# Stratigraphic Notes, 1980–1982

CONTRIBUTIONS TO STRATIGRAPHY

GEOLOGICAL SURVEY BULLETIN 1529-H

*Seventeen short papers deal with changes in stratigraphic nomenclature: names adopted, revised, reinstated, or abandoned and ages of rocks changed*

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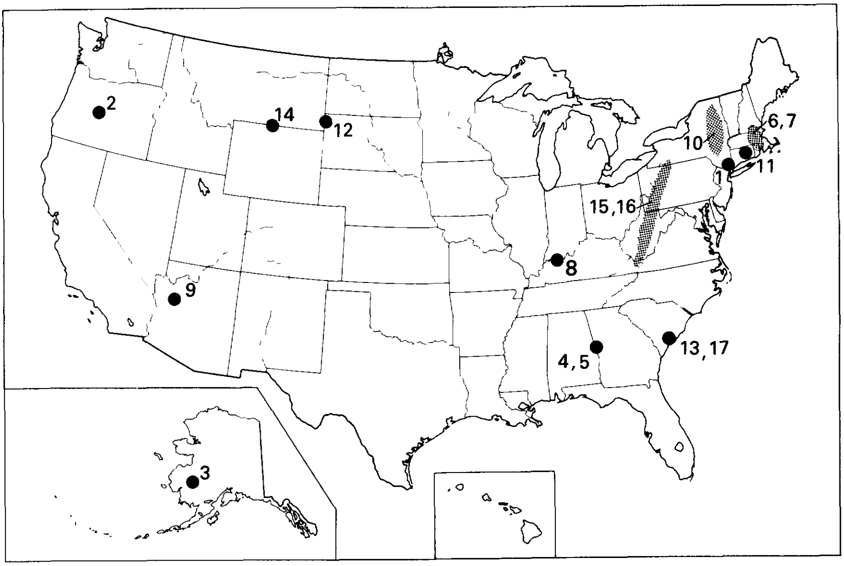


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# ADOPTION OF THE NAME HUTCHINSON RIVER GROUP AND ITS SUBDIVISIONS IN BRONX AND WESTCHESTER COUNTIES, SOUTHEASTERN NEW YORK

By Charles A. Baskerville

## INTRODUCTION

The Manhattan Prong of the New England upland, southeast of the Webster Avenue valley in Bronx County and the Bronx River valley in Westchester County, N.Y., consists of a series of low northeast-trending ridges and intervening valleys. The ridges generally are underlain by granite, gneissoid granite, and gneiss containing some interbedded amphibolite. The valleys are underlain by schist, gneiss, and amphibolite.

The purpose of this paper is to formally adopt the name Hutchinson River Group and to define the group which is composed of the rock types described above, and its included formations and subdivisions.

## HUTCHINSON RIVER GROUP (HERE FORMALLY NAMED)

The Hutchinson River Group is here formally named for the Hutchinson River that traverses most of the subject terrane from Scarsdale, Westchester County, on the north to Eastchester Bay (fig. 1; also see U.S. Geological Survey topographic map of the Mount Vernon and Flushing, N.Y. 7 1/2-min quadrangles) in the east Bronx on the south. Seyfert and Leveson (1968, 1969) used the Hutchinson River Group on the basis of an informal oral communication from C. A. Baskerville (1967). Most of the rocks of the group are well exposed along the length of the river and (or) within a mile of it. The group consists of (in ascending order) the Hartland Formation with its members and the Harrison Gneiss. Figure 1 outlines the known outcrop area of the Hutchinson River Group.

The contact between the Hutchinson River Group and the New York City Group of Prucha (1956) is shown on the Geologic Map of New York (New York State Museum and Science Service, 1971) as a queried dashed line. This line is Cameron's Line (a fault) of western New England (Robinson and Hall, 1979, fig. 2; Rodgers, 1970; Hall, 1979, fig. 3). The line, as it passes through the White Plains area of Westchester County on the Geologic Map of New York, is reasonably well defined; it fits the detailed mapping done by Hall (1968a, 1976) in the White Plains, N.Y.-Glenville, Conn., area. Pellegrini (1977) mapped the Mamaroneck, N.Y.-Conn., quadrangle, his work does not extend far enough west to further define Cameron's Line south of Hall's area. More detailed work in the Mount Vernon, N.Y., quadrangle, to the west of Pellegrini's area would fill in the gap between the work done by Hall and this author's area in the Bronx (southern part of the Mount Vernon quadrangle and the Flushing and Central Park, N.Y., sheets). Cameron's Line, therefore, has been extended south of White Plains on the Geologic Map of New York (New York State Museum and Science Service, 1971), bisecting the area east of the Bronx River valley, as an unchecked approximation.

In this paper, on the basis of 1980 mapping by the author (unpub. data), Cameron's Line has been adjusted to what appears to be a closer

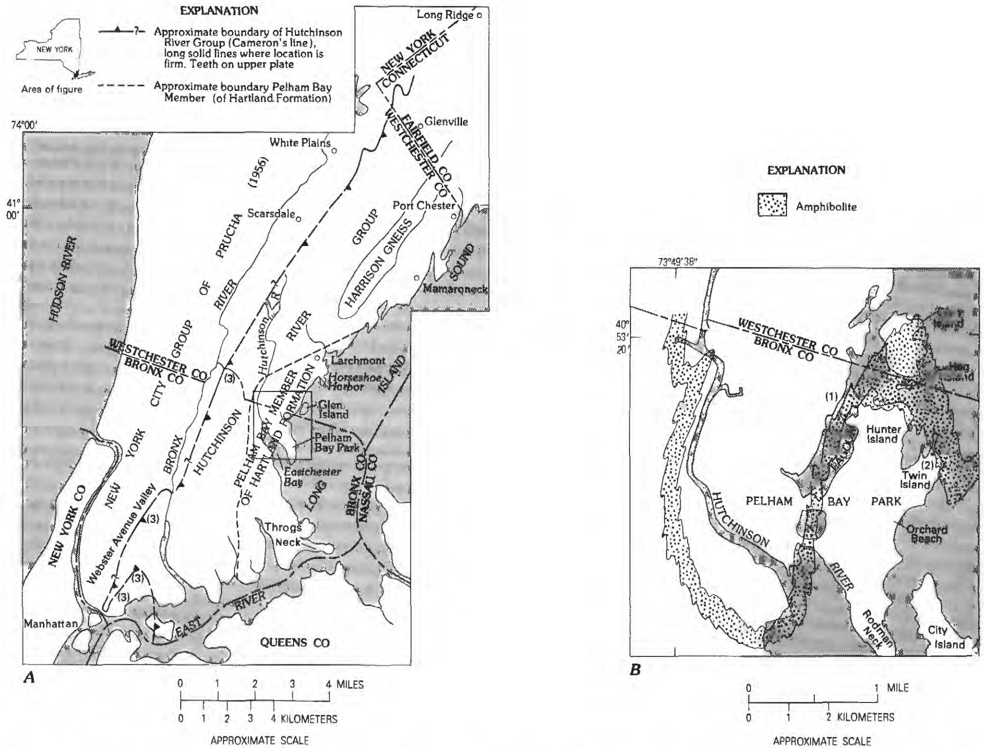


FIGURE 1.—*A*, Relation of the study area, Manhattan Prong of New England upland, to the State of New York and the known outcrop area of the Hutchinson River Group in Westchester and Bronx Counties, N.Y. Square enclosing Pelham Bay Park is location of *B*, approximate outcrop area of the Pelham Bay Member of the Hartland Formation, a large mappable amphibolite. The location of quartz-sillimanite gneiss (1) and of a K-Ar-dated sample obtained by Long and Kulp (1962) (2) are shown on *B*. Location of Cameron's Line (3) is indicated on *A*.

alinement with the known locations of Hartland rocks to the east and Prucha's New York City rocks to the west. In the vicinity of the Cross Bronx Expressway, east of Third Avenue, where the highway cuts through Crotona Park, outcrops separated by about 10 m can be seen; the rocks to the west are Manhattan Schist, those to the east are schists of the Hartland Formation (see solid portion on Cameron's Line in fig. 1(3)). Cameron's Line is hypothesized to be a Taconian thrust surface separating the western Proterozoic Y (Fordham Gneiss in the New York City area) and miogeosynclinal Cambrian and Ordovician (Prucha's New York City Group) rocks from the eastern Cambrian and Ordovician eugeosynclinal rocks (Hartland Formation and Harrison Gneiss). The allochthonous eastern units are thought to be thrust westward over Prucha's miogeosynclinal New York City Group (Hall, 1968b, 1979; Rodgers, 1970, 1971; Gates and Martin, 1976).

The Hartland rocks along strike to the northeast exhibit the same lithologies from the east Bronx northeastward to near Bridgeport Conn., where they contact the Straits Schist and Prospect Gneiss on their eastern border (Crowley, 1968; Hatch and Stanley, 1973). The Prospect is followed eastward by Gregory's (in Rice and Gregory, 1906) "Orange phyllite" and "Milford chlorite schist" units, respectively; the latter borders the Triassic and Jurassic basin at New Haven, Conn. The Prospect, "Orange," and "Milford" units probably would be the rocks in contact on the east with the Hartland rocks southward beneath Long Island Sound; this Prospect-"Orange"-"Milford" sequence is considered to range in age from Middle Ordovician to Silurian or even Devonian from west to east (Rodgers and others, 1956, 1959). If similar units do exist east of the Hartland Formation in New York, they are beneath Long Island Sound and extend under Long Island east of the New York City line between Queens and Nassau Counties (fig. 1).

The name Hutchinson River Group was proposed informally by the author while he was engaged in a reconnaissance study of the rocks in Westchester and Bronx Counties, N.Y., in 1965 (Baskerville, 1967). Seyfert and Leveson (1968) later mapped Twin Islands in Pelham Bay Park in detail. The author suggested that they use the name Hutchinson River Group as so little was known about the rocks east of Prucha's New York City Group at the time. The Geologic Map of New York showed these rocks as "schists and gneisses, undivided" (New York State Museum and Science Service, 1962). Subsequent publication by Seyfert (1968) and Seyfert and Leveson (1969) of their Twin Islands work placed the name Hutchinson River Group in the literature. Work by Hall (1968b) in the Glenville area of Connecticut (fig. 1) and southeastern New York resulted in a subdivision of the Hartland stratigraphy into four informal members.

