sandstone. In a similar setting, quartzite apparently rests unconformably on basement rocks in the Sauratown Mountains of North Carolina (fig. 2), and Rankin (1975b, p. 315) suggests that the quartzite was deposited on an isolated topographic high of basement rocks. A similar depositional setting may be postulated for the Setters of the Baltimore domes, and, thus, the Setters is not continuous or necessarily coeval with the Chilhowee (Rankin, 1975b, p. 316).

Northeastward from the Pennsylvania salient, along strike of basement highlands within the New York recess and northward, basal Cambrian sandstones (Hardyston, Poughquag, Lowerre, Dalton, and Cheshire) rest unconformably on Grenville-age basement rocks (fig. 2), and no Late Precambrian strata are preserved (Doll and others, 1961; Hall, 1968; Palmer, 1971, figs. 5 and 8; Rodgers, 1972, fig. 1). In the New York recess, the basal Cambrian sandstone is unusually thin and may be younger than that in the Appalachians farther southwest and northeast (Palmer, 1971, p. 189).

Along part of the Green Mountains, the Lower Cambrian Cheshire Quartzite is separated from Grenville-age basement by clastic sedimentary rocks that possibly are Late Precambrian (Bird and Dewey, 1970, fig. 5; Rodgers, 1972, p. 510), but part may be Cambrian (Palmer, 1971, p. 184). Bird and Dewey (1970, p. 1044) interpret the pre-Cheshire rocks to be continental rise sediments derived from the craton. Similarly, along the Long Range in Newfoundland, Late Precambrian clastic and volcanic rocks overlie Grenville-age basement (Bird and Dewey, 1970, fig. 5; Rodgers, 1972, p. 510). Late Precambrian volcanic rocks in Newfoundland are attributed to rifting (Bird and Dewey, 1970, p. 1045; Williams and Stevens, 1974, p. 785).

Distribution of the Late Precambrian part of the clastic-volcanic sequence suggests deposition in steep-sided basins along a rift system (Rankin, 1975b). The cratonward limit of Late Precambrian rocks marks the approximate continental margin at the edge of the rift system, although some thinner Late Precambrian units (for example, part of Catoctin) may extend cratonward from the rift. The same line marks the approximate oceanward limit of Early Cambrian shelf deposits, but some of the quartzose sandstone extends eastward over part of the Late Precambrian basin fill. Farther east, within the rift system, Grenville-age basement rocks evidently were broken into a spreading horst and graben complex. Some of the basement horst blocks supplied clastic sediment to the Late Precambrian basin fill, and some evolved as platforms where quartzose sand similar to that of the craton was deposited on Grenville-age basement. Such local basement-block shelves are indicated in the Baltimore domes, Sauratown Mountains, and Pine Mountain of Georgia (fig. 2). Graben floors within the rift system may have included oceanic crust in addition to continental crust which was thinned by crustal extension (see also Rankin, 1975b, p. 316). Rankin's (p. 313) review of chemical compositions of the Late Precambrian volcanic rocks indicates characteristics associated with tension and rifting, and Lowry (1974, p. 595) concludes that the basalts of the Catoctin and Lynchburg are related to tension along a hinge line. The approximate trace of the Late Precambrian rift is reflected in the present shape of Appalachian salients and recesses (figs. 1, 2); Late Precambrian rocks extend farthest toward the craton (across present structural strike) within the salients.

CARBONATE BANK

The quartzose sandstone at the top of the clastic-volcanic sequence is overlain by an extensive carbonate facies that ranges from Early Cambrian to Middle Ordovician and younger (fig. 3). The shallow-marine carbonate facies extends over much of the craton, but, away from the craton (oceanward) along the Appalachian–Ouachita system, the carbonate facies changes into a deep-water dark-colored shale facies. The shale facies includes interbeds of quartzose sandstone, chert, and locally, near the facies boundary, carbonate boulders. Rodgers (1968) interprets the facies boundary to be the steep edge of a carbonate bank (fig. 3). That interpretation may be extended along the Appalachian–Ouachita system from Newfoundland to Texas (Bird and Dewey, 1970, p. 1044; Palmer, 1971, p. 208; Thomas, 1972a, p. 103, 1976, p. 328; Nicholas and Rozendal, 1975, p. 211).

In the middle part of the Quebec salient, overthrust rocks along Logan's Line are in the shale facies (Rodgers, 1970, p. 117). But southward toward the New York recess, the facies boundary crosses structural strike diagonally, and the carbonate facies extends eastward into the deformed belt (Rodgers, 1968, p. 146). In the New York recess, the carbonate facies (Inwood Marble) is infolded with basement rocks in the interior structures (Hall, 1968, p. 124). At the eastern end of Gaspé, Cambrian limestone comparable to that of the frontal Appalachians is exposed locally within the deformed belt (Rodgers, 1970, p. 130). Along the middle part of the salient, the position of the bank edge is covered by overthrust rocks of the shale facies, and allochthonous masses (for example, the Taconic slates) of the shale facies overlie parts of the carbonate bank (Rodgers, 1968, p. 146). The bank-edge boundary of the carbonate facies apparently defines a broad curve in which the Quebec salient is centered, but the radius of curvature of the facies boundary is less than that of the structural front (fig. 3). The bank-edge facies boundary is identified in Newfoundland southeast of the Long Range (Rodgers, 1968, p. 146; Bird and Dewey, 1970, fig. 6).

In the Pennsylvania salient, shallow-water carbonate rocks grade southeastward into the deep-water euxinic Conestoga (and Frederick) Limestone (Rodgers, 1968, p. 145; Wise, 1970, p. 318; Palmer, 1971, p. 196). The arcuate outline of the facies boundary approximately coincides with the structural curve near the center of the salient (Rodgers, 1968, fig. 10-3; Wise, 1970, fig. 3); however, the basinal facies does not

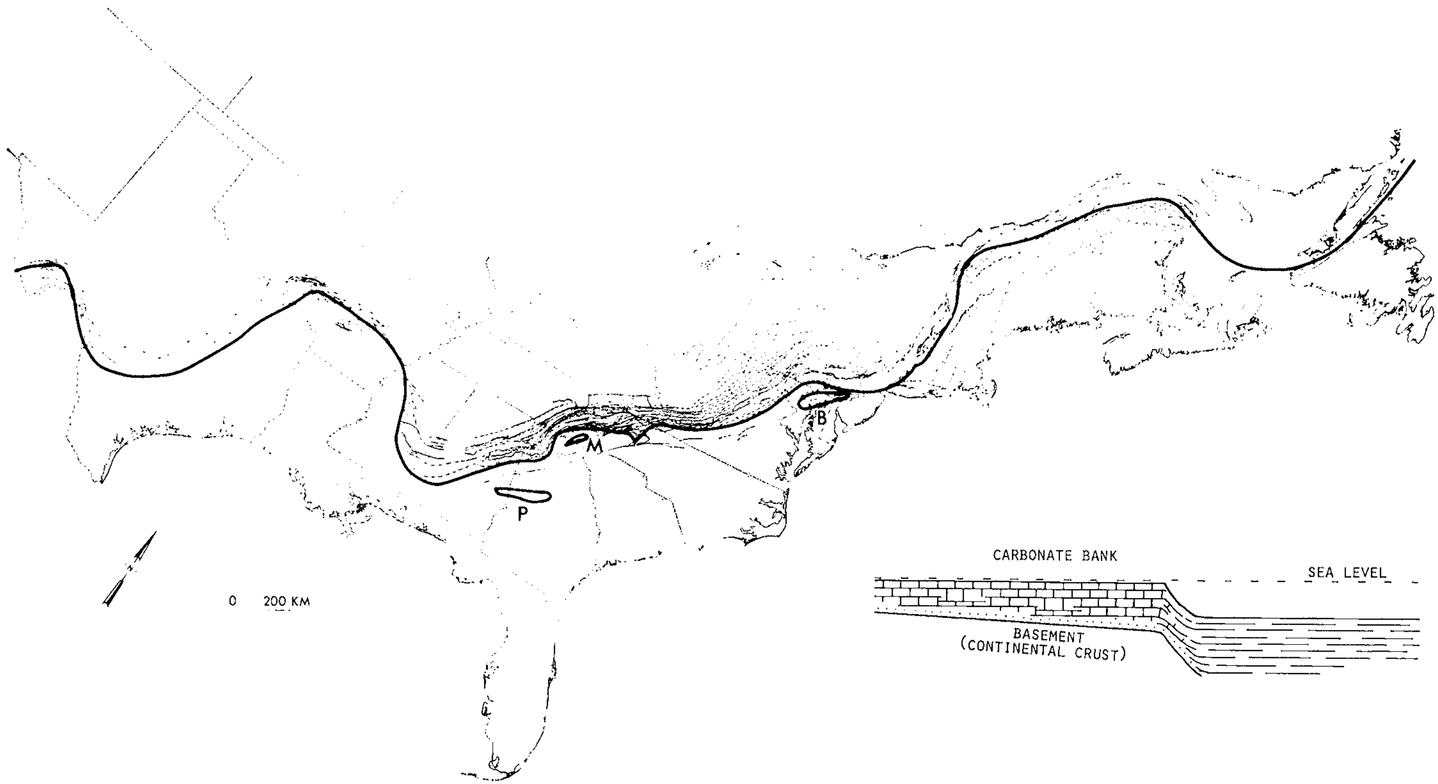


Fig. 3. Outline of Cambrian-Ordovician carbonate-bank facies. Heavy line on map shows approximate position of carbonate-bank edge, and diagrammatic cross section shows interpreted depositional framework of the carbonate bank and deep-water shale facies. Outlying areas of the carbonate-bank facies are interpreted to be isolated banks formed on microcontinents (B = Baltimore domes, P = Pine Mountain of Georgia). In the Murphy Marble belt (M), shallow-water carbonate rocks are within a stratigraphic sequence above Late Precambrian clastic sedimentary rocks and are not associated with a microcontinent. Data from Flawn and others (1961), Rodgers (1968), Wise (1970), Thomas (1972a, 1976), Nicholas and Rozendal (1975), Rankin (1975b). Structural base map from figure 1.

extend far across strike toward the frontal structures (fig. 3). Along the southern arc of the Pennsylvania salient, the bank-edge facies boundary follows the eastern side of the Blue Ridge in northern Virginia (Rodgers, 1968, p. 148).

Southward from Virginia, the edge of the carbonate bank is not clearly defined. The carbonate facies in the Tennessee salient extends eastward beneath the Blue Ridge overthrust. The Rome-Conasauga sandstone and shale interbeds within the carbonate facies in Tennessee were derived from the craton (Rodgers, 1953, p. 44, 47; Palmer, 1971, p. 207) and are not part of a marginal-basin shale facies. The carbonate facies may be represented by marble units in the Piedmont of Georgia and Alabama (Rodgers, 1968, p. 148), but correlation and possible continuity of the marble units with the carbonate facies farther northwest are not established. The Sylacauga Marble of the Alabama Piedmont is probably part of the lower Paleozoic carbonate facies (Shaw, 1970, p. 257; Carrington, 1973). The structural setting of the Sylacauga Marble in the Alabama recess is similar to that of the Inwood Marble in the New York recess, but no basement rocks are seen to be involved in the structures of the Sylacauga area.

The carbonate facies extends in the subsurface from the Alabama Appalachians westward through the Black Warrior basin to northern Arkansas and farther west and south around the Ouachita Mountains to the Arbuckle Mountains (fig. 3). The lower Paleozoic succession in the Ouachita Mountains is characterized by black shale and contains interbeds of chert, sandstone, limestone, and boulder conglomerate (Goldstein, *in* Flawn and others, 1961; Sterling, Stone, and Holbrook, 1966; Stone, Haley, and Viele, 1973; Thomas, 1976, p. 328). Distribution of black shale and limestone interbeds suggests that the carbonate facies grades westward to the shale facies at a northwest-trending bank edge in Mississippi (Thomas, 1972a, p. 103). In the Ouachita salient, the frontal structures are within the shale facies which has been thrust over the carbonate-bank edge. Eastward toward the Alabama recess, the structural front crosses the facies boundary diagonally, and, in eastern Mississippi and Alabama, the carbonate facies is deformed by Appalachian structures (fig. 3).

Southward from the Ouachita salient, in the subsurface of Texas, the carbonate facies west of the Ouachita deformed belt extends southwest beyond the Llano uplift (fig. 3). Along the southern arc of the Ouachita salient, the Shell Barrett well on the Waco uplift penetrated an intensely deformed carbonate (marble) sequence above basement (Rozendal and Erskine, 1971; Nicholas and Rozendal, 1975, p. 202). On the Waco uplift, the carbonate facies is involved in Ouachita deformation and is overlain by allochthonous metasedimentary rocks (Nicholas and Rozendal, p. 211). By analogy with the southern arc of the Quebec salient, the facies boundary in Texas may cross structural strike diagonally. Known distribution of the two facies permits the possibility that in the Ouachita salient the structural front is in the shale facies, but southward toward the Texas recess the facies boundary crosses structural strike diagonally, and the carbonate facies extends across strike into the deformed belt (perhaps to the east side of the Waco uplift). West of the Texas recess the carbonate facies on the Devils River uplift is overridden by allochthonous interior zone rocks (Flawn, *in* Flawn and others, 1961, p. 172; Nicholas and Rozendal, 1975, p. 199), but the bank edge (possibly on the southwest limb of the uplift, see Flawn, *in* Flawn and others, 1961, pl. 2) is not defined by subsurface data.