Rocks of the group have been considered to be pre-Triassic (Rodgers and others, 1956, 1959; Carr, 1960) and, more specifically, Cambrian and Ordovician (Hall, 1976). Clark and Kulp (1968) obtained a K-Ar age of 320 m.y. ± 12 m.y. on muscovite from a muscovite-biotite schist near Port Chester, N.Y., in the Interstate 287-Interstate 95 interchange area. Long and others (1959) obtained a 375-m.y. K-Ar age on biotite from the Harrison Gneiss at Long Ridge, Conn. (fig. 1). Long and Kulp (1962) determined a K-Ar age of 380 m.y. on biotite from gneiss in the east Bronx. Long and Kulp's east Bronx biotite sample was assumed to be Fordham Gneiss of Prucha's New York City Group, but, from the map location (fig. 1B), the author thinks that it probably belongs to the Hartland Formation. An earlier event, affecting New England upland rocks, is indicated by remnant K-Ar ages of 460 m.y. to 480 m.y and is probably an early Taconian phase (Clark and Kulp, 1968; Long and Kulp, 1962). The Taconic orogeny took place at the end of the Ordovician about 435 m.y. ago, and the Acadian orogeny extended from the Middle Devonian into the Mississippian (Rodgers, 1970, p. 216). These radiogenic dates indicate that the Hutchinson River Group was subjected to both the Taconian and Acadian events. Thus, the rocks composing the Hutchinson River Group may range in age from Early Cambrian to as young as Middle Ordovician.

The Hartland Formation and Harrison Gneiss have been mapped from southwestern Connecticut (Hall, 1968a, b, 1976, Rodgers and others, 1956, 1959) through eastern Westchester County (Hall 1968a, b, 1976; Pellegrini, 1977; Baskerville, 1967, and unpub. data, 1980) to the East River. The thickness of the group or any of its subdivisions is impossible to determine accurately because of isoclinal folding, great variations in outcrop width along strike, questionable contact locations, and lack of continuity along strike. Figure 2 is a diagrammatic stratigraphic column of the Hutchinson River Group.

## HARTLAND FORMATION AND ITS SUBDIVISIONS

The name Hartland was first used by Gregory (in Rice and Gregory, 1906) to describe the rocks in Connecticut that he called Hartland (Hoosac) Schist; no type locality was indicated. The Hartland Formation contains three (Carr, 1960; Gates, 1959; Pelligrini, 1977) or four (Gates and Martin, 1976; Hall, 1976) units or informal members and is named for Hartland township, Hartford County, Conn.

The informal units of The Hartland as used by Hall (1976) and here adopted are (in ascending order) the amphibolite member, not everywhere present; the schist-gneiss-amphibolite member; the light-gray gneiss member; and the schist and granulite member. These rocks crop out over a width of 6 to 8 km. Hall (1976) calculated an average thickness of 2,650 m in the White Plains-Glenville area. Tight isoclinal folding in the east Bronx has probably repeated the stratigraphic section at least twice in the 3.5- to 5.5-km width of outcrop. Therefore, the average thickness for the Hartland in the east Bronx is probably 1,750 to 2,750 m.

## PELHAM BAY MEMBER (HERE NAMED)

The Pelham Bay Member consists of a complex of interbedded thick mappable amphibolites (fig. 1B), some of which contain large garnet porphyroblasts, light- and medium-gray ptygmatically folded gneisses, schists (all of which are very high grade metamorphic rocks, almost gneissic in this unit), granulite, and coarse pegmatites, some of which are quite thick. On the west side of the lagoon, west of Hunter Island ((1) in fig. 1B), biotite gneiss containing garnet is found interbedded with granulite and amphibolite and gray to white muscovite gneiss. The biotite gneiss has quartz-sillimanite bands containing large sillimanite crystals. On the basis of composition, this quartz-sillimanite biotite gneiss unit appears to correlate with the schist and granulite member of Hall (1976 and oral commun., 1980).

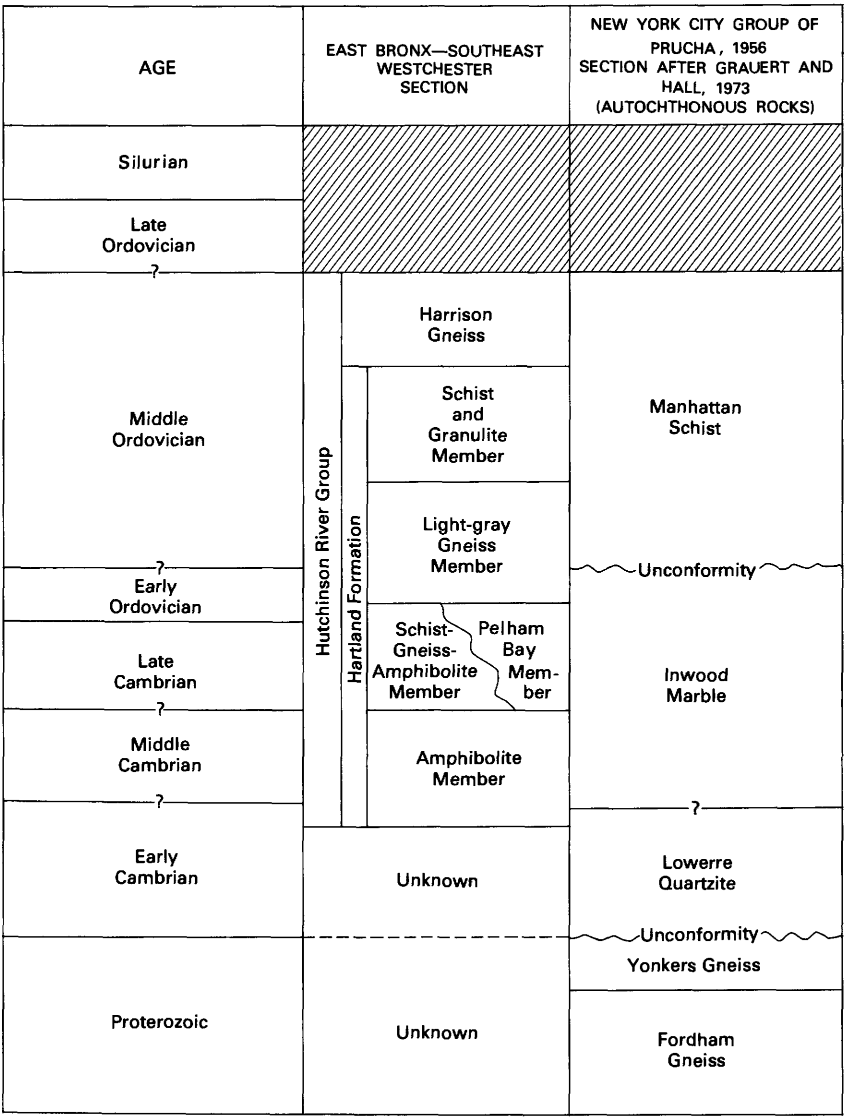


FIGURE 2 – Diagrammatic stratigraphic column of the Hutchinson River Group at its type locality near Hutchinson River in Manhattan Prong of New England upland southeast of Webster Avenue valley in Bronx County and east of Bronx River valley in Westchester County, southeastern New York. Informal members of the Hartland Formation adopted from Hall (1976).

The entire rock suite in the Pelham Bay Park area appears to be the result of very high grade progressive metamorphism beyond that normally found in the Hartland Formation. Enough relict textures or fabrics remain from "typical" Hartland lower grade rocks that the author can see a relation to the schist and granulite in some facies and possibly to the schist-gneiss-amphibolite members of Hall (1976) with the rocks seen west of the Hutchinson River valley and north of Pelham Bay Park in eastern Westchester County.

Land exposures indicate that the Pelham Bay rocks may be as thick as 1,500 m. The complex disappears to the east beneath Long Island Sound and may be much thicker. The Pelham Bay Member is here named for, and its type area is given as, Pelham Bay Park in northeast Bronx County.

Pelham Bay Park (see U.S. Geological Survey Flushing, N.Y., 7 1/2-min quadrangle) is chosen as the type area because it is a park and, as such, the outcrops will be protected from urbanization. They should remain easily accessible for study in the future. Part of this unit was mapped in detail on Twin Islands by Seyfert and Leveson (1969). The author has mapped the same units to the west, south (west of Eastchester Bay), and north (Hunter Island) of the Twin Islands, as well as into Westchester County in Glen Island Park (unpub. data, 1980; see U.S. Geological Survey Mount Vernon, N.Y., 7 1/2-min quadrangle).

Pellegrini (1977) mapped similar rocks in the Mamaroneck, N.Y.-Conn., 7 1/2-min quadrangle farther northeast in Westchester County. Pellegrini (1977) included rocks in the Horseshoe Harbor area of Mamaroneck that are equivalent to the high-grade metamorphic rocks of the Pelham Bay Member in Hall's (1976) schist and granulite member of the Hartland.

The author's reconnaissance work (unpub. data, 1980) has not discovered similar rocks in southwestern Fairfield County, Conn.; they may very well be there, however. The exposures of this unit, then, are confined to the Long Island Sound coast of Westchester and Bronx Counties of New York. The Pelham Bay rocks may be beneath the sound off the Connecticut coast and probably occur at depth in eastern Queens County, Long Island. South of Eastchester Bay to the Throgs Neck peninsula, at the confluence of Long Island Sound and the East River, no rock crops out along the shore. If there were outcrops inland, they are now covered by buildings and streets.

In this paper, the Pelham Bay Member is considered to be pre-Silurian in age, probably from Late Cambrian to Early Ordovician. This age is based on a 380-m.y. K-Ar date obtained by Long and Kulp (1962) on biotite from a sample in this area of the east Bronx. Hall (1976) has suggested that the rocks of southwestern Connecticut are increasingly younger across strike from the west to the Harrison Gneiss on the east. This decrease in age implies a synform in the Westchester area, according to Hall (written commun., 1981), with the Harrison Gneiss in the core. The rocks east of the Harrison are increasingly older.

The map pattern (fig. 1B) and field measurements (Baskerville, unpub. data, 1980) in Pelham Bay Park indicate a synform plunging northerly. If the Harrision Gneiss in the core of this synform in Larchmont, Westchester County, were projected south up plunge to the Pelham Bay area of the Bronx, it would be above present topography. If the units in these two locations (Larchmont and Pelham Bay) are indeed part of the same structure, the Pelham Bay rocks may be surface exposures of a deep-seated root part of the Hartland. These Pelham Bay rocks appear to have been subjected to more igneous activity than the more "typical" Hartland rocks, as the increased amount of amphibolite, pegmatite, and gneissoid rocks present indicates.

## HARRISON GNEISS (NAME ADOPTED)

Merrill (1898) named the Harrison diorite for extensive outcrops in the town of Harrison, Westchester County, southeastern New York. Eckel (1902), Berkey (1907), and Ziegler (1911), among others, also used the name Harrison diorite. Rodgers and others (1956, 1959) renamed and redescribed the Harrison as the Harrison Gneiss; this usage is herein adopted by the U.S. Geological Survey. More recently, Harrison diorite was used by Long and Kulp (1962) and Fluhr (1969). Clark and Kulp (1968) changed the name somewhat by calling it Harrison granodiorite gneiss. Fisher and others (in New York Museum and Science Service, 1971), used the name Harrison Gneiss in the compilation of the Geologic Map of New York. Hall (1968b), Sanders (1974), and Pellegrini (1975, 1977) most recently have used Harrison Gneiss in reference to the type rocks cropping out in the town of Harrison, Westchester County, N.Y.

Rodgers and others (1956, 1959) redescribed this unit as a dark-gray to greenish gneiss having an andesine, quartz, hornblende, and biotite composition and local augen structure. Their description fits the rock much better than descriptions of this unit as diorite by other workers. The term diorite applied to this rock unit would not be all encompassing, and the rock does not fit the plutonic definition of diorite. The Harrison has a predominant metamorphic gneiss texture (Ziegler, 1911).

Most geologists consider the Harrison Gneiss to be a single unit of rock that is well exposed in Harrison township, Westchester County, N.Y. Pellegrini (1977) proposed a second unit to the Harrison in the Mamaroneck, N.Y.-Conn., 7 1/2-min quadrangle. This second unit is similar to part of Hall's (1976) schist and granulite member of the Hartland Formation and is believed by the author to be part of the Hartland. Ries (1895) studied the petrography of the Harrison and gave locations in Westchester County where it can be seen.

Both the Hartland Formation and the Harrison Gneiss may be observed in almost continuous exposures from west (approximately from the Lake Street overpass, White Plains) to east (Port Chester, N.Y.) along the Cross Westchester Expressway (Interstate 287) and in the driveways and parking lots of vast adjoining industrial and office parks along the parallel service roads for about 11 km.

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# CARBONIFEROUS FORMATIONS IN CENTRAL OREGON

By Harold J. Buddenhagen[1]

U.S. Geological Survey Professional Paper 1110-CC, "The Mississippian and Pennsylvanian (Carboniferous) Systems in the United States--California, Oregon, and Washington" contained a report by Ewart M. Baldwin (1979), which included a map credited to me showing these formations in central Oregon. Said map was taken from an uncolored copy that was on open-file in the office of the Oregon State Department of Geology and Mineral Industries in Portland. For some reason, the map in Professional Paper 1110-CC differs substantially from my mapping. The geology of the area is very complex and obscure so it is understandable that different geologists would probably develop different interpretations of it. However, in this case, no new mapping was undertaken by Baldwin so it appears that the error resulted from misreading of my map: roads and other lines were read as formation boundaries, notably in section 25, T. 18 S., R. 24 E., where a strip of Upper Triassic Wade Butte Formation about 2,000 feet wide extending westerly across this section is shown erroneously as the Pennsylvanian Spotted Ridge Formation. To remedy this situation, the map on the following page (fig. 3) is offered as a substitute for figure 20, which is credited to me, in Professional Paper 1110-CC. For lack of space, this map does not show all occurrences of Carboniferous rocks present in this area, but it does show the principal ones.

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U.S. Geol. Survey Bull. 1529-H, 1982, p. H11-12

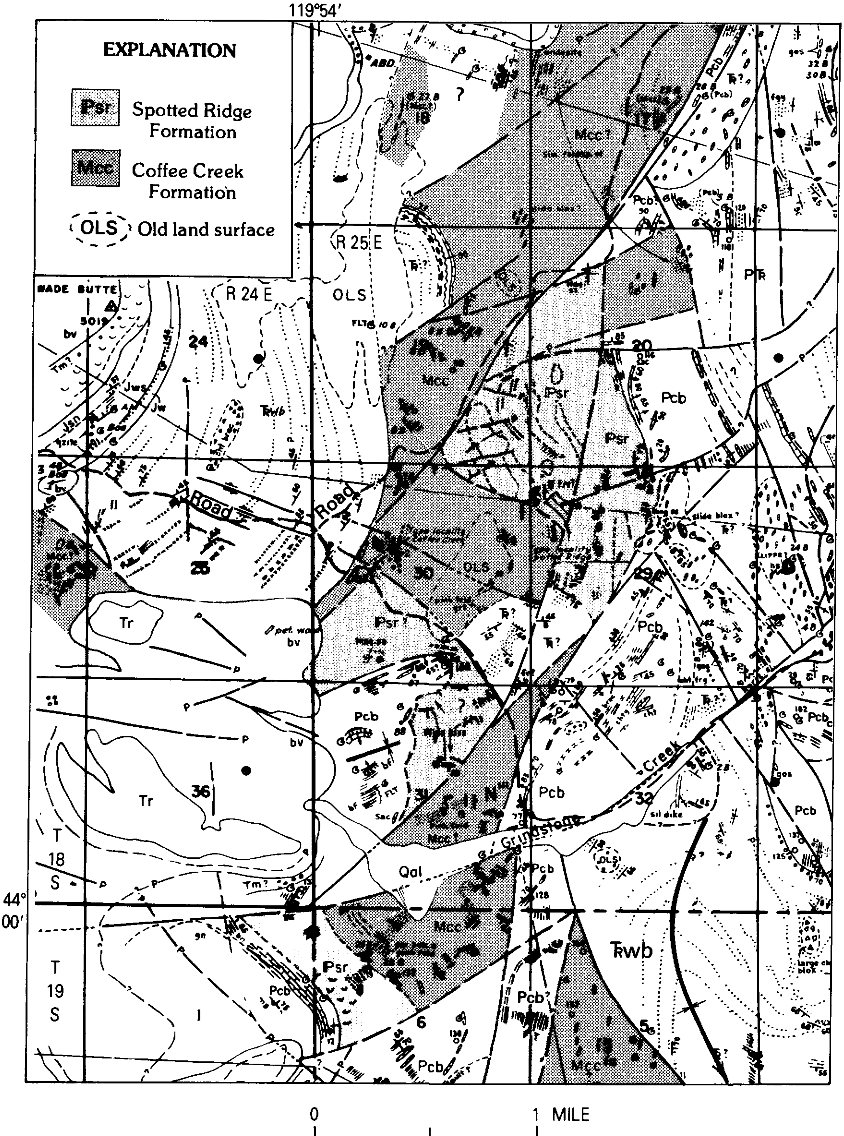



FIGURE 3.—Carboniferous formations in central Oregon: A correction to Baldwin (1979, fig. 20, p. CC38).

# NEW PALEOZOIC FORMATIONS IN THE NORTHERN KUSKOKWIM MOUNTAINS, WEST-CENTRAL ALASKA

By J. Thomas Dutro, Jr., and William W. Patton, Jr.

## ABSTRACT

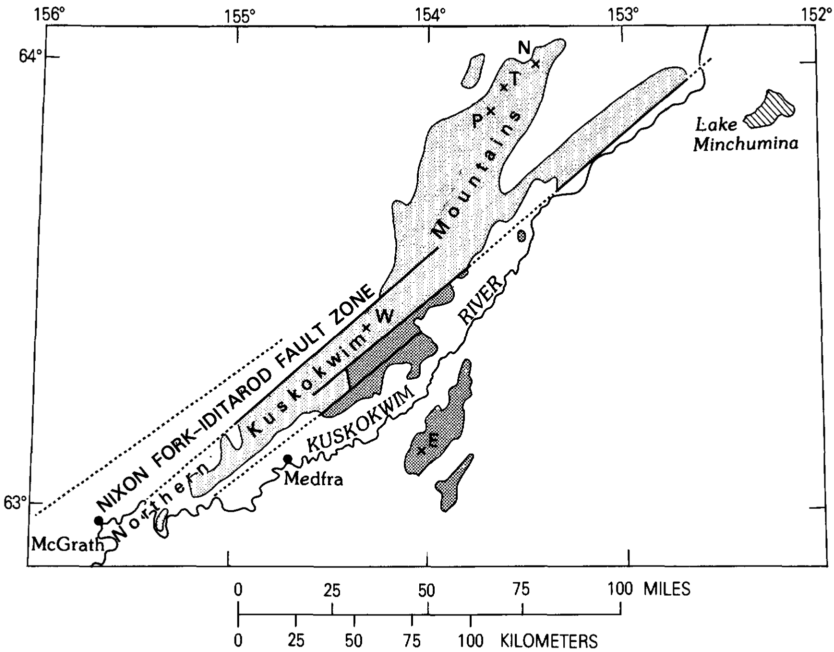
Five new formations are proposed for the lower and middle Paleozoic sequence in the northern Kuskokwim Mountains, west-central Alaska. Four of these units, exposed in the Nixon Fork terrane, reflect dominantly shallow-water carbonate depositional environments. The Novi Mountain Formation, of Early Ordovician age, is overlain conformably by the Middle and Upper Ordovician Telsitna Formation. The Paradise Fork Formation, deeper water dark platy limestone and shale, lies disconformably above the Telsitna Formation and is of latest Llandoverian to Wenlockian (Silurian) Age. The Whirlwind Creek Formation, of Late Silurian to Late Devonian (Frasnian) age, lies unconformably on either the Paradise Fork or the Telsitna. The East Fork Hills Formation, a deep-water facies equivalent of the other four formations, crops out southeast of the platform sequence and is separated from it by a northeast-trending fault zone.