The stratigraphic section in the Marathon salient is similar to that in the Ouachita salient. The lower Paleozoic shale and chert interval in the Marathon salient contains more limestone than the Ouachita sequence, and the Marathon section appears transitional between the shelf carbonate facies and the deep-water black shale facies (Flawn, *in* Flawn and others, 1961, p. 51). Carbonate boulders in the Ordovician Woods Hollow Shale are interpreted to be bank-edge boulder slides (King, 1937, p. 47; Nicholas and Rozendal, 1975, p. 195).

The edge of the carbonate bank follows a sinuous trace along the Appalachian–Ouachita system. Reentrants (curves concave oceanward) in the bank edge coincide with structural salients, and promontories of the bank coincide with structural recesses. However, the trace of the facies boundary at the edge of the bank is not exactly parallel with structural strike. The carbonate facies extends farther oceanward across structural strike in recesses than in salients, and locally the facies boundary crosses structural strike diagonally. In the Quebec and Ouachita salients, the position of the bank edge is obscured beneath overthrust basinal shale facies, and the shale facies extends to the surficial frontal structures in the salients. In contrast, the carbonate facies can be seen to be deformed in the interior structures in the New York and Alabama recesses. In the Pennsylvania salient, overthrust shale facies does not cover the bank edge, and, there, outcrops show that the facies boundary curves around the salient. The coincidence of curves in the bank edge with salients and recesses indicates that the shape of the structural system has conformed to a fundamental structural framework on which the original bank edge was established. Rodgers (1968, p. 148) has suggested that the bank edge may have formed along the existing margin of continental crust. The carbonate-bank facies boundary delimits two mutually exclusive sedimentary regimes. However, oceanward from the facies boundary (within the regime of basinal shale facies), some isolated areas contain carbonate rocks (fig. 3). On the Baltimore domes and the Pine Mountain block of Georgia, a quartzite and marble sequence overlies Grenville-age basement rocks. These are interpreted to be isolated small carbonate banks which formed on fragments (microcontinents) of continental basement rocks within the opening geosyncline (see also Rankin, 1975b, p. 316).

CLASTIC WEDGES

Introduction, definitions, and summary.—The carbonate-bank facies is terminated upward by a complex array of cratonward-prograding clastic wedges (as defined by King, 1959, p. 59; Krumbein and Sloss, 1963, p. 535). King (1959, p. 59) described clastic wedges as semicircular (fish-scale) in shape. Isopach lines and facies distributions are roughly concentric about the center of each wedge. The wedges thin and grade to carbonate facies not only toward the craton but also laterally along strike within the structural system. King (fig. 34) identified a number of clastic wedges and mapped the approximate outline and location of the wedges along the Appalachian system. Data compiled in this paper show that the thickest (central) part of each wedge coincides approximately with the center of a structural salient, and that the wedges thin laterally toward recesses (fig. 4). The different salients are characterized by one to three distinct clastic wedges, and the timing of wedge deposition appears unique for each salient.

A single clastic wedge is considered here to be a large-scale sedimentary unit of interrelated clastic rocks and is interpreted to represent a single major depositional cycle related to a single orogenic episode (see also Meckel, 1970, p. 49). Some clastic wedges begin with deep-water sediments, but others lack deep-water elements. Clastic wedges generally prograde radially from near the center. Concomitantly with an increase in extent of younger clastic strata, rocks in the central part of the wedge grade upward to more shallow-water or non-marine facies. The clastic wedges prograde over the area of the older carbonate bank, and parts of some wedges are overlapped by transgressive limestones that indicate reestablishment of the carbonate-bank environment as the supply of clastic sediment diminished. Limestone units within the succession of clastic wedges are referred to here as "interwedge carbonates." Commonly, intertonguing with the carbonate facies is restricted to the distal parts of a wedge. In some wedges, the proximal (central) part is truncated, and the upward gradation of facies is interrupted by unconformity. Meckel (1970) has described the vertical and lateral compositional variations in three clastic wedges and has presented a general model for cratonward prograding of clastic-wedge sediments. Meckel (1970) described the clastic wedges in terms of a cross section perpendicular to Appalachian strike; however, application of the model parallel with strike shows thinning of wedge sediments along Appalachian strike away from the wedge center.

In the frontal part of a wedge, stratigraphic strike (as defined by linear sedimentary units, trend of isopach lines, trend of diachronous facies boundaries, et cetera) is approximately parallel with structural strike in the frontal structures of a salient. However, the radius of curvature of stratigraphic outlines is less than that of salients, and stratigraphic strike curves to intersect structural strike at the lateral margins of wedges (generally in recesses). In some places where stratigraphic strike crosses structural strike, the semicircular outline of wedge sedi-

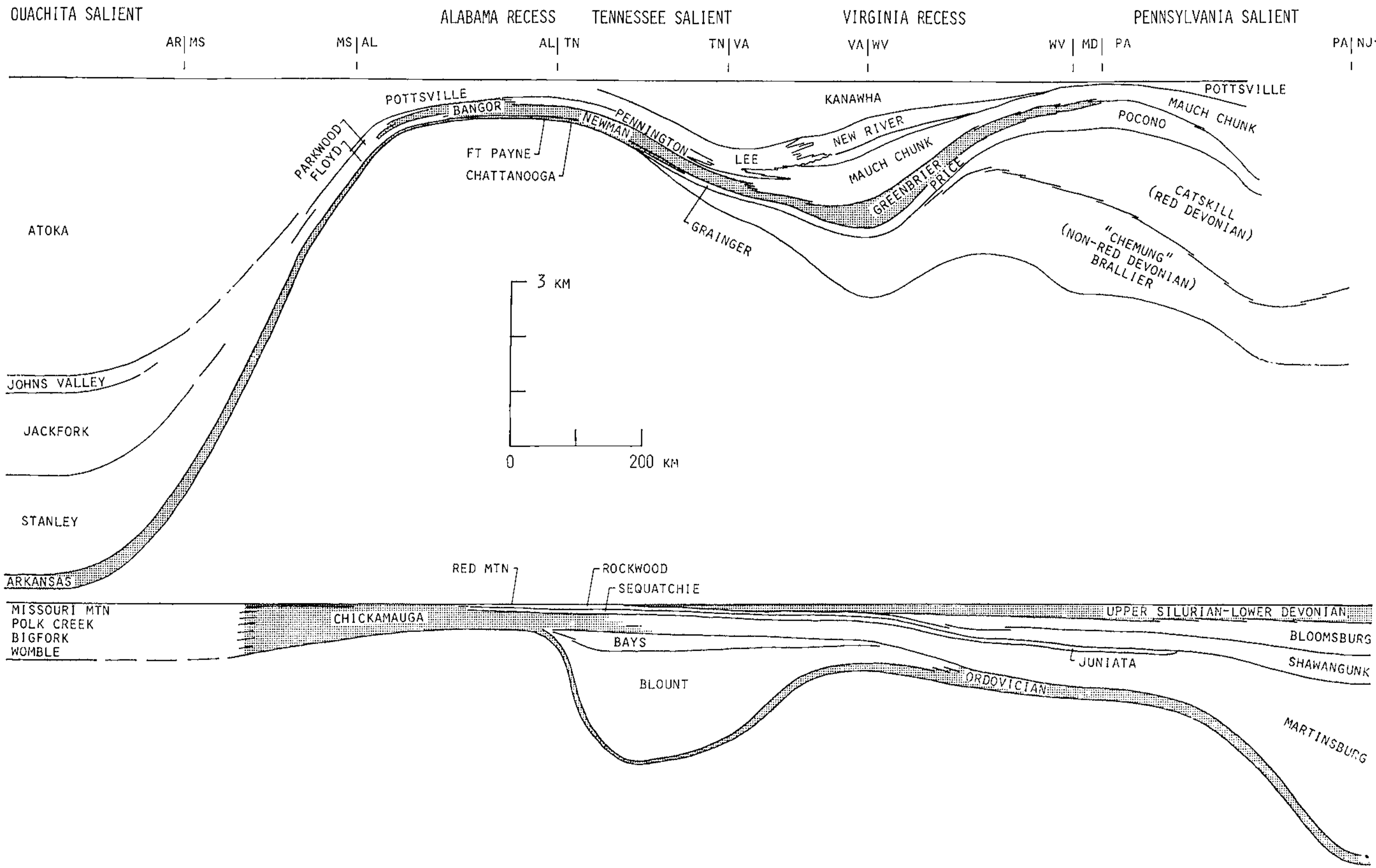
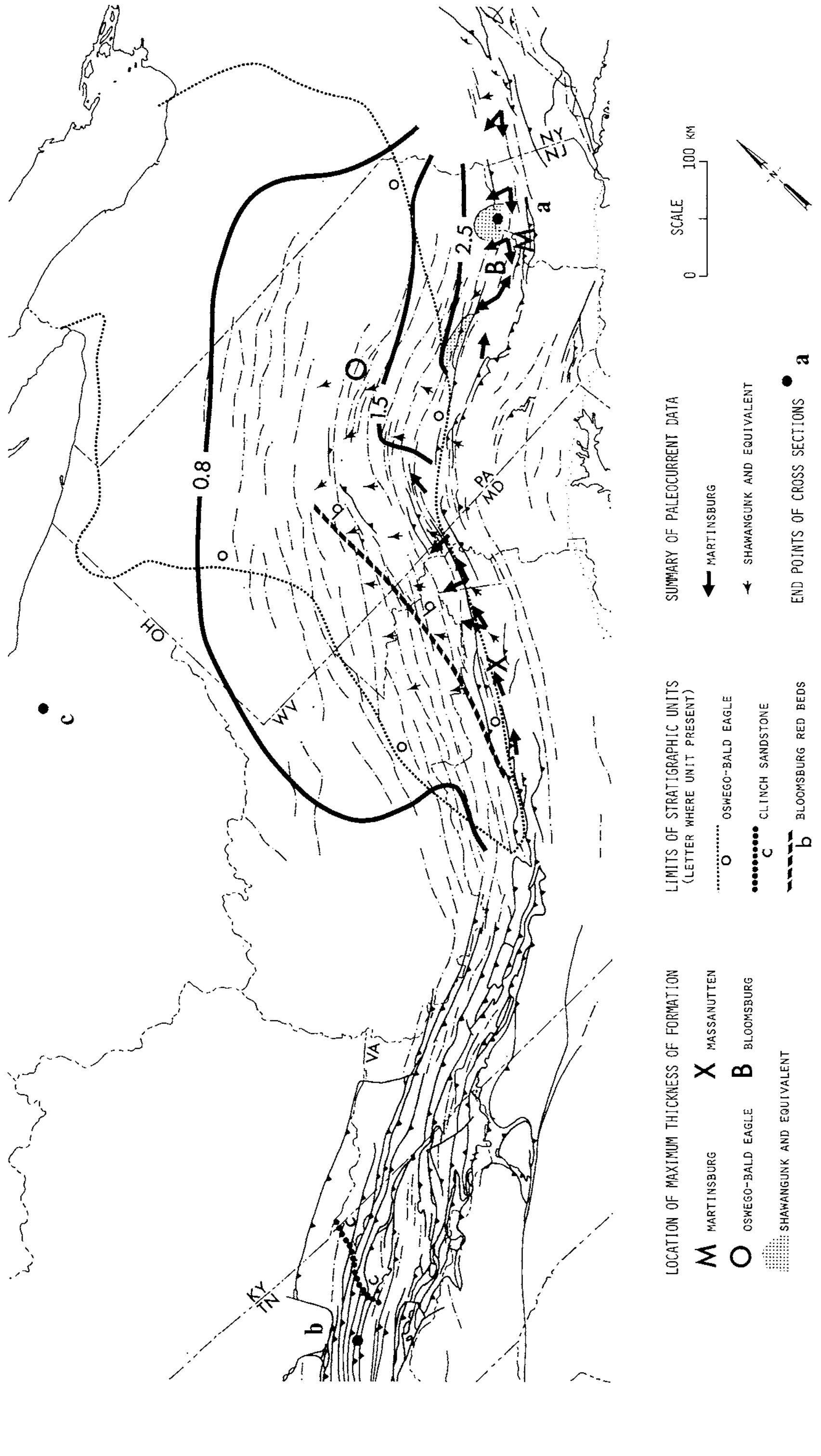


Fig. 4. Diagrammatic stratigraphic cross section of post-Early Ordovician clastic wedges along Appalachian-Ouachita structural system between New York and Arkansas. Datum of upper part of cross section is top of Pennsylvanian Pottsville-Kanawha-Atoka; datum of lower part of cross section is approximate top of Lower Devonian (base of Devonian in Ouachita salient). Base of upper part of cross section coincides with top of lower part. Explanation: no pattern = clastic-wedge rocks; stipple = interwedge carbonate units; pre-Stanley rocks in the Ouachita salient do not constitute a clastic wedge or interwedge carbonate but are interpreted to be a deep-water basin facies marginal to the carbonate bank. Compiled from figures 5, 6, 7, 8, and 9. The formations shown are not all in a single continuous stratigraphic succession; the formation names shown do not include all units in the succession at some localities.