## INTRODUCTION

Paleozoic rocks in the northern Kuskokwim Mountains occur in two different tectonic terranes that reflect widely separated depositional sites that subsequently have been juxtaposed by strike-slip faulting (fig. 4; Patton and others, 1980). The Nixon Fork terrane, composed of more than 5,500 m of predominantly shallow-water carbonate rocks of early and middle Paleozoic age, is fault-bounded on the southeast by the East Fork terrane composed of slightly metamorphosed, locally sheared and foliated, deep-water shaly carbonate rocks also of early and middle Paleozoic age. The Nixon Fork terrane is well exposed along a northeast-trending belt that underlies the higher parts of the northern Kuskokwim Mountains. The East Fork terrane is poorly exposed and is confined chiefly to low, densely forested hills bordering the Kuskokwim River valley. Both terranes are sliced by pervasive northeast-trending high-angle faults that parallel the nearby Nixon Fork-Iditarod fault zone.

## NIXON FORK TERRANE

The platform carbonate sequence of the Nixon Fork terrane is here divided into four new formations, ranging in age from Early Ordovician to Late Devonian (fig. 5). Depositional environments range from mainly supratidal laminated silty limestone in the Lower Ordovician through a complex array of shallow-water carbonate facies that include reefoid bodies in the Upper Ordovician and Middle Devonian. Dark platy limestones and shales containing mid-Silurian graptolites indicate deeper water paleoenvironments in that part of the Paleozoic column.




**EXPLANATION**

Nixon Fork terrane—Approximate location of type section of formation shown by letter: N, Novi Formation; T, Telsitna Formation; P, Paradise Fork Formation; W, Whirlwind Creek Formation

East Fork terrane—Approximate location of type section of formation shown by letter: E, East Fork Hills Formation

Contact

Fault—Dotted where concealed

FIGURE 4.—Areas of outcrop of Nixon Fork and East Fork terranes, northern Kuskokwim Mountains, west-central Alaska. From Patton and others, 1980.

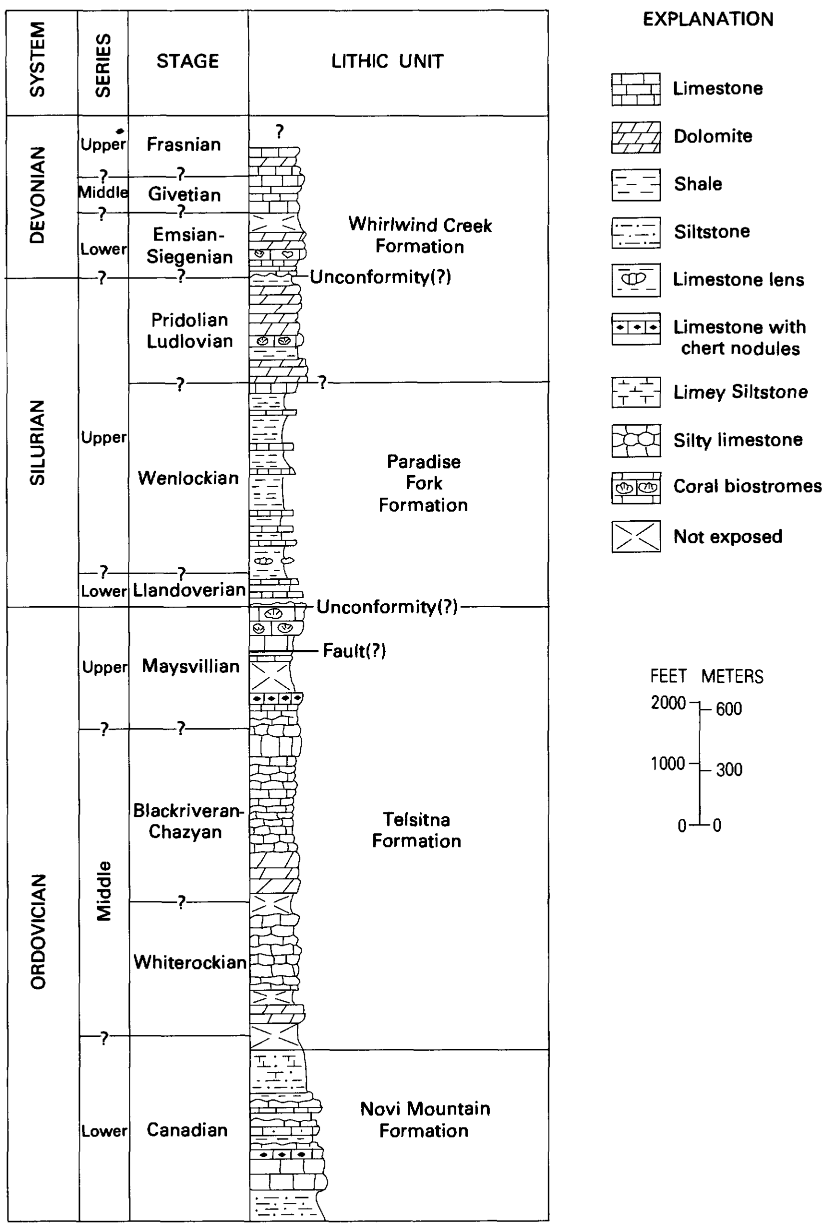



FIGURE 5.—Generalized columnar section of lower and middle Paleozoic rocks, northern Kuskokwim Mountains, west-central Alaska.

## NOVI MOUNTAIN FORMATION (HERE NAMED)

### Name and distribution

The name Novi Mountain Formation is here applied to a nearly 900-m-thick cyclically interbedded sequence of thin-bedded silty to micritic Lower Ordovician limestone and calcareous siltstone exposed in the vicinity of Novi Mountain. The type section is located on the east and west flanks of Novi Mountain, in sections 29, 30, and 32, T. 17 S., R. 28 E., Medfra (D-1) quadrangle.

The Novi Mountain Formation outcrops are characteristically variegated gray carbonate rocks alternating with yellow-weathering siltier members. The formation is mapped southwestward from its type section for about 25 km along a fault-bounded anticlinal structure (Patton and others, 1980). In addition, it is present in a tightly folded anticline along the same trend, about 50 km southwest of Novi Mountain. The formation also is mapped in the area south of Browns Fork and east of White Mountain Creek.

### Stratigraphy

The upper two-thirds of the Novi Mountain Formation is characterized by 5- to 30-m-thick carbonate cycles. Each cycle begins with massive limestone, usually containing flat carbonate pebbles at the base and locally abundant oolites. The thick-bedded limestone grades upward through thin irregularly bedded shaly limestone into calcareous siltstone or shale in the uppermost part of the cycle.

The type section of the Novi Mountain Formation, on Novi Mountain, is a composite of two measured sections with beds dipping generally northwest. The lower section is on the southeast spur of the mountain, measured from a low saddle at about 700 m elevation up to the crest of the ridge at 1,040 m, near triangulation station Higher. The upper part of the formation is exposed on the west side of the mountain where a second section was measured downhill and downdip from the ridgetop to an elevation of 640 m where the stratigraphically highest outcrops occur. The two sections are correlated along the ridge by lateral tracing of cyclic units.

The lower 150 m is predominantly yellow-weathering calcareous siltstone and shale with much evidence of bioturbation. The interval from 150-300 m is characterized by massive limestone beds, 3-5 m thick. At least a dozen cycles, as described above, occur from 300-600 m above the base of the formation. The upper 300 m of the Novi Mountain consists chiefly of thin irregularly bedded silty limestone and shale, in less clearly delineated cycles, with some units of dolomite, interpreted by us to be of supratidal origin. The Novi Mountain appears to grade upward into the Telsitna Formation in the high ridge southeast of the Telsitna River. No clear lower contact is mapped in that area, but the Novi Mountain lies directly on a lower Paleozoic or Precambrian calc-schist unit east of White Mountain Creek.

### Fossils and age

The Novi Mountain contains few megafossils, but a sequence of sparse conodont faunas studied by J. E. Repetski (written commun., 1977-80) indicates that the entire formation is of Early Ordovician age. Conodonts from the middle part of the formation, 300 to 600 m above its base, indicate North American Midcontinent Province Faunas C and D of the Lower Ordovician, according to Repetski. Included in these assemblages are Drepanodus parallelus Branson and Mehl s.f. (= in the form sense), Acontiodus iowensis Furnish s.f., Acontiodus cf. A. staufferi Furnish s.f., Drepanoistodus suberectus (Branson and Mehl), Juanognathus? sp., ?Scolopodus filosus Ethington and Clark s.f., Scolopodus cf. S. gracilis Ethington and Clark, Scolopodus sp., and Oneotodus cf. O. variabilis Lindstrom.

Megafossils include gastropod steinkerns, poorly preserved indeterminate orthoconic cephalopods, and a few fragmental trilobites. E. L. Yochelson (written commun., 1976) identified Sinuites sp. and Liospira sp., both common Ordovician genera. The trilobites include a possible hystricurinid and a pilekiinid, the latter limited to the Lower Ordovician, according to R. J. Ross, Jr. (written commun., 1978).

## TELSITNA FORMATION (HERE NAMED)

### Name and distribution

The name Telsitna Formation is here applied to an approximately 2,000 m-thick sequence of Middle and Upper Ordovician limestone and dolomite that conformably overlies the Novi Mountain Formation. The type section and type area are located on a high northeast-trending ridge situated along the divide between the head of the Telsitna River and Paradise Fork. The type section is almost completely exposed in southwest-dipping beds along the crest of the ridge in sections 17, 19, and 20, T. 18 S., R. 27 E., Medfra (D-2) quadrangle.

The Telsitna Formation is the most widely distributed Paleozoic map unit in the northern Kuskokwim Mountains. It extends from the type section southwestward to the Kuskokwim River near McGrath, essentially forming the northwest front of the mountains from near Novi Mountain on the north to Halfway Mountain on the south. Most of the prominent hills, including Limestone Mountain and Greens Head, consist of Telsitna Formation carbonate strata (Patton and others, 1980).

### Stratigraphy

The lower 300 m of the Telsitna Formation is dominated by variegated light-gray to dark-brown dolomite beds. From 300 to 600 m, the sequence consists mainly of thin-bedded fine-grained medium-gray limestone with silty yellow-weathering interbeds. This part of the formation is highly fossiliferous, including silicified brachiopods and molluscs. Nonfossiliferous dolomite occurs in the interval between 600 and 1,000 m, limestone beds becoming more abundant in the upper 200 m. The upper 600 m of the formation is predominantly limestone that ranges from thick bedded and fine grained to thin bedded, silty, and micritic. Some black chert is present as scattered small nodules and lenses about 200 m below the top of the formation. The overlying Silurian deep-water strata of the Paradise Fork Formation appear to lie disconformably on the Telsitna Formation, but the contact may be faulted.

### Fossils and age

Conodonts have been identified from nearly the entire Telsitna Formation (A. G. Harris and J. E. Repetski, written commun., 1979). Samples from about 300 m above the base, in beds that also include the lowest occurence of Maclurites, yielded conodonts of North American Midcontinent Province Middle Ordovician Fauna 3. Fauna 4 conodonts are reported from about 900 m above the base, and a Blackriveran (middle Caradocian) assemblage is identified from about 1,200 m above the base. Late Ordovician (probably Maysvillian) conodonts are present in the upper 50 m of the Telsitna Formation.

The common Ordovician gastropod Maclurites ranges through much of the formation, from 300 m to at least 1,600 m above the base (E. L. Yochelson, written commun., 1978). The upper beds also contain a number of corals, stromatoporoids, and trilobites of Late Ordovician age. W. A. Oliver, Jr. (written commun., 1977-78) reports Catenipora sp., cf. Fletcheria sp., Pycnolithus? sp., Saffordophyllum sp., Tetradium sp., Labyrinthites sp., and Labechia spp., of possible Maysvillian Age. Trilobites from the upper part of the formation, probably within the upper 100 m, are Bumastoides cf. B. milleri (Billings) and Sphaerexochus sp., of late Middle or Late Ordovician age (R. J. Ross, Jr., written commun., 1976). Ostracodes from about 1,300 m above the base are Blackriveran in age, according to J. M. Berdan (written commun., 1980). The assemblage includes: Platybolbina (Platybolbina) sp., Craspedopyxion? aff. C.? tumblingrunensis (Kraft, 1962), Eurychilina aff. E. strasburgensis Kraft, 1962, Leperditella? aff. L.? asymmetrica (Kraft, 1962), "Leperditella" aff. "L." altiforma Harris, 1957, Steusloffina? sp., and Krausella sp.

## PARADISE FORK FORMATION (HERE NAMED)

### Name and distribution

The name Paradise Fork Formation is here applied to a poorly exposed Silurian sequence, at least 1,000 m thick, of dark-gray thin-bedded platy limestone and black shale. The best exposures and type locality of the formation are located in low hills between upper Paradise Fork and upper Sulukna River drainages in sections 11, 12, 14, and 23, T. 19 S., R. 26 E., Medfra (D-2) quadrangle.

This formation is mapped southwestward from its type area to White Mountain Creek, where it is terminated by a major northwest-trending vertical fault (Patton and others, 1980). The Paradise Fork Formation forms a generally southwest-trending synform that includes younger Devonian beds just north of Stone Mountain. Stone Mountain itself is composed mostly of undifferentiated Cretaceous strata and Upper Cretaceous and lower Tertiary monzonite bodies. No strata assignable to the Paradise Fork are recognized in the southern part of the carbonate outcrop belt.

### Stratigraphy

The Paradise Fork Formation overlies, possibly disconformably, the Telsitna Formation and appears to reflect a change to a deeper water depositional environment. The dominant rock types in the lower half of the formation are dark-gray platy silty limestone and interbedded black silt shale containing nodular concretions at several horizons. Isolated limestone lenses and bodies up to 5 m thick occur in the upper part of the sequence.

The concretions, up to 12 cm in diameter, are composed of dark silty limestone and generally contain fossils or other foreign objects in their cores. Fossils include graptolites, ostracodes, gastropods, and cephalopods. In some instances, pebbles form the cores of the concretions, most of which also contain bits of bituminous material. Silty, yellow-weathering laminae, probably dolomitic, are common at several levels throughout the formation.

### Fossils and age

The only diagnostic fossils from the lower part of the Paradise Fork Formation are graptolites that are identified by Claire Carter (written commun., 1978) as Monograptus cf. M. parapriodon Boucek and Paraplectograptus aff. P. eiseli (Manck). Carter suggested a latest Llandoverian to early Wenlockian Age assignment. No beds dated as latest Ordovician or early Llandoverian were found between the uppermost dated Telsitna strata (Maysvillian) and the lowermost graptolite-bearing beds that are no more than 50 m above the base of the Paradise Fork. The absence of these beds suggests that there is a disconformity between the two formations, but field relations also allow for minor faulting along the contact. The age of the upper part of the formation is only roughly estimated, but ostracodes from near the top of the exposed sequence are identified by J. M. Berdan (written commun., 1979) as Herrmannina cf. H. caeca (Jones, 1891). Berdan states, "...Copeland (1976) shows Herrmannina consistently occurring below Leperditia in northern North America, and H. caeca below L. arctica in beds dated as Wenlockian in the south-central Arctic Islands." Consequently, the age of the uppermost part of the Paradise Fork is probably not younger than Wenlockian.

## WHIRLWIND CREEK FORMATION (HERE NAMED)

### Name and distribution

The name Whirlwind Creek Formation is here applied to an Upper Silurian to Upper Devonian sequence of predominantly shallow-water carbonate rocks, 1,000-1,500 m thick. The type section is a south-dipping sequence exposed on the ridge between Whirlwind and Soda Creeks in section 23, T. 24 S., R. 23 E., Medfra (B-3) quadrangle. The higher beds of the formation are measured in a supplemental section on the north flank of the syncline in sections 29 and 30, T. 23 S., R. 25 E., Medfra (B-3) quadrangle.

The main outcrop belt of the Whirlwind Creek Formation stretches northeastward for about 100 km from the type area to approximately latitude 64° N. (Patton and others, 1980). Exposures are generally poor northeast of Hardscrabble Creek. The southwestern limit of the belt is controlled by two intersecting faults north of Limestone Mountain. Two outlying masses of Whirlwind Creek Formation occur north of Stone Mountain and along a northwest-trending ridge west of White Mountain Creek. These two exposures are regionally significant because they show stratigraphic relations with adjacent formations. North of Stone Mountain, the Whirlwind Creek apparently lies above the Paradise Fork Formation, whereas it lies unconformably on the Telsitna Formation throughout most of the region. West of White Mountain Creek, the Whirlwind Creek is unconformably overlain by Permian through Lower Cretaceous strata.

### Stratigraphy

Although not well exposed, the base of the formation in the type section appears to overlie the Telsitna Formation disconformably. The Whirlwind Creek consists of relatively thick cycles of dolomite and limestone that include some silty intervals and reefoid units with Favosites and other corals and stromatoporoids.

The lowermost cycle in the type section, about 250 m thick, starts with 80 m of algal laminated dolomite that grades upward into pelletoidal limestone and, finally, into silty limestone and siltstone between 80 and 170 m above the base. The upper 80 m of this major cycle is thick-bedded reefy limestone with Favosites that grades into thin-bedded limestone in the uppermost 30 m. A second major cycle, much like the lowermost one, is about 400 m thick. A third cycle, also about 400 m thick, occurs at the top of the type section.

In the supplemental section, cycles do not appear to be as well developed, although the lower 250 m represents a major cycle that correlates with the highest cycle in the type section. The upper 300 m of the supplemental section consists of reefy coral-bearing limestone and dolomite, the latter containing several intervals of dolomite-breccia in the upper 100 m.

### Fossils and age

The best age control for the lower part of the Whirlwind Creek is provided by ostracodes present in the upper parts of each major cycle. Ostracodes from two lower horizons, about 170 and 500 m above the base of the formation, are identified by J. M. Berdan (written commun., 1976) as Leperditia aff. L. arctica Salter, 1853, Sulcatiella Polenova, 1968, and Tubulibairdia? sp. Berdan states, "... It seems most probable that both collections are of Late Silurian age, but the possibility of a Devonian age cannot be ruled out."

Ostracodes from about 600-900 m above the base of the formation are probably of Siegenian (Early Devonian) Age, according to Berdan who states (written commun., 1979), "There does not appear to be any distinct change in the]character of the ostracode assemblage as one goes up through [this part of the section. The taxa listed are close to forms described by Polenova from the Lower Devonian of Siberia, from localities ranging from Novaya Zemlya on the west to the Sette-Daban Range on the east, and south to the Salair Range on the southwest margin of the Kuznetz Basin. Except for cosmopolitan forms, there are few taxa in common with described North American faunas....There appear to be more Siegenian taxa than either Gedinnian or Emsian but, as ostracode ranges are not yet well known, both Gedinnian and Emsian beds could be present as well as Siegenian."

From the supplemental section, a relatively rich brachiopod assemblage and the coral Rhizophyllum, from about 250 m below the top, suggest an Emsian Age. Corals from elsewhere in the outcrop belt were identified by W. A. Oliver, Jr. (written commun., 1977). Alaiophyllum, a Middle Devonian genus, occurs in the outcrop north of Stone Mountain. Strata on both limbs of the syncline on upper Soda Creek contain both Middle Devonian and Late Devonian (Frasnian) corals. The Frasnian beds contain Smithiphyllum sp. These corals occur together with abundant Amphipora in beds that are correlated with the dominantly dolomitic unit in the upper 100 m of the supplementary section.

## EAST FORK TERRANE

## EAST FORK HILLS FORMATION (HERE NAMED)

### Name and distribution

The name East Fork Hills Formation is here applied to a poorly exposed sequence of at least several hundred meters of Lower Ordovician to Middle Devonian dark-gray orange-weathering finely laminated limestone and dolomitic limestone. Because of the scattered outcrops, no continuous section is exposed, although several tens of meters can be examined at the designated type locality along the crest of the East Fork Hills in section 14, T. 27 S., R. 25 E., Medfra (A-3) quadrangle.

The East Fork Hills Formation is mapped along both sides of the North Fork of the Kuskokwim River, from just east of Limestone Mountain to Hardscrabble Creek and southeastward to Moose Hill (Patton and others, 1980). None of the outcrops is particularly well exposed, and relations with other mapped units are obscure. Everywhere along the northwest edge of the outcrop belt, the East Fork Hills Formation is in fault contact with the Whirlwind Creek and Telsitna Formations of the Nixon Fork terrane. Paleozoic chert and phyllite in Grayling Hill, east of the East Fork Hills, may be a deeper water facies equivalent, as they are correlated with similar rocks near Lake Minchumina that have yielded probable Ordovician graptolites and radiolaria.