LOCATION OF MAXIMUM THICKNESS OF FORMATION
M MARTINSBURG
O OSWEGO-BALD EAGLE
SHAWANGUNK AND EQUIVALENT
X MASSANUTTEN
B BLOOMSBURG
LIMITS OF STRATIGRAPHIC UNITS (LETTER WHERE UNIT PRESENT)
o OSWEGO-BALD EAGLE
c CLINCH SANDSTONE
b BLOOMSBURG RED BEDS
SUMMARY OF PALEOCURRENT DATA
MARTINSBURG
SHAWANGUNK AND EQUIVALENT
END POINTS OF CROSS SECTIONS a
SCALE
0
100 KM
N
NY
NJ
PA
MD
OH
WV
VA
KY
TN
0.8
1.5
2.5

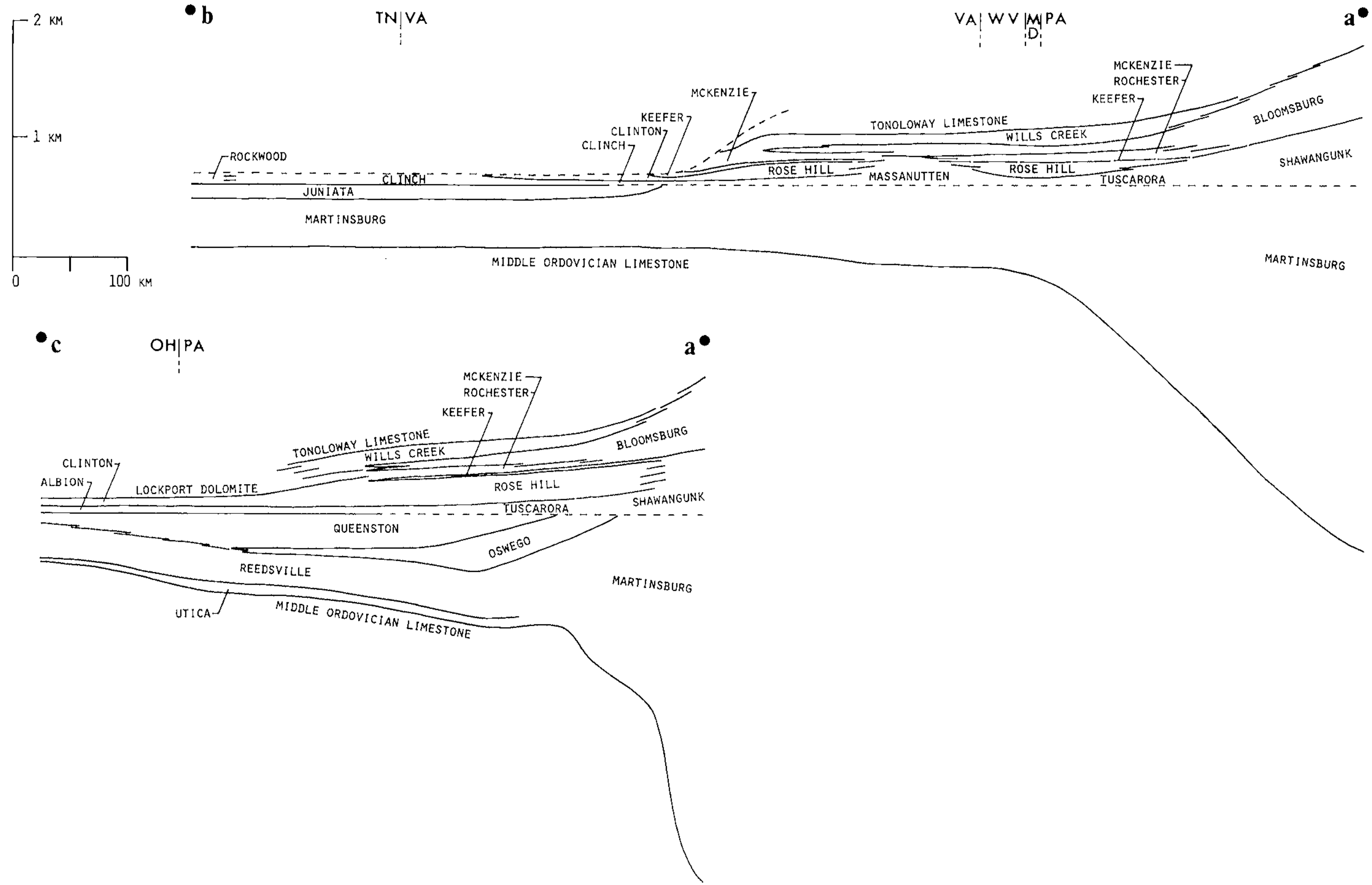


Fig. 5. Martinsburg-Shawangunk (Ordovician-Silurian) clastic wedge in Pennsylvania salient. Composite isopach map of Ordovician-Silurian clastic wedge (contour values in km). Datum of cross sections is top of Ordovician; dashed contact lines on cross sections indicate unconformities. Data from Butts (1940), Woodward (1941, 1959), Amsden (1955), Shearrow (1957), Harris and Miller (1958), Fettke (1961), Hoskins (1961), McBride (1962), Yeakel (1962), Cooper (1963), Tillman (1963), Drake and Epstein (1967), Colton (1970), Walker (1970), Dennison and Boucot (1974), Dennison and Wheeler (1975). Structural base map from figure 1.

ments is distorted, evidently because contemporaneous synclines provided channels along which clastic sediment was selectively transported farther from the wedge center.

The following discussions provide an overview of facies and thickness distributions and of sediment dispersal patterns in each of the clastic wedges in the Pennsylvania, Tennessee, and Ouachita salients (wedges described in order of decreasing age in each salient). Brief outlines of clastic wedges in the Marathon, Quebec, and Newfoundland salients are included for comparison.

Martinsburg-Shawangunk wedge.—The Martinsburg-Shawangunk clastic wedge in the Pennsylvania salient includes Ordovician and Silurian rocks (fig. 5). A turbidite sandstone and shale succession (Martinsburg Formation) grades upward into prograding deltaic and fluvial facies of the Oswego (Bald Eagle) Sandstone and Queenston (Juniata) Formation red beds (McBride, 1962, p. 87; Yeakel, 1962, p. 1534; Horowitz, 1966, p. 160; Meckel, 1970, p. 54). In the interior (southeastern) structures, the Oswego and Juniata are unconformably absent, and Silurian Shawangunk Conglomerate overlies Martinsburg. The Shawangunk thins and grades laterally into the extensive Tuscarora (Clinch) Sandstone (fig. 5). The Bloomsburg red beds at the top of the wedge are overlapped by a transgressive Silurian and Lower Devonian sequence of carbonate rocks, chert, and quartzose sandstone (Colton, 1970, p. 31; Oliver and others, 1967).

The maximum thickness of the Martinsburg-Shawangunk wedge, about 4 km, is in eastern Pennsylvania near the center of the Pennsylvania salient (fig. 5). The Martinsburg thins and the lower part grades to a limestone facies northwestward (cratonward) and southwestward along strike in the Virginia recess. Red beds of the Juniata extend southwestward into the Tennessee salient, and Clinch (Tuscarora) Sandstone extends across the Virginia recess and grades out southwestward in the Tennessee salient (fig. 5). Other elements of the wedge also thin southwestward into the Virginia recess.

Paleocurrent data from the Martinsburg indicate both transverse and longitudinal currents (McBride, 1962, p. 83). Dominant longitudinal currents flowed northeastward along strike around the curve of the salient in Pennsylvania and farther south but flowed southwestward in New York and New Jersey (McBride, p. 83). Petrographic data show that Martinsburg sandstones in the south were derived from a sedimentary terrain, but those in the north were derived from a provenance of metamorphic and plutonic rocks (McBride, p. 66). McBride (p. 84-87) concludes that a lateral sediment source on the southeast supplied clastic sediment to a longitudinal transport system.

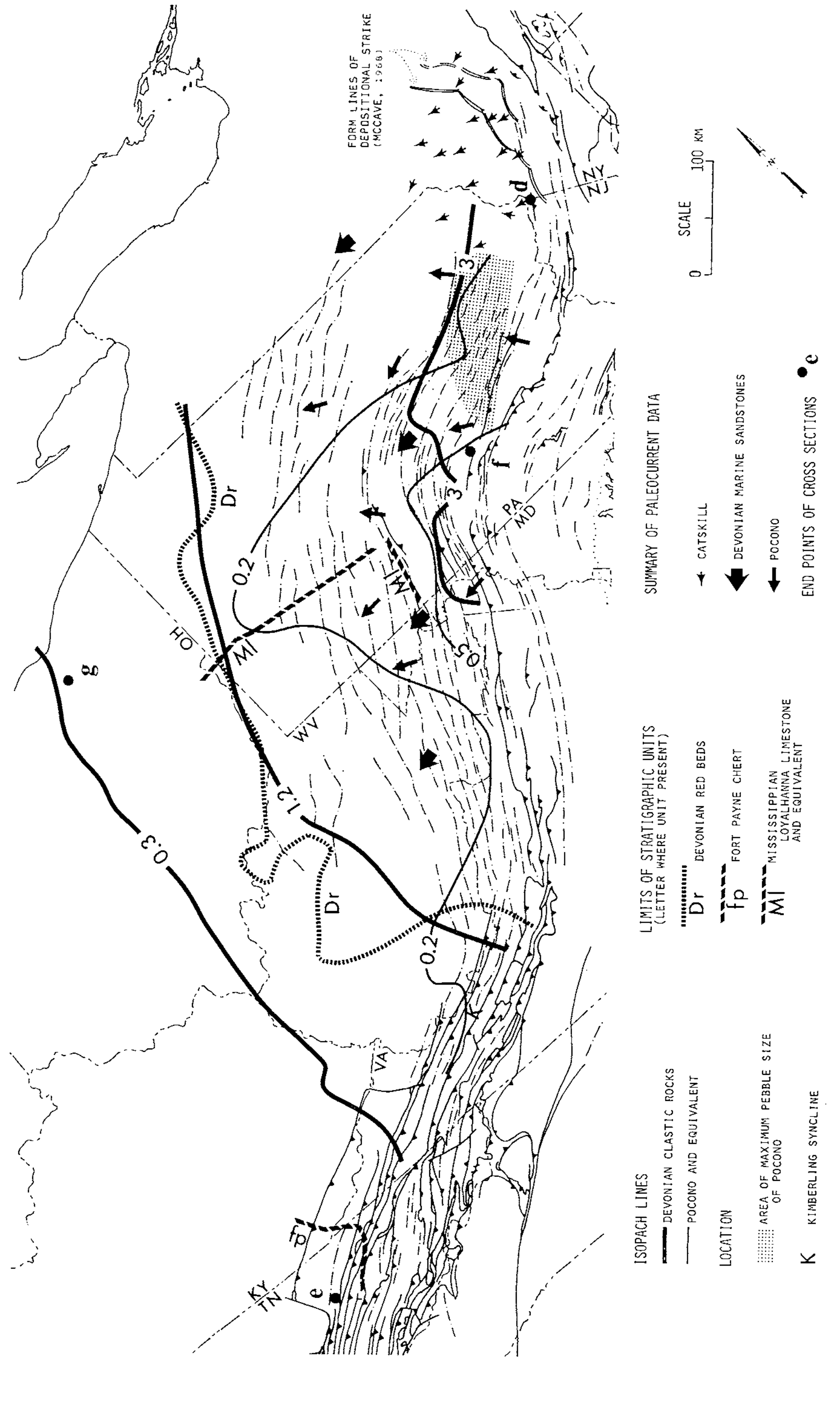
The Shawangunk Conglomerate thins radially from a maximum of about 750 m in eastern Pennsylvania (Colton, 1970, fig. 18); however, an isolated area of relatively thick sandstone occurs in the Massanutten syncline (Yeakel, 1962, fig. 4). Yeakel (p. 1534) concludes that Bald Eagle, Juniata, and Tuscarora sediments were transported northwestward from various points along a source on the southeast. A similar depositional setting is postulated for the Bloomsburg Formation (Hoskins, 1961, p. 105).

Catskill-Pocono wedge.—The Devonian-Mississippian Catskill-Pocono wedge has a maximum thickness of more than 3.5 km near the center of the Pennsylvania salient (fig. 6). The wedge thins southwestward along structural strike, but the distal fringe extends into the Tennessee salient. The Devonian clastic sequence is generally less than 1.2 km thick in southwestern Virginia (Oliver and others, 1967, fig. 9; Colton, 1970, fig. 22); but, an anomalous thickness of 1.5 km is localized in the Kimberling syncline, evidently because of contemporaneous downwarp (Cooper, 1964, p. 101). In the Tennessee salient, the Devonian Chattanooga Shale and the Lower Mississippian Grainger Formation (shale and sandstone) constitute the southwestern fringe of the wedge, and the Grainger grades southwestward (fig. 6) into the Fort Payne Chert (Rodgers, 1953, p. 106; Englund, 1968, fig. 6). Thickness distribution in the northern part of the wedge is obscured by erosion; however, available data indicate northward thinning of Upper Devonian rocks in northeastern Pennsylvania (Glaeser, 1974, fig. 13).

The Devonian clastic sequence grades upward from black shale, through a sandstone-shale succession ("Chemung"), to red beds (Catskill). The various components prograde northwestward toward the craton and southwestward along structural strike (Oliver and others, 1967, figs. 11-1 and 12-1; McIver, 1970, fig. 7; Meckel, 1970, p. 54). The Catskill red-bed facies grades out southwestward across the Virginia recess (fig. 6). Proportion of sandstone decreases southwestward toward the Virginia recess and northwestward across strike (Oliver and others, 1967, fig. 9), and grain sizes in the sandstone units decrease southwestward from Pennsylvania (Dennison, 1963, p. 226, 1970, p. 65). Paleocurrent data from the most northerly outcrops in southeastern New York indicate transport from the east-northeast, but those from outcrops farther south indicate currents from the east-southeast (Burtner, 1963). Possibly the Devonian of southeastern New York (Catskill Mountains) includes sediment derived from the northeast (Quebec salient) as well as clastic-wedge sediment derived from a source on the southeast within the Pennsylvania salient (McCave, 1968, figs. 8 and 9; see also lithofacies maps by Woodrow and Fletcher, 1967, figs. 5-7).

For the Pocono Formation, paleocurrent data, distribution of pebble sizes, and sand-shale ratio mapping indicate transport to the west and north from a sediment source southeast of the center of the Pennsylvania salient (Pelletier, 1958, fig. 16, p. 1061). Sandstone petrography indicates a source terrain of sedimentary and low-grade metamorphic rocks (Pelletier, p. 1041). Regionally the Pocono thins southwestward and grades to finer grained rocks (Price Sandstone and Grainger Formation).

A transgressive interwedge carbonate, the Loyalhanna Limestone (fig. 6) and related carbonate units, overlaps the Catskill-Pocono wedge


FORM LINES OF DEPOSITIONAL STRIKE (MCCAVE, 1968)
SCALE
0 100 KM
NY
NJ
PA
MD
OH
WV
VA
KY
TN
Dr
MI
fp
0.2
0.3
0.5
1.2
3
d
e
f
g
ISOPACH LINES
DEVONIAN CLASTIC ROCKS
POCONO AND EQUIVALENT
LOCATION
AREA OF MAXIMUM PEBBLE SIZE OF POCONO
K KIMBERLING SYNCLINE
LIMITS OF STRATIGRAPHIC UNITS (LETTER WHERE UNIT PRESENT)
Dr DEVONIAN RED BEDS
fp FORT PAYNE CHERT
MI MISSISSIPPIAN LOYALHANNA LIMESTONE AND EQUIVALENT
SUMMARY OF PALEOCURRENT DATA
CATSKILL
DEVONIAN MARINE SANDSTONES
POCONO
END POINTS OF CROSS SECTIONS

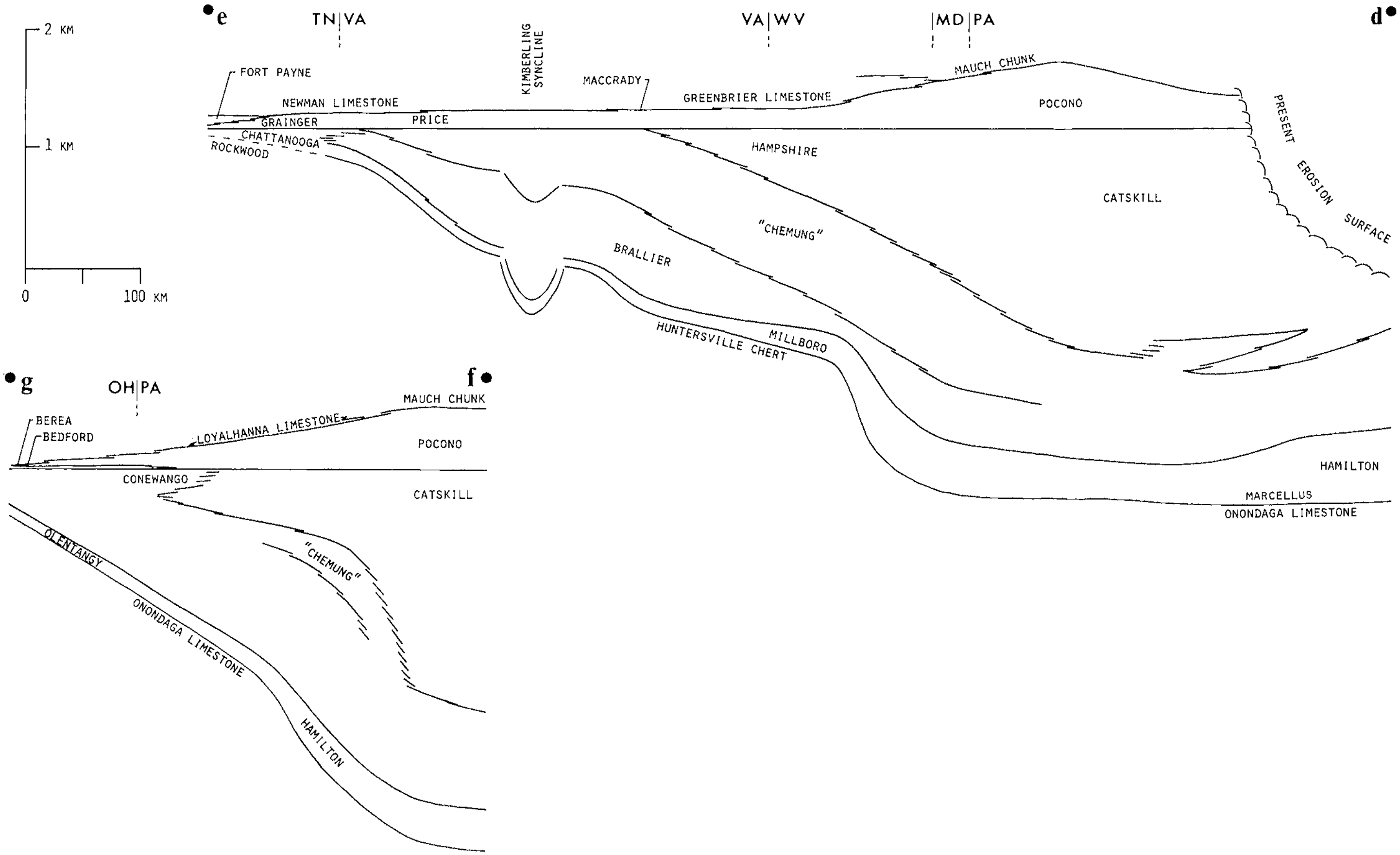
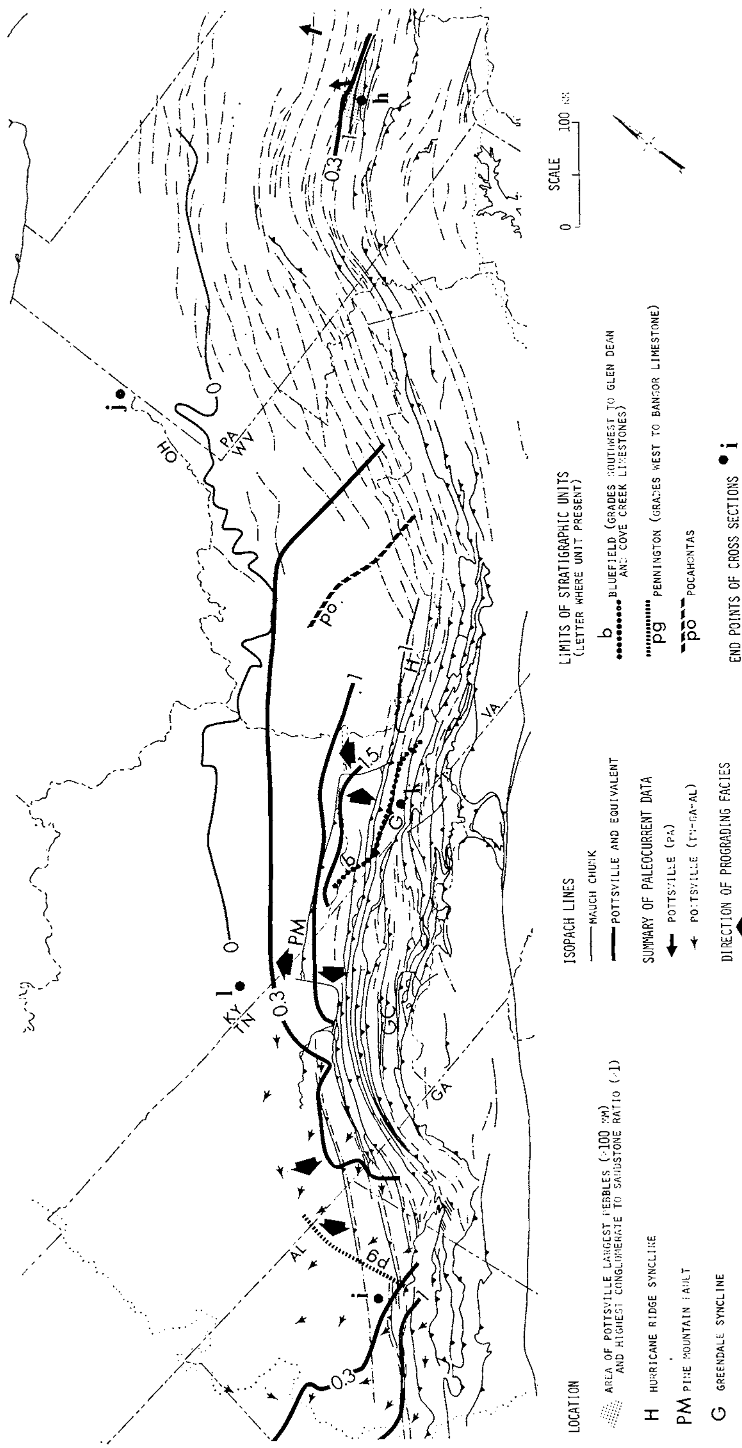


Fig. 6. Catskill-Pocono (Devonian-Mississippian) clastic wedge in Pennsylvania salient. Isopach maps of Devonian and Mississippian clastic units (contour values in km). Datum of cross sections is top of Devonian; dashed contact line on cross section indicates unconformity. Data from Butts (1940), Harris and Miller (1958), Pelletier (1958), Fettke (1961), Walker (1962), Ayrton (1963), Leeper (1963), Wagner (1963), Wolfe (1963), Cooper (1964), Oliver and others (1967), Englund (1968), Adams (1970), Colton (1970), Meckel (1970), Arkle (1974), Glaeser (1974), Dennison and Wheeler (1975). Structural base map from figure 1.


LOCATION
AREA OF POTTSVILLE LARGEST PEBBLES (>100 MM) AND HIGHEST CONGLOMERATE TO SANDSTONE RATIO (>1)
H HURRICANE RIDGE SYNCLINE
PM PINE MOUNTAIN FAULT
G GREENDALE SYNCLINE
GC GREASY COVE
ISOPACH LINES
MAUCH CHUNK
POTTSVILLE AND EQUIVALENT
SUMMARY OF PALEOCURRENT DATA
POTTSVILLE (PA)
POTTSVILLE (TN-GA-AL)
DIRECTION OF PROGRADING FACIES
LIMITS OF STRATIGRAPHIC UNITS (LETTER WHERE UNIT PRESENT)
b BLUEFIELD (GRADES SOUTHWEST TO GLEN DEAN AND COVE CREEK LIMESTONES)
pg PENNINGTON (GRADES WEST TO BANGOR LIMESTONE)
po POCAHONTAS
END POINTS OF CROSS SECTIONS
SCALE
0
100
OH
PA
WV
KY
TN
VA
GA
AL