### Stratigraphy

Outcrops of the East Fork Hills Formation are characterized by alternating thin bands of limestone and orange-weathering dolomite, locally sheared and foliated. Small-scale crossbedding and penecontemporaneous slump structures are common. Exposures on upper Soda Creek and Beaver Creek contain subordinate amounts of laminated dolomite, dark chert, and siliceous siltstone. Most of the outcrops are shattered by small-scale shearing and faulting, resulting in rubbly colluvial slopes. This formation is interpreted as a deep-water facies, approximately equivalent to the entire shallow-water carbonate sequence of the Nixon Fork terrane represented by the Novi Mountain, Telsitna, Paradise Fork, and Whirlwind Creek Formations.

### Fossils and age

Conodonts are reported from several localities in the East Fork Hills Formation (J. E. Repetski, written commun., 1977-79), and ages range from Early Ordovician to Middle Devonian. This nearly complete overlap with the age range of the platform sequence supports the interpretation of the East

Fork Hills sequence as a deep-water equivalent of the Nixon Fork terrane. The original relative positions of the two terranes, before the period of lateral faulting that has juxtaposed them, are unknown.

In addition, we suggest that the East Fork Hills Formation is at least partly correlative with the lime mudstone and shale of the Dillinger River region in the western part of the Talkeetna quadrangle to the east in the southern Alaska Range (Reed and Nelson, 1977).

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# NEW STRATIGRAPHIC UNIT IN THE WILCOX GROUP (UPPER PALEOCENE-LOWER EOCENE) IN ALABAMA AND GEORGIA

By Thomas G. Gibson

This report defines and names a new lithostratigraphic unit in the Wilcox Group on the basis of exposures in the coastal plains of eastern Alabama and western Georgia. The Baker Hill Formation of late Paleocene age is proposed for largely kaolinitic clay and crossbedded sand strata previously included in the time-equivalent, but lithologically distinct, Nanafalia Formation.

Mapping of Paleocene through middle Eocene strata in eastern Alabama and western Georgia has shown that the proposed unit is widespread in areas updip from those that contain the "classical" exposures of largely shelly glauconitic sand that are retained in the Nanafalia Formation. The Baker Hill Formation is well exposed in the Chattahoochee River valley drainage system in westernmost Georgia and easternmost Alabama, but also occurs over a wide area in the eastern Gulf Coastal Plain.

## BAKER HILL FORMATION (HERE NAMED)

### Stratigraphy

This new formation includes the kaolinitic and bauxitic massively bedded clay, carbonaceous clay, and crossbedded micaceous sand found in the vicinity of Baker Hill, Ala. (fig. 6-A), in the Eufaula bauxite district; the formation extends a considerable distance westward into Alabama and eastward into Georgia. These strata typically have been placed in the Nanafalia Formation by previous authors (Warren and Clark, 1965; Clarke, 1972). Although the strata herein placed in the Baker Hill Formation can be shown to be time equivalent to the shelly glauconitic sand of the Nanafalia (Edwards, 1980; Gibson, 1980), they are easily separable lithologically. The Baker Hill Formation occurs to the north and thus updip from exposures of the typical Nanafalia Formation. The Baker Hill represents deposition in fluvial and estuarine environments during the coastal onlap cycle in which the Nanafalia was deposited in inner neritic environments to the south.

The type locality of the Baker Hill Formation is herein designated as the Lynn Griffin #1 mine in Henry County, southeastern Alabama (fig. 6-A), 6 miles southwest of Baker Hill. The exposure on the south wall of the mine exhibits most of the common facies of the formation (fig. 7). The upper contact of the formation is exposed here, but the lower contact is below water level in the mine. This mine probably will continue to be a good exposure, but numerous other abandoned mines in the Baker Hill area also are available as reference sections. In addition, three other reference sections in Barbour and Henry Counties, Ala., are designated; these show various lithologies of the upper and lower portions of the formation and the contacts. Although kaolinitic clay is the most conspicuous lithology in the area, probably because of exposures in the mines, a significant thickness of the formation is composed of crossbedded quartz sand. The sand is fine- to

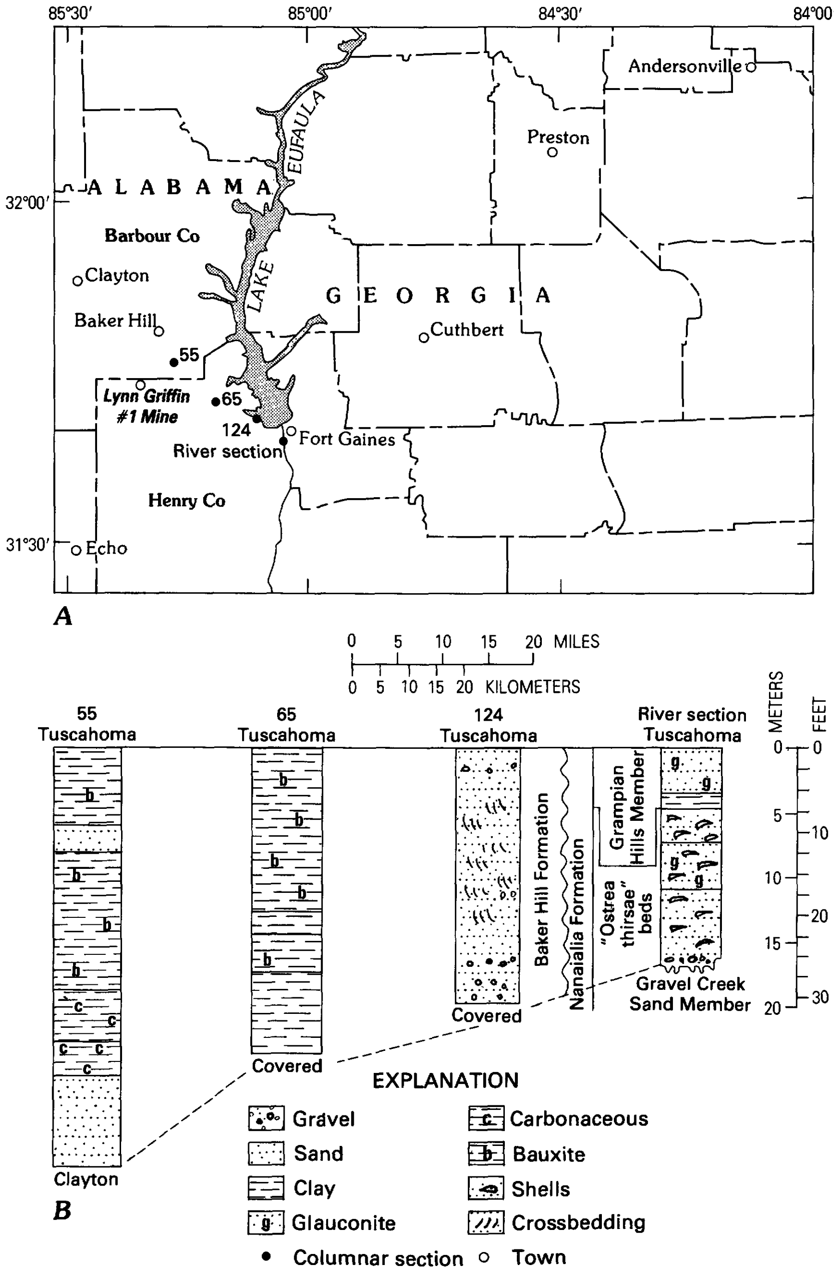


FIGURE 6.—*A*, Location of Baker Hill Formation, Ala. and Ga. *B*, Columnar sections of Baker Hill Formation. Sections show change in facies downdip in Baker Hill Formation and into Nanafalia Formation.

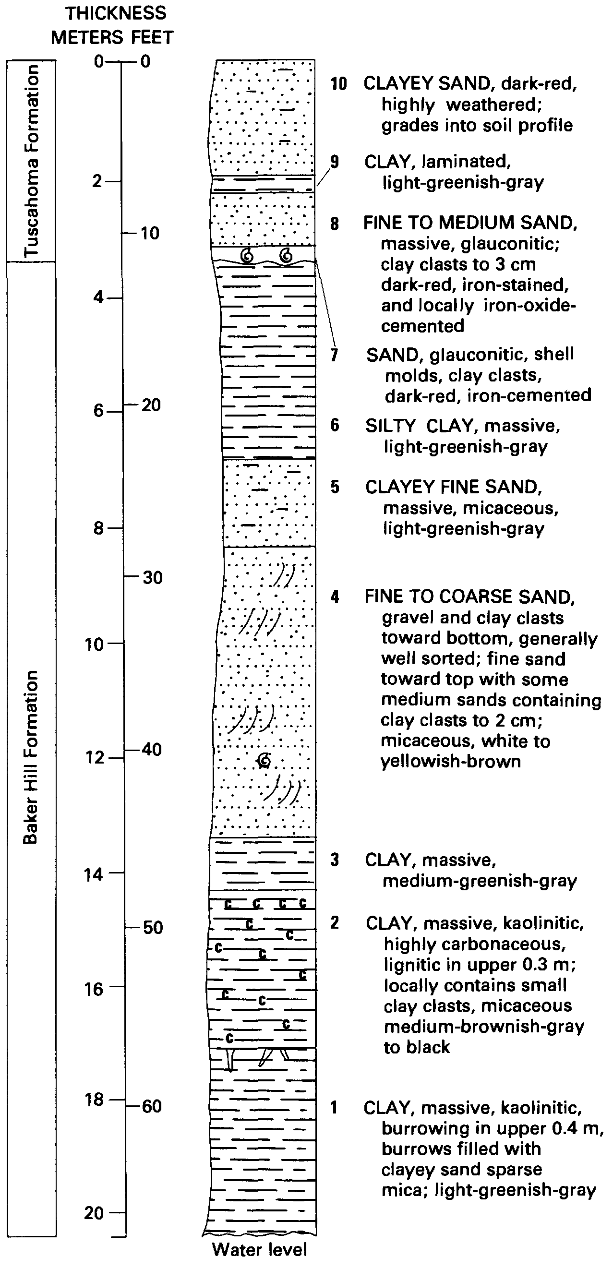


FIGURE 7.—South wall of Lynn Griffin mine #1, Henry County, Ala., type locality of Baker Hill Formation. Top of section 444 ft (135.3 m).

coarse-grained, poorly to well-sorted, and highly micaceous and commonly contains clay clasts. Crossbedded sand that generally composes the lower beds of the formation also occurs between thick clay sequences and may be abundant near the top of the formation. The sand is deep red in color in updip more weathered exposures and white to buff in more downdip less weathered exposures. The crossbedding includes both planar and trough types, commonly is large scale with sets up to 3 to 4 feet high, and dominantly dips southeast to southwest although complex sets with several directions occur. The kaolinitic clay is generally massive in appearance; in the Mathison mine (section 65, fig. 6-B), a zoning pattern suggestive of a soil profile was observed. The clay is usually light gray to light greenish-gray in color, although highly weathered sections are color mottled gray-green and red to purple. The highest alumina content is in the pisolitic beds that generally occur in pod- to funnel-shaped masses tens of feet in diameter; in mine exposures, these beds largely have been removed. Occasionally, carbonaceous clay beds are present. They can be as thick as 10 feet; one of the thickest beds is well exposed in the Lynn Griffin #1 mine. In a few places, the organic matter becomes concentrated enough to be considered lignite; these lignitic beds are 6 inches to several feet in thickness. As seen in subsurface cores in the Baker Hill area, several intervals of carbonaceous clay up to 8 to 10 feet thick may occur between kaolinitic clay beds. The facies changes among these lithologies are complex and rapid, both vertically and horizontally.