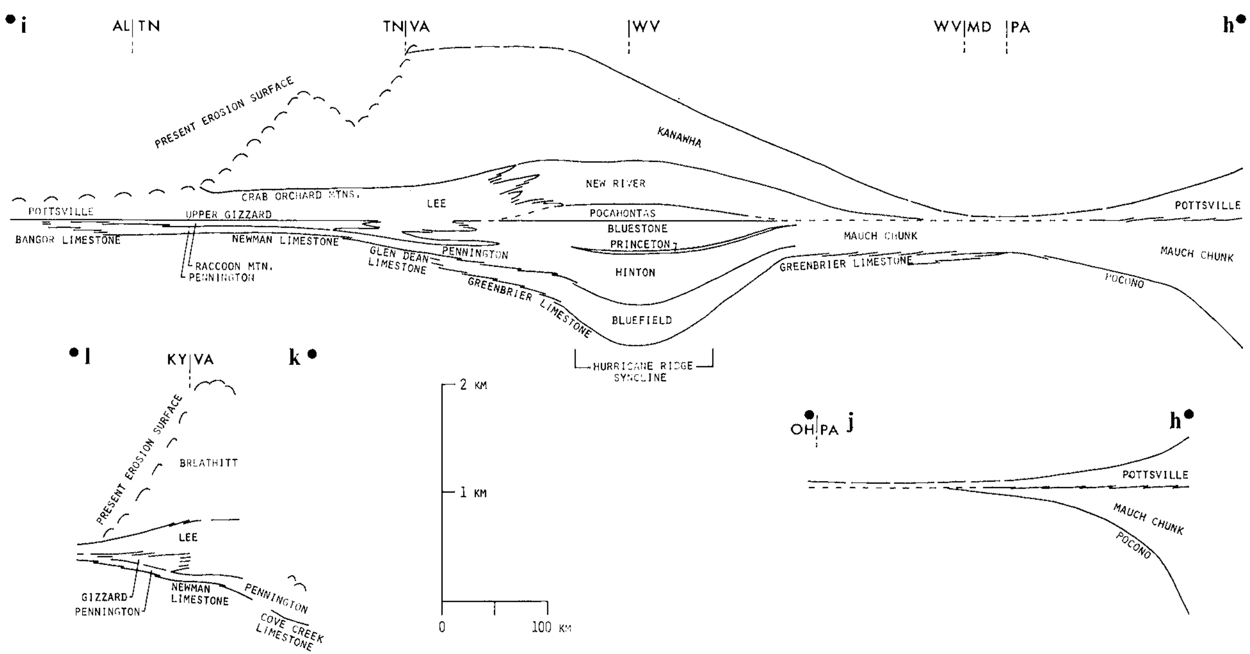


Fig. 7. Mauch Chunk-Pottsville (Mississippian-Pennsylvanian) clastic wedge in Pennsylvania salient and Pennington-Lee (Mississippian-Pennsylvanian) clastic wedge in Tennessee salient. Isopach maps of Mauch Chunk and Pottsville and equivalent clastic units (contour values in km). Datum of cross sections is top of Mississippian; dashed contact lines on cross sections indicate unconformities. Data from Butts (1940), Wanless (1946), Cooper (1948), Wilpolt and Marden (1949), Freeman (1951), Stearns (1954), Wilson, Jewell, and Luther (1956), Harris and Miller (1958), Neuman and Wilson (1960), Branson (1962a), Stearns and Mitchum (1962), Culbertson (1963a, b), Schlee (1963), Englund and DeLaney (1966), Thomas (1966, 1972b), Meckel (1967, 1970), Englund (1968, 1974), Ferm, Milici, and Eason (1972), Thomas and Drahovzal (1973), Arkle (1974), Hobday (1974), Miller (1974), Dennison and Wheeler (1975). Structural base map from figure 1; limit of Mesozoic-Cenozoic coastal plain strata shown by light dotted line.

in western Pennsylvania (Adams, 1970, p. 98). The limestone tongue pinches out eastward, and, in the central part of the salient, the Pocono is succeeded by the Mauch Chunk Formation of the next higher clastic wedge (Adams, 1970; Meckel, 1970, fig. 3). The Loyalhanna is the most northeasterly part of an extensive interwedge carbonate, which includes the Greenbrier Limestone of the Virginia recess and the Newman Limestone of the Tennessee salient (figs. 4, 6).

Mauch Chunk-Pottsville wedge.—The Mississippian Mauch Chunk Formation red-bed facies and Pennsylvanian conglomerate, sandstone, shale, and coal comprise the youngest wedge in the Pennsylvania salient. Maximum thickness of the wedge is in southeastern Pennsylvania (fig. 7), where the Mauch Chunk is more than 1 km thick (Meckel, 1970, fig. 3; Arkle, 1974, fig. 4). The Mauch Chunk thins northwestward across strike and extends southwestward to the Tennessee salient (fig. 7). A regionally anomalous thickness of more than 1 km is concentrated locally in the contemporaneous Hurricane Ridge syncline in the Virginia recess (Thomas, 1966, fig. 3). The Pennsylvanian Pottsville Group is more than 400 m thick in the southern anthracite field of southeastern Pennsylvania (Meckel, 1967, p. 228; Arkle, 1974, p. 23), and it thins northward and westward (fig. 7). Preserved post-Pottsville Pennsylvanian rocks are thickest in the southern anthracite field, and the equivalent section is thinner to the west and southwest (Meckel, 1970, fig. 3; Arkle, 1974, fig. 9).

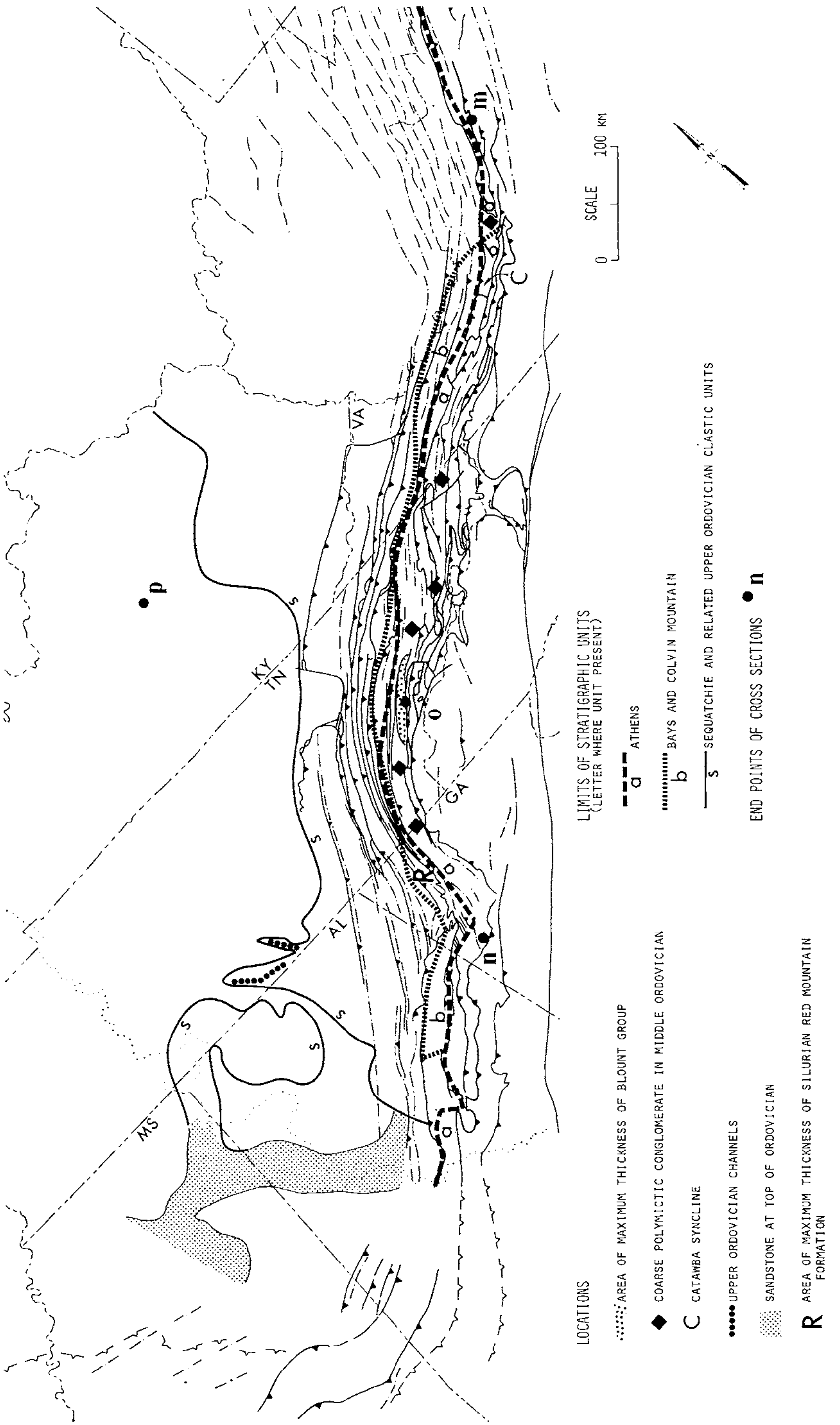
The highest conglomerate-sandstone ratio and the largest pebbles in the Pottsville are centered in southeastern Pennsylvania (Meckle, 1967, figs. 3 and 7). Paleocurrent data, pebble-size distribution, and isopach mapping indicate that Pottsville sediment was transported northward and northwestward from a source in southeastern Pennsylvania (Meckel, p. 240). Sandstone petrography demonstrates a source terrain of sedimentary and metamorphic rocks (Meckel, p. 233).

The Mauch Chunk progrades southwestward over the Mississippian (Loyalhanna-Greenbrier) limestone, and the lower clastic strata of the Mauch Chunk of Pennsylvania are time-correlative with part of the Greenbrier Limestone to the southwest (Cooper, 1948, p. 259). The lower part of the Mauch Chunk Group (partly calcareous shale of the Bluefield Formation) of southern West Virginia and adjacent Virginia grades southwestward (fig. 7) into the Greenbrier (Glen Dean) and Cove Creek Limestones in southwestern Virginia (Butts, 1940, p. 382; Wilpolt and Marden, 1949; Miller, 1974).

The southwestern limit of the Mauch Chunk-Pottsville wedge is indistinct in the Virginia recess, because Upper Mississippian and Pennsylvanian clastic rocks extend along the Appalachians from Pennsylvania to Alabama. It is suggested here that a lithologically similar and essentially coeval wedge (Pennington-Lee wedge) is centered on the Tennessee salient, and the oppositely directed wedges merge in the Virginia recess. Arkle (1974, p. 9) describes Pennsylvanian stratigraphy of West Virginia in terms of two basins, one exemplified by a northeastward-thickening sedimentary wedge (here called Mauch Chunk-Pottsville) and the other by a southward-thickening wedge (here called Pennington-Lee). The oppositely directed wedges overlap in West Virginia (Arkle, fig. 3).

Blount wedge.—The Blount wedge in the Tennessee salient records a Middle Ordovician orogenic phase (Rodgers, 1953, p. 94, 1971, p. 1165; Neuman, 1955, p. 171). Maximum thickness of the wedge, about 2.5 km, is in the southeastern outcrop belts near the center of the Tennessee salient (fig. 8). The wedge thins and grades to carbonate northwestward across strike (Rodgers, 1953, fig. 4), as well as northeastward and southwestward along strike (fig. 8). Gray shale and sandstone in the lower part of the wedge (Blount Group) is overlain by a red-bed facies (Moccasin and Bays Formations) which grades upward to quartzose sandstone (Bays Formation) in the southeastern belts (Rodgers, 1953, p. 81, 1971, p. 1165). Coarse polymictic conglomerates in the Middle Ordovician of the southeastern belts (fig. 8) were derived from local sources on the southeast; the clasts are sedimentary rocks from older Ordovician and Cambrian formations (Kellberg and Grant, 1956). The distal fringe of the wedge is deflected northeastward along the Appalachian structures, and the Bays Formation (sandstone) is locally very thick near its northeastern limit in the contemporaneous Catawba syncline (fig. 8) in the Virginia recess (Cooper, 1964, p. 93). In the interior belts of the Alabama recess, the Athens (black) Shale is evidently the distal fringe of the wedge. The Colvin Mountain Sandstone (equivalent to Bays) pinches out westward in the Alabama recess (Drahovzal and Neathery, 1971, p. 49).

In eastern Tennessee, the Bays Formation is unconformably overlain by Devonian strata, and the top of the Blount wedge is truncated. The hiatus diminishes to southwest, northwest, and northeast, and the Upper Ordovician (Sequatchie) and Silurian (Red Mountain, Rockwood) succession includes red sandstone, siltstone, and shale. The western (cratonward) limit of Sequatchie is irregular, and possible correlatives of the Upper Ordovician clastic sequence extend to channel-filling sandstones in south-central Tennessee (Wilson, 1949, p. 223) and a sandstone at the top of the Ordovician limestone in northeastern Mississippi (Thomas, 1972a, p. 92). Silurian clastic rocks grade westward to a carbonate facies and are truncated eastward; maximum thickness is more than 300 m in northwest Georgia (Chowns, 1972, fig. 1). The Sequatchie and Red Mountain (Rockwood) are commonly interpreted to be the distal part of the Martinsburg-Shawangunk wedge (Rodgers, 1971, p. 1164; Chowns, 1972, p. 13; Dennison and Wheeler, 1975, p. 47); if so, the distal fringe of that wedge extends entirely across the Tennessee salient. Alternatively, the clastic rocks of the Sequatchie-Red Mountain (Rockwood) may be the distal upper part of the Blount wedge; however, their proximal (eastern) equivalents are not preserved. The Silurian Clinch Sandstone (distal fringe of Martinsburg-Shawangunk wedge) grades out southwestward into the Rockwood Formation across northeastern Tennessee (Rodgers, 1953, p. 101). The Upper Silurian-Lower Devonian Hancock Limestone (Rodgers, 1953, p. 103) is an interwedge carbonate unit above the Blount


LOCATIONS
AREA OF MAXIMUM THICKNESS OF BLOUNT GROUP
COARSE POLYMICTIC CONGLOMERATE IN MIDDLE ORDOVICIAN
C CATAWBA SYNCLINE
UPPER ORDOVICIAN CHANNELS
SANDSTONE AT TOP OF ORDOVICIAN
R AREA OF MAXIMUM THICKNESS OF SILURIAN RED MOUNTAIN FORMATION
LIMITS OF STRATIGRAPHIC UNITS (LETTER WHERE UNIT PRESENT)
a ATHENS
b BAYS AND COLVIN MOUNTAIN
s SEQUATCHIE AND RELATED UPPER ORDOVICIAN CLASTIC UNITS
END POINTS OF CROSS SECTIONS n
SCALE
0 100 KM
VA
KY
TN
GA
AL
MS

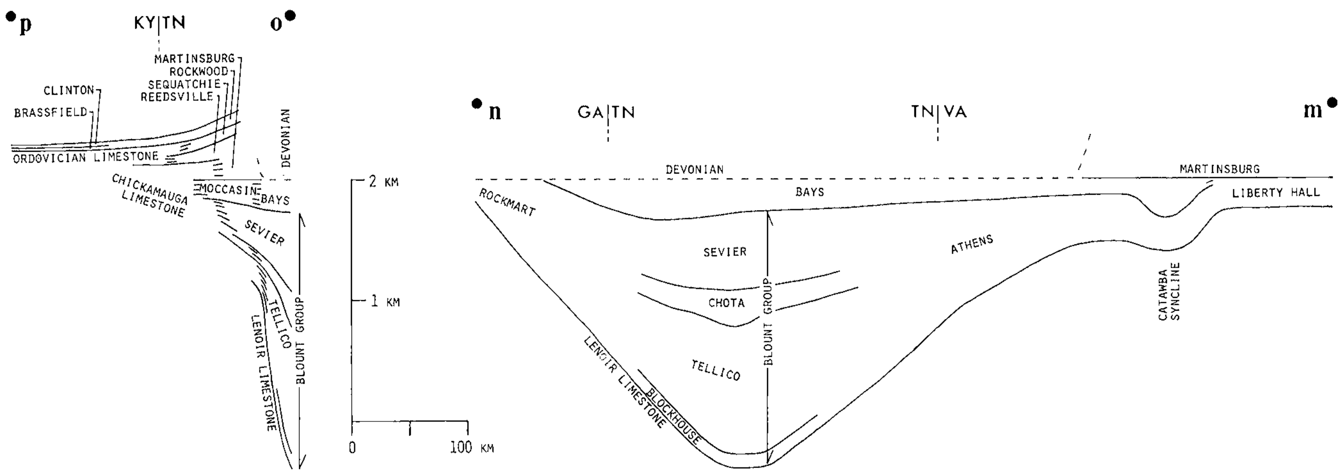


Fig. 8. Blount (Ordovician-Silurian) clastic wedge in Tennessee salient. Datum of cross sections is top of Bays Formation and equivalent; dashed contact line on cross sections indicates unconformity. Data from Butts (1940), Cooper and Cooper (1946), Wilson (1949), Freeman (1951), Munyan (1951), Rodgers (1953), Cooper (1955, 1964), Neuman (1955), Kellberg and Grant (1956), Allen and Lester (1957), Neuman and Wilson (1960), Tillman (1963), Englund (1968), Cressler (1970), Kidd and Copeland (1971), Drahovzal and Neathery (1971), Chowns (1972), Thomas (1972a), Thomas and Drahovzal (1973), Dennison and Wheeler (1975), Kidd (1975). Structural base map from figure 1; limit of Mesozoic-Cenozoic Gulf coastal plain strata shown by light dotted line.

wedge, but the Silurian-Devonian section is thin and incomplete in the Tennessee salient.

Pennington-Lee wedge.—The Pennington-Lee wedge in the Tennessee salient includes shale and sandstone of the Upper Mississippian Pennington Formation and laterally discontinuous units of Pennsylvanian (Lee, Kanawha, and equivalent) conglomerate, sandstone, shale, and coal (fig. 7). Rocks of this wedge are preserved in the frontal structures in Tennessee, but the distribution and center of the wedge are obscured by erosion on the present surface. Northeastward thinning of the wedge away from the Tennessee salient is well documented (fig. 7). Because of deeper erosion, the southwestward extent of the wedge is not shown by preserved thicknesses; paleocurrent data and facies distribution demonstrate southwestward and westward prograding from the Tennessee salient into the Alabama recess. A possible remnant of the wedge in the interior structures in the Tennessee salient is the Mississippian Greasy Cove Formation (of Neuman and Wilson, 1960) which consists of limestone, sandstone, and shale. The Greasy Cove clastic rocks are older than the basal Pennington of the frontal structures.

The Pennington Formation is about 300 m thick in the southeastern part of the Pine Mountain fault block in southwestern Virginia, and it has a maximum thickness of more than 400 m (the top is eroded) in the Greendale syncline southeast of the Pine Mountain block (Cooper, 1948, p. 262). The Pennsylvanian part of the wedge is thickest in the area of the Pine Mountain fault block (fig. 7); northeastward thinning across the Virginia recess reflects overstep and thinning out of the older (Lee-equivalent) Pocahontas and New River Formations (Arkle, 1974, fig. 5; Dennison and Wheeler, 1975, p. 121).

On the Pine Mountain fault block, conglomeratic quartzose sandstone beds, characteristic of the lower part of the Lee Formation, intertongue with marine fine clastic rocks of the upper Pennington (Englund, 1968, p. 13). The lower sandstone tongues pinch out both northeast and southwest along structural strike (fig. 7) as well as across strike to the northwest (Englund, 1964, p. B33, 1968, fig. 11; Englund and DeLaney, 1966, p. D49). The stratigraphically lowest Lee sandstone beds are near the center of the Pine Mountain fault block, and the basal Lee beds prograde southwestward, northwestward, and northeastward from that point (fig. 7). On a more regional scale, the clastic Pennington Formation progrades southwestward over the Newman (Bangor) Limestone in Tennessee (Ferm, Milici, and Eason, 1972, fig. 3), and massive sandstones (Lee of northern Tennessee) prograde southwestward over the Pennington. In northeastern Alabama, the Pennington clastic facies grades southwestward into the upper part of the Mississippian Bangor Limestone (Thomas, 1972b, p. 89, 1974, p. 205). Pottsville-equivalent massive sandstones extend southwestward beyond the limit of the Pennington and prograde over the Bangor in northern Alabama (Thomas, 1974, p. 205), and cross bedding in the post-Bangor sandstones indicates generally westward or southwestward sediment transport in northeastern Alabama

(Tanner, 1959, p. 224; Schlee, 1963, p. 1446; Chen and Goodell, 1964, p. 70; Metzger, 1965, p. 27).

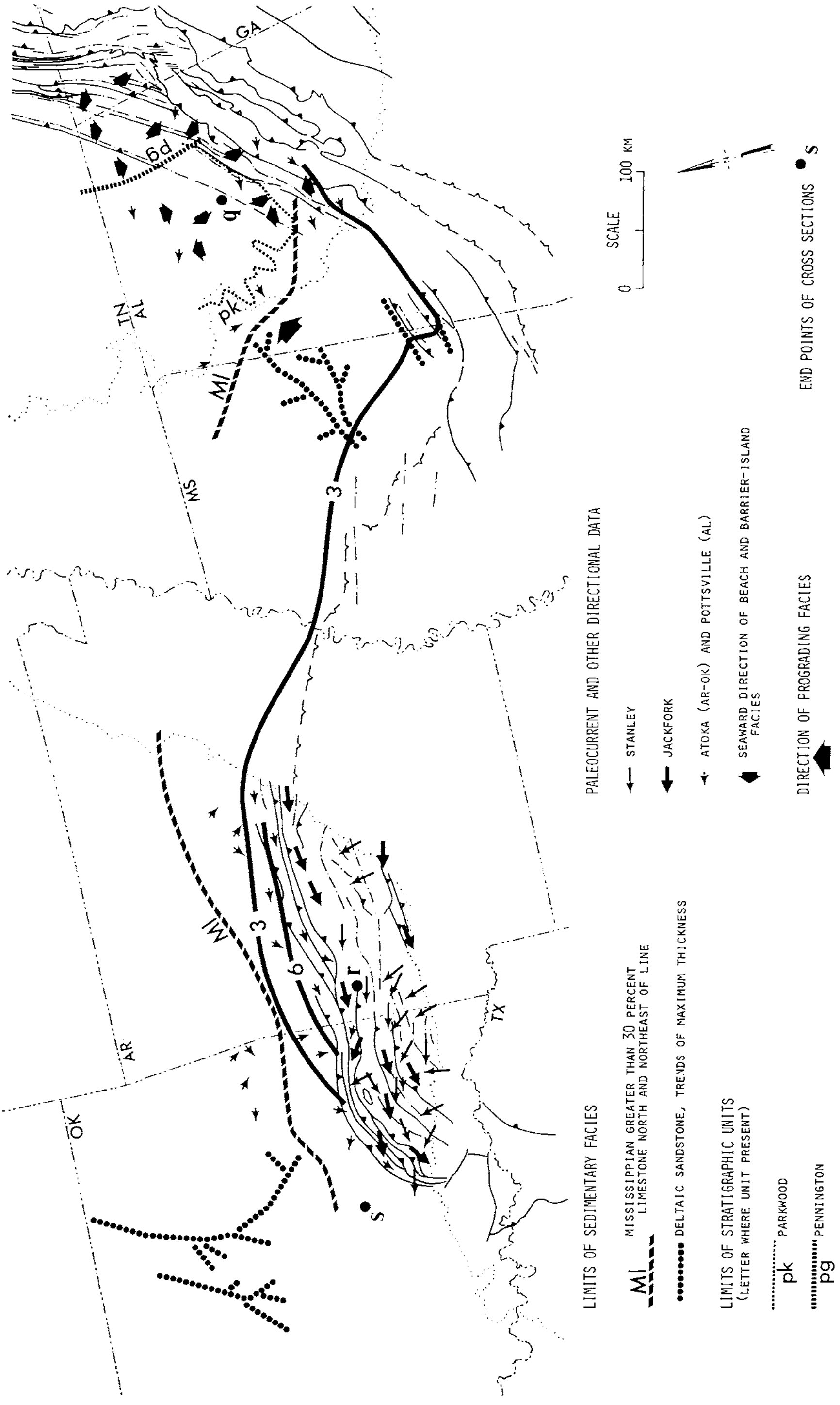
Distribution of thickness and direction of facies progradation (fig. 7) suggest that the depositional center of the Pennington-Lee wedge was southeast of the Pine Mountain fault block in the Tennessee salient. Sediment supply from southeast of the salient is indicated by facies distribution (for example, the southwestward prograding clastic sediments in Alabama).

Ouachita wedge.—The Mississippian-Pennsylvanian clastic wedge in the Ouachita Mountains is more than 9 km thick (fig. 9). The succession is interpreted to be a deep-water flysch facies (Cline, 1960, p. 100, 1970; Morris, 1974, p. 124) which grades upward from shale (Stanley) into a sandstone-shale succession (Jackfork-Atoka).

North and northwest of the Ouachitas in Arkansas and Oklahoma, the Mississippian-Pennsylvanian succession is much thinner and consists of shallow-marine and deltaic sandstones, shales, and limestones (Maher and Lantz, 1953; Laudon, 1959; Scull, Glover, and Planalp, 1959, p. 167; Ogren, 1968; Visher, Saitta B., and Phares, 1971; Glick, 1973; Zachry and Haley, 1973). In Oklahoma, the Pennsylvanian includes southward prograding deltaic sandstones (Scull, Glover, and Planalp, 1959, p. 167; Visher, Saitta B., and Phares, 1971, p. 1212). Sediment thickness distribution indicates active down-to-basin faults contemporaneous with Atoka deposition at the northern edge of the Ouachita trough (Koinm and Dickey, 1967; Buchanan and Johnson, 1968; Haley and Hendricks, 1968), and erratic boulders in the Johns Valley Shale were supplied from similar scarps (Shideler, 1970, p. 803).

The wedge thins northeastward to less than 3 km in the Black Warrior basin of Mississippi and Alabama (fig. 9), where the succession consists of deltaic and shallow-marine strata (Thomas, 1974, p. 200). Distribution of sandstone units indicates a northeastwardly prograding delta system, and the Mississippian part of the clastic wedge (Floyd-Parkwood) grades northeastward into a shelf carbonate facies (Bangor Limestone) in Alabama (Thomas, 1972a, p. 98, 1972b, p. 105, 1974, p. 194). The strike of the facies boundary is nearly perpendicular to Appalachian structural strike in Alabama, but it is approximately parallel with the more distant Ouachita structural front (fig. 9). Distribution of the clastic facies indicates a sediment supply from the southwest. The northeastward prograding Mississippian clastic facies extends farthest northeast along contemporaneous Appalachian synclines in Alabama (Thomas, 1974, p. 203).