The type area of the Nanafalia Formation, by which these beds were previously referenced, is in western Alabama. There, the Nanafalia consists of a micaceous crossbedded sand named the Gravel Creek Sand Member at the base of the formation, a major overlying unit of glauconitic fossiliferous sand informally called the "Ostrea thirsae beds," and less fossiliferous glauconitic sand and clay of the Grampian Hills Member at the top. These three units are recognizable eastward into the Chattahoochee River valley, where they occur to the south or downdip from the area exposing the distinctive kaolinitic, bauxitic, and carbonaceous clay and crossbedded clay-clast-bearing sand of the Baker Hill Formation.

## Contact relations

The lower beds of the Baker Hill Formation usually are composed of red fine- to medium-grained sand that is crossbedded and micaceous and that contains clay clasts as large as several inches in diameter. These basal beds rest upon buff-colored limestone of the Clayton Formation, dark-red residual clay left from dissolution of the limestone, or light- to medium-gray massively bedded clay of the Porters Creek Formation. The surface is highly undulatory with 5 to 8 feet of relief noticeable within a single outcrop and approximately 70 feet found over a distance of a mile just north of Baker Hill. In the more downdip areas of the Baker Hill Formation, the crossbedded clay-clast-bearing basal sand of the Baker Hill rests, in places, upon micaceous carbonaceous sand of the Gravel Creek Sand Member of the Nanafalia Formation, which is sometimes found in sinkholes developed in the upper part of the Clayton Formation.

Throughout the Eufaula bauxite district and for a considerable distance west and east, the Baker Hill Formation is overlain by the Tuscahoma Formation. The considerably greater coastal onlap of the

Tuscahoma, which has an extensive marine transgressive deposit at the base (Gibson, 1980), gives a strongly planar surface to the top of the Baker Hill because of marine erosion. The basal beds of the Tuscahoma consist of shelly glauconitic sand containing abundant clay clasts and fine-grained quartz and phosphate gravel. Thus, the fresh greensand of the basal Tuscahoma (which weathers a deep red) contrasts sharply with the light-gray clay or buff sand of the uppermost part of the Baker Hill.

### Geographic extent and thickness

The Baker Hill Formation is widespread throughout eastern Alabama and western and west-central Georgia. It has been recognized southwestward to Echo, Ala., where field investigations stopped (fig. 6-A). To the east, the formation is well exposed in the Springvale bauxite district in western Georgia near Cuthbert (fig. 6-A), is exposed also between Cuthbert and Fort Gaines, and can be traced eastward to the Andersonville bauxite district at Andersonville, Ga. Between the Springvale and Andersonville districts, the exposed sections are dominated by crossbedded sand and impure clay. This known distribution gives a lateral extent of at least 90 miles for the formation with the high probability of an even greater extent both eastward and westward of its presently known limits. The northwest-southeast or updip-downdip extent also is considerable. In eastern Alabama, kaolinitic clay and clay-clast-bearing crossbedded sand extend from Clayton southward to about 2 miles northwest of Fort Gaines, Ga., a distance of 30 miles, essentially at right angles to the strike of the beds in this area (fig. 6-A). The northern limit at Clayton occurs at elevations around 600 feet; the Baker Hill is absent to the north because of the erosion surface that has cut down to Cretaceous rocks.

The formation has an average thickness of 80 to 100 feet and a probable range of 50 to 150 feet.

### Age

The age of the Baker Hill Formation in terms of intercontinental zonations can be established by its equivalency with the Nanafalia Formation. The unconformities both above and below the Nanafalia Formation continue updip and bound the Baker Hill Formation also (Gibson, 1980). In addition, Edwards (1980) found similar dinoflagellates of the same age both in marine beds of the Nanafalia and in carbonaceous clay beds of the Baker Hill Formation. The Nanafalia Formation in eastern Alabama is placed in calcareous nannoplankton zones NP7-NP9 of the late Paleocene by Bybell (1980); thus a similar age for the Baker Hill is indicated (fig. 8).

Toulmin (1977) considered the Gravel Creek Sand Member of the Nanafalia Formation, which occurs in sinkholes in the underlying Clayton Formation, to expand updip to include the entire thickness of what herein is called the Baker Hill Formation. Marsalis and Friddell (1975) suggested that the Gravel Creek occurred only in the sinkholes and was not present updip. The burrowed surface at the base of the "Ostrea thirsae beds" of the Nanafalia Formation where they overlie the Gravel Creek Sand Member in sinkholes, as at Franklin Landing near Fort Gaines, suggests that a time gap exists between the two units. Evidence from spore-pollen and dinoflagellate assemblages suggests that the Gravel Creek beds along the Chattahoochee

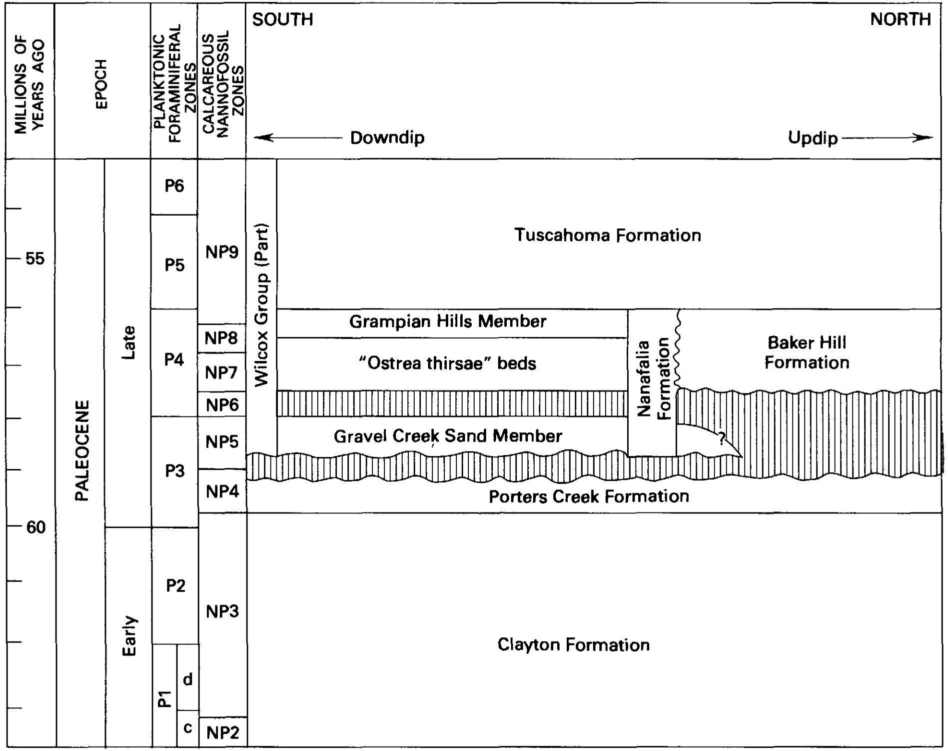


FIGURE 8. – Age relations of units of late Paleocene and early Eocene age in eastern Alabama and western Georgia. Patterned area indicates nondeposition. Epoch and zonal placements from Vail and Mitchum (1979).

River correlate with at least part of the Naheola Formation of western Alabama (Frederiksen, 1980; L. E. Edwards, oral commun., 1981), the upper part of which belongs to calcareous nannofossil zone NP5 (L. M. Bybell, oral commun., 1981) (fig. 8). Thus, the Gravel Creek beds are considerably older than the Baker Hill Formation and the "Ostrea thirsae beds" of the Nanafalia. These relationships strongly support the idea of Marsalis and Friddell that the Gravel Creek is present only in the downdip sinkholes. The biostratigraphic information suggests that it is a remnant of an older depositional cycle than that represented by the "Ostrea thirsae beds", the Grampian Hills Member, and the Baker Hill Formation.

**Environments of deposition**

Environmental interpretations based on geographic distribution of the sedimentological characteristics and on paleontologic information suggest that the Baker Hill Formation was deposited in fluvial to estuarine environments separated by barrier bars from the shallow, inner neritic environments in which the Nanafalia Formation was deposited.