The Pennsylvanian part of the Ouachita wedge (Pottsville of Alabama) progrades northeastward over the stratigraphically highest part of the Mississippian carbonate facies in the Alabama recess. Evidently the northeastward prograding clastic sediments (Ouachita wedge) merged with southwestward prograding clastic sediments (Pennington-Lee wedge) above the Bangor Limestone in northern Alabama; however, the bound-


LIMITS OF SEDIMENTARY FACIES
MI
MISSISSIPPIAN GREATER THAN 30 PERCENT LIMESTONE NORTH AND NORTHEAST OF LINE
DELTAIC SANDSTONE, TRENDS OF MAXIMUM THICKNESS
LIMITS OF STRATIGRAPHIC UNITS (LETTER WHERE UNIT PRESENT)
pk PARKWOOD
pg PENNINGTON
PALEOCURRENT AND OTHER DIRECTIONAL DATA
STANLEY
JACKFORK
ATOKA (AR-OK) AND POTTSVILLE (AL)
SEAWARD DIRECTION OF BEACH AND BARRIER-ISLAND FACIES
DIRECTION OF PROGRADING FACIES
SCALE
0
100 KM
END POINTS OF CROSS SECTIONS
s
OK
AR
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MS
TN
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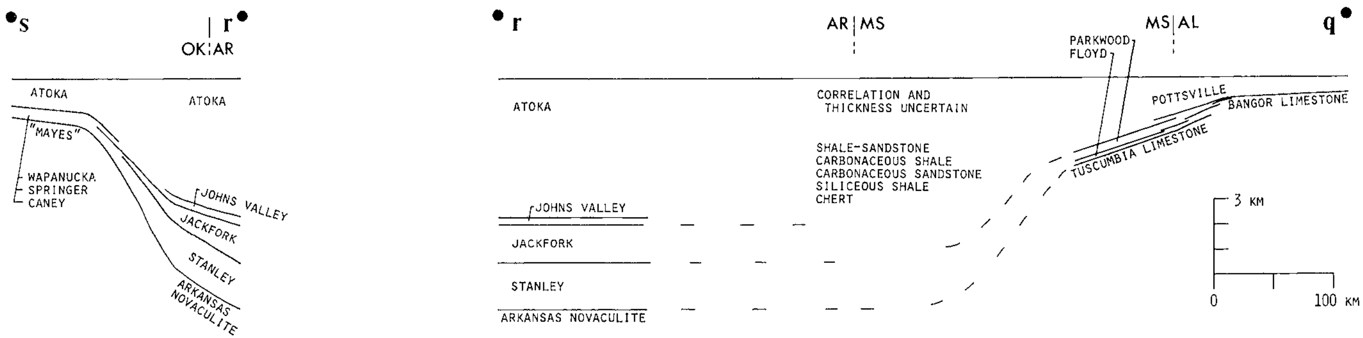


Fig. 9. Mississippian-Pennsylvanian clastic wedge in Ouachita salient. Composite isopach map of clastic wedge (contour values in km). Datum of cross sections is top of Atoka. Data from Maher and Lantz (1953), Caplan (1957), Reinemund and Danilchik (1957), Frezon and Glick (1959), Laudon (1959), Scull, Glover, and Planalp (1959), Cline (1960), Branson (1962b), Frezon (1962), Goldstein and Hendricks (1962), Schlee (1963), Haley and Frezon (1965), Stone (1966), Ogren (1968), Visher, Saitta B., and Phares (1971), Thomas (1972a, b, 1974), Thomas and Drahovzal (1973), Hobday (1974), Morris (1974). Structural base map from figure 1; limit of Mesozoic-Cenozoic Gulf coastal plain strata shown by light dotted line.

ary between the wedges is unidentified (Thomas, 1972b, p. 90, 1974, p. 205). On the basis of geometry of beach and barrier-island sandstones of the basal Pottsville, Hobday (1974, p. 223) concludes that two sediment supply systems (from the northeast and from the south) merged in north-central Alabama, "expelling the sea toward the northwest." Cross bedding in the Pottsville of northeastern Alabama indicates westward or southwestward transport (figs. 7, 9), but in northwestern Alabama cross-bed orientation is more diverse (Schlee, 1963, pl. 1; Metzger, 1965, p. 27). Compositional variation in Pottsville sandstones in central Alabama indicates a southerly (Ouachita) sediment source (Davis and Ehrlich, 1974, p. 177). Populations of lithic grains in sandstones of the Alabama Pottsville and the Arkansas Jackfork-Atoka are essentially indistinguishable, and derivation from the same source rocks is indicated (Graham, Ingersoll, and Dickinson, 1976, p. 622).

The great volume of Mississippian-Pennsylvanian clastic sediment in the Ouachita wedge has been interpreted to indicate a geologically complex, tectonically active sediment source south or southeast of the trough (Miser, 1921; Miser and Purdue, 1929, p. 134; Bokman, 1953, p. 168; King, *in* Flawn and others, 1961, p. 184; Goldstein and Hendricks, 1962, p. 421; Hill, 1966, p. 120; Johnson, 1966, p. 156; Klein, 1966, p. 316; Walthall, 1967, p. 323; Cline, 1970, p. 100; summary by Morris, 1974, p. 130; Niem, 1976, p. 644). Other petrographic data indicate a supply of quartzose sand from the craton on the north (Klein, 1966, p. 316; Morris, 1971, p. 398); however, that supply did not become active until after Stanley deposition (Morris, 1974, p. 138). Volcanism on the south is indicated by the tuff beds (for example, Hatton Tuff) in the Ouachita sequence (Johnson, 1966, p. 156; Niem, 1976, p. 644). Paleocurrent data indicate northwestward transport during deposition of the lower (Stanley) part of the wedge (Johnson, 1966, p. 150; Morris, 1974, p. 133). During deposition of the higher (Jackfork-Atoka) part of the wedge, transport was predominantly westward (axial) along the trough (Briggs and Cline, 1967, p. 991; Cline, 1970, p. 93; Morris, 1971, p. 399, 1974, p. 133). A comprehensive interpretation suggests that sediment was introduced into the Ouachita trough from both south and north and was transported westward along the axis (Klein, 1966, p. 323; Cline, 1970, p. 100; Thomas, 1976, p. 338). The regional tectonic setting suggests that the sediment source south or southeast of the Ouachita trough also supplied the northeastward prograding clastic sediments of Mississippi and Alabama (Thomas, 1972b, p. 106, 1974, p. 202, 1976, p. 337, 1977, p. 18). The dominant supply of sediment from the south into the Ouachita trough, Black Warrior basin, and southwestern Appalachians persisted through the Late Mississippian (Stanley and Floyd-Parkwood) and into the Pennsylvanian. During Early Pennsylvanian, the supply of sediment became more complex and included quartzose sand from the craton north of the Ouachita trough. Possibly parts of the Pennington-Lee wedge prograded from the Tennessee salient westward to the eastern end of the Ouachita trough during Pennsylvanian (compare Morris, 1971, 1974; Graham, Dickinson, and Ingersoll, 1975).

Southwestward from the Ouachita salient, elements of the clastic wedge are recognized in the subsurface (Flawn, *in* Flawn and others, 1961, pl. 2; Goldstein and Hendricks, 1962, p. 423), but insufficient data are available to define the thickness of the wedge. However, the frontal zone is more narrow southwestward from the Ouachita salient (Flawn, *in* Flawn and others, 1961, p. 166), suggesting that the greatest volume of clastic sediment originally accumulated in the salient (King, *in* Flawn and others, 1961, p. 184).

Marathon wedge.—The Mississippian-Pennsylvanian succession in the Marathon salient (fig. 10) is a clastic wedge about 3.5 km thick (Flawn, *in* Flawn and others, 1961, p. 53; McBride, 1970, fig. 2). The lower part of the wedge (Tesnus-Dimple-Haymond) is a flysch facies, and the upper part (Gaptank) is interpreted to include a molasse facies (reviewed by Flawn, *in* Flawn and others, 1961, p. 53; McBride, 1970, p. 69). Thickness distribution and paleocurrent data indicate a supply of clastic sediment (Tesnus and Haymond) from a marginal sediment source (Llanoria) southeast of the salient (King, 1937, p. 135; McBride, 1970, p. 69). The Dimple Limestone is a carbonate flysch derived from a shelf on the craton to the northwest and from a smaller shelf to the south (Thomson and Thomasson, 1969, p. 84; McBride, 1970, p. 69). Possible extent of the clastic wedge eastward toward the Texas recess is obscured by the uncertain stratigraphy and structure of the Devils River uplift (Flawn, *in* Flawn and others, 1961, p. 167; Nicholas and Rozendal, 1975, p. 201); however, Ouachita frontal zone rocks are not recognized by Flawn (p. 167) along the Devils River uplift. As in the Ouachita salient, the frontal zone is relatively wide in the Marathon salient, and King (*in* Flawn and others, 1961, p. 184) infers original accumulation of a greater volume of clastic sediment in the salients.

Taconic wedge.—Sediments of a clastic wedge in the Quebec salient are preserved south of the Adirondack dome and along the St. Lawrence River (exogeosyncline of Cady, 1969, pl. 1; Rodgers, 1971, p. 1148; St. Julien and Hubert, 1975, p. 342). In New York, the carbonate-bank facies is overlain by westward prograding Middle and Upper Ordovician shale and sandstone (Rodgers, 1971, p. 1149), the Taconic wedge. At the edge of the carbonate bank, the oldest part of the clastic wedge is of Middle Ordovician age (Rodgers, p. 1148), and younger clastic strata extend farther west. Uplift of an orogenic sediment source on the east resulted in supply of sediment to the prograding clastic wedge and subsequently in gravity-slide emplacement of the Taconic allochthon (Zen, 1968, p. 136; Bird and Dewey, 1970, p. 1046; Rodgers, 1971, p. 1149). Emplacement of the early Taconic slices during deposition of the clastic-wedge sediments was accompanied by deposition of wildflysch conglomerate in front of the advancing slices (Zen, 1967, p. 35). The related tectonic events constitute the Taconic orogeny (Rodgers, 1971).

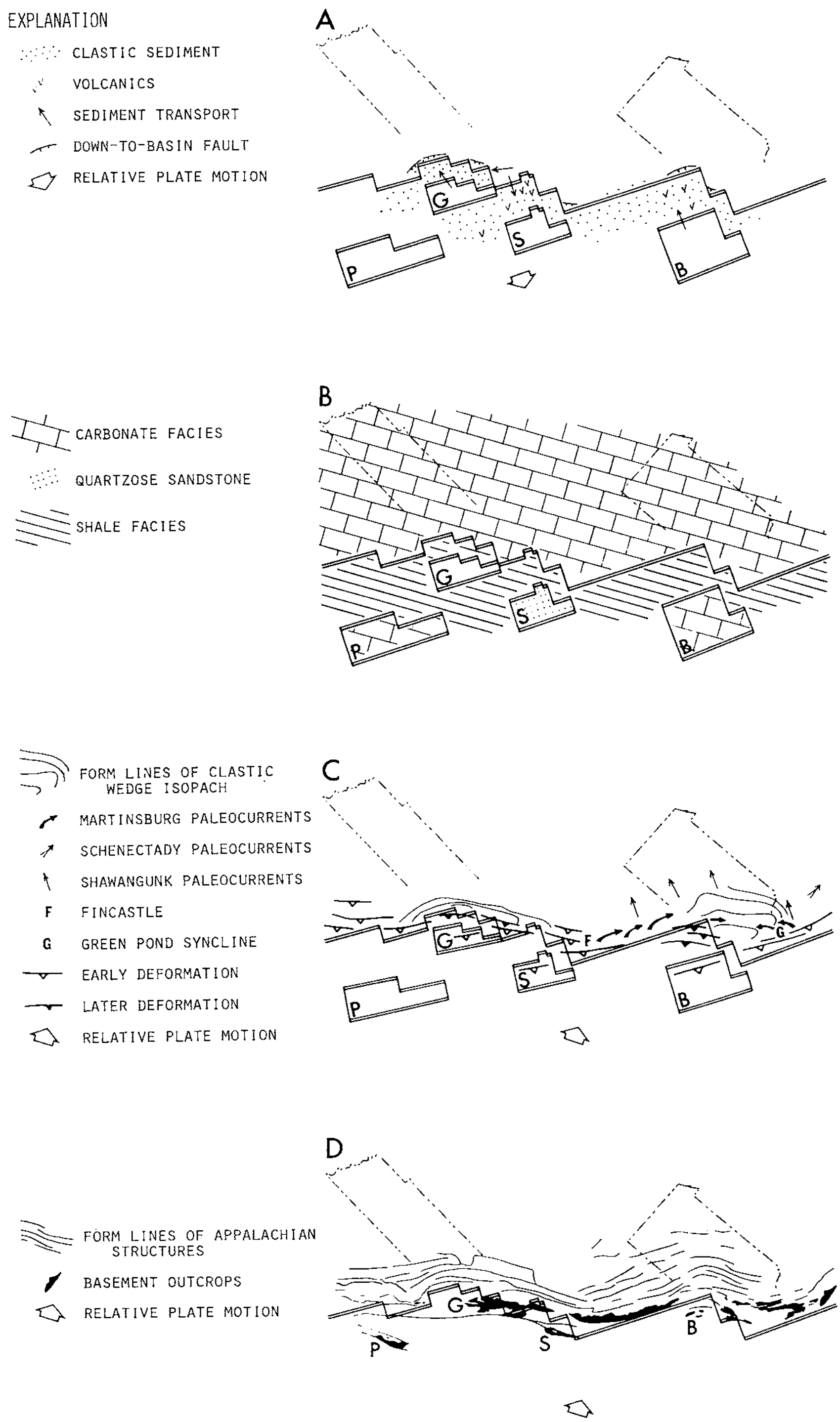
Fig. 10. Mississippian-Pennsylvanian clastic wedge in Marathon salient. Isopach map (from Galley, 1958, fig. 20) of pre-Desmoinesian Pennsylvanian clastic rocks (contour values in km). Schematic cross section modified from McBride (1970, fig. 6) is perpendicular to axis of trough in Marathon region. Data from Galley (1958), Thomson and Thomasson (1969), McBride (1970). Structural base map from figure 1; limit of Mesozoic-Cenozoic Gulf coastal plain strata shown by light dotted line.

Newfoundland.—The early Paleozoic geology of western Newfoundland is comparable to that of the Quebec salient (Rodgers and Neale, 1963; Bird and Dewey, 1970; Rodgers, 1971, p. 1161). The inception of a prograding clastic wedge above the carbonate-bank facies in Newfoundland and the emplacement of gravity slides within that clastic sequence are earlier than comparable events in the Taconic allochthon (Bird and Dewey, 1970, p. 1049; Rodgers, 1971, p. 1161-1162).

DISCUSSION AND INTERPRETATION

Distributions of thickness and facies of Late Precambrian and Paleozoic rocks bear specific relation to the shape of Appalachian-Ouachita salients and recesses. This relation indicates that the shape of the salients and recesses was inherited from a structural framework that persisted from the inception of Appalachian-Ouachita geosynclinal sedimentation. Three fundamental genetic types of sequences within the total stratigraphic succession reflect the evolution of the continental margin and sediment supply, but each type exhibits specific adaptation to the shape of the tectonic framework. Stratigraphic distribution patterns, as well as the curves of the Appalachian-Ouachita structural system, are consistent with the interpretation that the original tectonic framework was defined by an orthogonally zigzag continental margin. The shape of the continental margin may be interpreted to be the result of transform faults along a Late Precambrian rift. Transform-rift junctions define *reentrants* cut out of the continental margin (crustal margin concave oceanward) and *promontories* of the continent (crustal margin convex oceanward). Reentrants in the continental margin have evolved as structural salients, and promontories of the continent have become sites of structural recesses (fig. 11).

Initial deposits in the opening rift constitute the Late Precambrian-Early Cambrian clastic-volcanic sequence (fig. 11A). Abrupt thickness variations (particularly along the cratonward pinch-out of Late Precambrian rocks) reflect fault-bounded depositional basins along the rift and transforms. Greatest thicknesses evidently accumulated inside the angular reentrants in the continental margin. Composition of the volcanic rocks indicates their association with rifting (Rankin, 1975b, p. 313). Some local variations in sediment thickness may be the results of down-to-basin fault collapse. Late Precambrian sediments were deposited rapidly in the steep-sided basins, although several alternatives have been proposed for water depth and specific environments (compare Power and Forrest, 1973, p. 704; and Hadley, 1975). Much of the clastic sediment evidently was supplied from local sources near the continental margin; for example, the northeastern source of much of the Ocoee sediment in the Tennessee reentrant (later salient) may have been on the Virginia promontory (later recess). Other clastic sediments (lower Ocoee, Hadley, 1970, p. 249; parts of Wissahickon, Hopson, 1964, p. 131) evidently were derived from the east from extracratonic sources; however, composition of these sediments indicates a dominant supply from continental crustal


EXPLANATION
CLASTIC SEDIMENT
VOLCANICS
SEDIMENT TRANSPORT
DOWN-TO-BASIN FAULT
RELATIVE PLATE MOTION
A
G
S
P
B
CARBONATE FACIES
QUARTZOSE SANDSTONE
SHALE FACIES
B
G
S
P
B
FORM LINES OF CLASTIC WEDGE ISOPACH
MARTINSBURG PALEOCURRENTS
SCHENECTADY PALEOCURRENTS
SHAWANGUNK PALEOCURRENTS
F FINCASTLE
G GREEN POND SYNCLINE
EARLY DEFORMATION
LATER DEFORMATION
RELATIVE PLATE MOTION
C
G
F
G
S
P
B
FORM LINES OF APPALACHIAN STRUCTURES
BASEMENT OUTCROPS
RELATIVE PLATE MOTION
D
G
P
S
B

rocks. These sources may have been fragments of continental crust that were rifted from the edge of the continent but were left as microcontinents within the opening basin (for example, Grenville-age basement of the Sauratown Mountains and of the Baltimore domes).

The sharply angular and topographically rough continental margin was smoothed by Late Precambrian erosion and deposition. Although the outlines of reentrants and promontories persisted, the extensive quartzose sandstone of the Early Cambrian reflects general planation and establishment of a beach environment near the rifted edge of continental crust. As the beach environment transgressed onto the craton, a carbonate bank evolved on continental crust (fig. 11B). The trace of the bank edge may depart from the rifted continental margin locally because of fault collapse of the margin and possibly because of the shape of the earlier sedimentary fill. The bank facies changes abruptly into the deep-water shale facies. Outwash of clastic sediment over the bank supplied mud to the marginal basin, and collapse of the bank edge introduced carbonate boulders into the basin sediments. Quartz sand from the cratonic beach facies occasionally was transported across the bank into the deep basin. Intermittent volcanic activity in the basin is indicated by volcanic and volcaniclastic materials in the Piedmont.

Isolated small shallow-water carbonate banks formed on some of the microcontinents within the basin. On the Baltimore domes, the Setters-Cockeysville sandstone-carbonate sequence overlies basement rocks. The Setters-Cockeysville sequence, which has problematic relationship to the main cratonic carbonate facies, probably formed on an isolated microcontinent bank that never was connected to the similar (but not necessarily coeval) bank on the craton (fig. 11B). Other possible microcontinent banks include the postulated Texarkana platform south of the Ouachitas (Thomas, 1976, p. 337), Pine Mountain of Georgia, and Sauratown Mountains (but no carbonate rocks are preserved on the latter). The Murphy Marble (fig. 3) is also a shallow-water carbonate facies (Power and Forrest, 1973, p. 707); however, the Murphy is stratigraphically above a thick sequence of Late Precambrian clastic rocks and is not associated with an isolated fragment of continental basement.

The carbonate bank framework persisted at least through Cambrian and Early Ordovician. The bank environment then was disrupted by the influx of clastic wedge sediments from marginal sources. Each clastic

←Fig. 11. Sequential diagram of evolution of salients and recesses from reentrants and promontories in the continental margin. Outline of rift (double heavy line) and transforms (single heavy line) from figure 1; map shows area of Pennsylvania and Tennessee salients (outlines of states of Pennsylvania and Tennessee for geographic reference). Microcontinents: B = Baltimore domes, G = eastern Great Smoky Mountains, P = Pine Mountain, S = Sauratown Mountains.

A. Late Precambrian: opening rift, deposition of clastic-volcanic sequence.

B. Cambrian-Ordovician: deposition of carbonate-bank facies and deep-water shale facies; deposition of carbonate facies (and quartzose sandstone) on microcontinents.

C. Ordovician-Silurian: initial compression of continental margin and closing of marginal basin; deposition of clastic wedges. Details of evolution of Martinsburg-Shawangunk wedge in Pennsylvania salient and outline of Blount wedge in Tennessee salient.

D. Late Paleozoic: final compressive deformation (outline of continental margin shown as prior to deformation for comparison with other maps).

wedge is a discrete large-scale sedimentary unit, the center of which coincides geographically with a structural salient. The time of initial clastic wedge deposition varies along the system, and the timing of clastic pulses appears unique for each salient. However, the general aspects of composition and depositional history of each wedge are similar.

Compositional variations, grain-size distributions, paleocurrent data, and facies and thickness distributions indicate that wedge sediment was derived from marginal sources and transported onto the old carbonate bank and toward the craton. This indicates an "inversion of relief" from the configuration of the earlier carbonate bank and associated deep marginal basin (Rodgers, 1971, p. 1147). Composition of much of the wedge sediment indicates a provenance of sedimentary and low-grade metamorphic rocks. Some of the sediment was eroded from metamorphic and plutonic rocks. Intermittent volcanism characterized some source areas. Compressive deformation and uplift of older sedimentary rocks are implied, and perhaps a rising mobile core of deformed and partly metamorphosed older sedimentary rocks supplied clastic-wedge sediment. The older basement rocks of the microcontinents also were incorporated into some of the sediment sources.

Maximum thickness of each clastic wedge coincides approximately with a structural salient (earlier reentrant). Evidently differentially greater subsidence in the reentrants (later salients) resulted in concentration of clastic-wedge sediments, while the promontories (later recesses) stood as shallow platforms marked by thinner sedimentary accumulations. Sediment transport data indicate point sources along the system. Radial distribution from near the centers of the reentrants seems to be a common pattern; however, uplifts within promontories supplied longitudinally transported sediment into some reentrants. On a local scale, contemporaneous deformation resulted in thick sediment accumulations and in longitudinal sediment transport along single synclines, while contemporaneous uplifts locally supplied clastic sediment. The bulk of clastic-wedge sediment was derived from marginal tectonic source lands; however, quartzose sediment from the craton is included in the cratonward fringes of some wedges.

The interpretation shown in figure 11C is designed as a general model for clastic wedge deposition, although it specifically shows evolution of the Martinsburg-Shawangunk wedge. Modifications in timing and sedimentological details enable application of this model to other Appalachian–Ouachita clastic wedges. Maximum thickness of the Martinsburg-Shawangunk wedge is within the Pennsylvania reentrant on continental crust, and the wedge overlies the carbonate-bank facies (fig. 11C). During deposition of the Martinsburg, dominant longitudinal paleocurrents flowed northeastward in Pennsylvania and farther south but flowed southwestward in New York and New Jersey (McBride, 1962, p. 83). Evidently the bottom of the depositional basin was in eastern Pennsylvania. Petrographic data indicate two different sources of Martinsburg sediment: a provenance of sedimentary rocks from Pennsylvania southward, and a provenance of metamorphic and plutonic rocks in the north (McBride, p. 66). These sources and dispersal systems may reflect early compression and uplift on the promontories in Virginia and New York (fig. 11C). Pre-Martinsburg uplift of Blue Ridge basement and the overlying sedimentary succession in Virginia is indicated by clasts in the Middle Ordovician Fincastle Conglomerate (Lowry, 1974, p. 584), and that uplift may have later supplied northeastward-transported Martinsburg sediment. Silurian conglomerate rests unconformably on basement rocks along part of the Green Pond syncline in New Jersey (Finks, 1968, p. 117), and possibly basement uplifts in that area supplied southwestward-transported Martinsburg sediment. Later, during deposition of the Shawangunk Conglomerate, paleocurrents flowed predominantly northwestward from the Pennsylvania reentrant (fig. 11C). Apparently the later provenance was laterally more extensive along the basin margin, and radial transverse transport of clastic sediment from the southeast was dominant. The change from important longitudinal transport (Martinsburg) to generally radial transverse transport is suggested by northward-directed paleocurrents in the Schenectady Formation in New York (fig. 11C; data from Krueger, 1963; Bloomer, ms, fig. 14). These sediment dispersal patterns suggest a progressive evolution of compressive deformation of the continental margin. As compression began, deformation and uplift were initiated on the continental promontories in Virginia and New York. Later, continuing compression resulted in closing the basin in the reentrant and in deformation and uplift of the basinal sediments and volcanics, as well as continental basement rocks of the Baltimore domes microcontinent.

Terminal orogeny in the Appalachian–Ouachita system represents the end of compressive deformation which had begun with the earliest clastic-wedge sediments (fig. 11D). Like the time of initial compressive deformation, the time of termination of compression varies along the system. The ultimate shape of the Appalachian–Ouachita system reflects the original orthogonally zigzag outline of the continental margin, and the structural style reflects compression of a laterally non-uniform array of rocks that had accumulated along that margin. The direction of tectonic transport during convergence and compression of the continental margin was not parallel with the orientation of movement during rifting. Thus, the old transform faults have not become sites of extensive later strike-slip faults. Rather, convergence has been resolved into compressive stress along the entire margin, and the shape of the deformed belt has formed a best-fit curve around the old promontories and reentrants.

Within the reentrants (evolving as salients), compressive stress has been transmitted toward the continental crust through the medium of thick incompetent clastic sediments of all three genetic types. The effect has been intense deformation of the basinal facies along with incorporated continental crustal fragments and volcanic rocks. In the reentrants, these rocks have been thrust toward and, to some extent, over the old bank edge. Deformation has propagated farthest cratonward along decollements within clastic-wedge sequences, in part over undeformed carbonate-bank rocks and older basement. In contrast, on the promontories (evolving as recesses), compression has been transmitted directly through continental crust. The thinner and dominantly more competent sedimentary rocks in the promontories (recesses) exhibit a more narrow belt of deformation than the adjacent reentrants (salients). Moreover, compression of crustal rocks has resulted in thick-skinned deformation significantly involving pre-Appalachian Grenville-age basement rocks in Appalachian folds and thrust faults within promontories (recesses). The importance of the thick-skinned style in promontories (recesses) is reflected in the present distribution of basement outcrops (fig. 11D).

CONCLUSIONS

Appalachian–Ouachita Late Precambrian and Paleozoic rocks include three successive genetic types: (1) clastic-volcanic sequence, (2) carbonate bank, and (3) clastic wedges. The outline of Appalachian–Ouachita structural salients and recesses reflects the distribution of each of the three types. Because these sediments span the full range of Appalachian–Ouachita geosynclinal history, the general shape of salient-recess geometry must have been established at the inception of the Appalachian–Ouachita geosyncline. Distributions of sedimentary facies and thickness suggest that the original framework of the geosyncline was an orthogonally zigzag continental margin, and the shape of Appalachian–Ouachita salients and recesses constitutes a best-fit curve around that framework. The original shape of the continental margin may be the expression of several transform faults along a Late Precambrian rift. Reentrants (concave oceanward) in the continental margin have evolved as structural salients, and promontories (convex oceanward) have become sites of structural recesses.

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