A freshwater origin is postulated for the more updip deposits. This interpretation is indicated by the occurrence of pollen and spores only, dinoflagellates being absent (Gibson, Edwards, and Frederiksen, 1980). However, the carbonaceous clay and lignite in the more downdip Griffin mine yielded a low diversity dinoflagellate assemblage suggestive of brackish rather than freshwater environments of deposition (Gibson, Edwards, and Frederiksen, 1980). Samples from several carbonaceous clay intervals 1 mile north of Baker Hill, about 7 miles updip from the Griffin mine, yielded only pollen and spores and are considered to be of freshwater origin. The limited data seem to indicate that, at times, a transition from brackish to freshwater occurred in the area around Baker Hill. Additional evidence for brackish water environments in the more downdip areas of the Baker Hill Formation is the occurrence of Ophiomorpha, a noded burrow structure generally attributed to a callianassid shrimp. This burrow is considered by most workers to indicate brackish to marine waters, probably of at least 15 to 20 parts per 1,000 salinity or greater (R. W. Frey, oral commun., 1981). These burrows are found only in the downdip parts of the formation, generally within 5 miles of the bar sands, such as those found at corehole 124 (fig. 6-A). Here, in the most downdip part of the Baker Hill Formation, the section is composed almost entirely of sand, commonly gravelly and crossbedded and containing some clay clasts (figs. 6-B, 9). This area is interpreted as the barrier bar complex that allowed ponding of clay in estuarine and fluvial environments located to the north. Within 2 miles downdip (southeastward) from this corehole, equivalent strata are the shelly, glauconitic sand of the Nanafalia Formation as seen at Franklin Landing and other adjacent river sections at Fort Gaines (see River section, fig. 6-B).

**Type locality**

Lynn Griffin mine #1, Henry County, Ala., 5.2 miles southwest of Baker Hill, Ala., on west side of County Highway 29, 0.2 miles south of Barbour County-Henry County line, NE 1/4 sec. 4, T. 8 N., R. 27 E. (Lawrenceville, Alabama, 7 1/2-min quadrangle); exposes middle and upper parts of the formation and contact with the overlying Tuscahoma Formation.

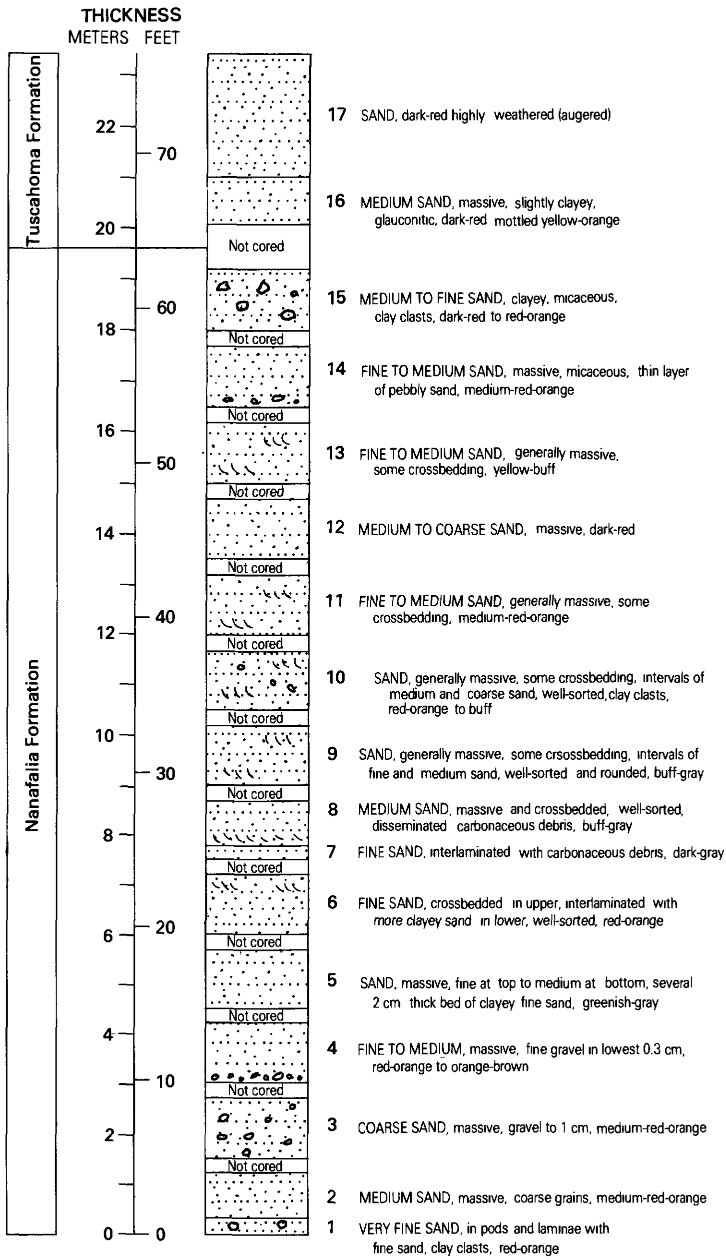


FIGURE 9.—Corehole section at Highland Park, Henry County, Ala. (124 on fig. 6*B*), showing sand and gravel lithologies of the farthest downdip, Baker Hill Formation. Top of section 232 ft (70.7 m).

## Reference localities

Two Mathison mines, Henry County, Ala., 1 mile north-northeast of Screamer, Ala., 200 and 400 yards north of County Highway 57, NW 1/4 sec. 13, T. 8 N., R. 28 E. (Fort Gaines NW, Alabama, 7 1/2-min quadrangle); exposes middle and upper parts of the formation and contact with the overlying Tuscahoma Formation.

Large abandoned mine, Henry County, Ala., in SE 1/4 sec. 5, T. 8 N., R. 29 E., immediately east of County Highway 93 (Fort Gaines NW, Alabama, 7 1/2-min quadrangle); exposes middle and upper part of the formation and contact with the overlying Tuscahoma Formation.

Roadcut, Barbour County, Ala., on north side of County Highway 24, 4.1 miles northwest of Baker Hill, Ala., SE 1/4 sec. 20, T. 10 N., R. 27 E. (Baker Hill, Alabama, 7 1/2-min quadrangle); exposes lower part of the formation and contact with the underlying Porters Creek Formation.

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# REVISION OF THE HATCHETIGBEE AND BASHI FORMATIONS (LOWER EOCENE) IN THE EASTERN GULF COASTAL PLAIN

By Thomas G. Gibson

## INTRODUCTION

Mapping in eastern Alabama and western Georgia, combined with biostratigraphic studies in this area and in western Alabama, has led to new interpretations of the stratigraphic relationship of the Hatchetigbee Formation. The Hatchetigbee Formation usually has been subdivided into a basal Bashi Marl Member that is overlain by an "unnamed upper member" (Toulmin, 1977). As shown by Gibson and Bybell (1981), the entire "unnamed upper member" is an updip facies equivalent of the Bashi Marl Member, and both represent an equal and relatively short time interval in the early Eocene. Because only the "unnamed upper member" occurs in the more updip areas and only the Bashi Marl Member occurs over the most downdip with an intervening large area containing one or more tongues of each unit, each of these units should be treated as coeval formations. This usage eliminates an informal stratigraphic name (the "unnamed upper member") and clarifies the previous concepts that the "unnamed upper member" represents a considerably greater time range than the Bashi Marl Member, which it does not, and that its lithologies occur above the "marly" lithologies of the Bashi, which is not always the situation. Each formation, as revised, extends at least from eastern Mississippi into western Georgia.

## HATCHETIGBEE FORMATION (HERE REDEFINED)

### Stratigraphy

The Hatchetigbee Formation, as redefined, includes the various lithologies that have been placed in its "unnamed upper member" and also the lithologies termed Hatchetigbee by some authors to separate mainly noncalcareous strata from its Bashi Marl Member. The characteristic lithologies of the Hatchetigbee include (1) massively bedded very fine to fine-grained quartz sand that is well sorted and contains little or no glauconite, (2) crossbedded fine- to rarely medium grained sand, and (3) interlaminated sequences of very fine grained sand, silt, and clay. Carbonaceous debris is abundant in the interlaminated sequences in western Alabama (Scott, 1972), but it is much less abundant in these sequences in eastern Alabama and western Georgia. Glauconite occurs in some beds of the formation but generally composes less than 2 percent. Indistinct lamination is present in some updip localities of the massively bedded sand, but the lamination is even more diffused by bioturbation in more downdip exposures. Scattered molluskan molds may be present, and occasionally silicified shells are found in thin lenticular pockets several inches in thickness and several to 10 feet in length. Also present is "sawdust sand," used in the sense of Pryor and Vanwie (1971), which is composed of silt and clay particles that have been flocculated and (or) rolled together into medium sand-sized grains. Sand units in the most updip areas in eastern Alabama commonly are crossbedded; the crossbeds may be undirectional and large scale, up to 3 ft in height, or they may be smaller scale and trough

U.S. Geol. Survey Bull. 1529-H, 1982, p. H33-H41

crossbedded with several directions present. The laminae in lithology 3 may be parallel and continuous but commonly are discontinuous with a length of several inches to 1 to 2 ft. Clay also may occur as flaser bedding within the laminated sequences and as thin clay drapes on trough crossbedded strata.

The type section for the Hatchetigbee Formation is at Hatchetigbee Bluff on the Tombigbee River (fig. 10); here, the Hatchetigbee anticline brings the unit up to the surface a considerable distance downdip from most outcrops of the formation, and the exposure exhibits the interfingering of the Bashi Formation (here raised in rank) and the Hatchetigbee. Beds 2, 3, 5, 6, 8, and 9 of this section as described by Smith, Johnson, and Langdon (1894) belong to the Hatchetigbee; beds 4, 7, and 10 are placed in the Bashi. Reference sections at Yellow Bluff and Woods Bluff (fig. 10) also are contained in Smith, Johnson, and Langdon (1894). Reference sections in eastern Alabama and western Georgia are contained herein; these consist of one complete section and several partial sections and include the contacts with the underlying and overlying formations.

### Contact relations

The Hatchetigbee Formation is underlain by the Tuscahoma Formation or the Bashi Formation. In western Alabama and in some downdip exposures in eastern Alabama, sand or interlaminated silt and clay beds of the lowermost part of the Hatchetigbee conformably rest upon highly glauconitic and fossiliferous fine-grained sand of the Bashi. In updip exposures in eastern Alabama, fine- to medium-grained crossbedded sand composing the basal beds of the Hatchetigbee rests upon interlaminated silt and clay of the underlying Tuscahoma Formation. At localities in eastern Alabama and western Georgia where much of the Hatchetigbee Formation is composed of interlaminated sand and clay, a 1-foot- to several-foot-thick sand bed commonly marks the base. The lower laminated beds of the Hatchetigbee generally are distinguished by discontinuous laminae containing flaser bedding, whereas the underlying Tuscahoma has planar continuous laminae. The contact between the Hatchetigbee and Tuscahoma Formations is a marine planation surface with essentially no relief.

Unconformably overlying the Hatchetigbee is either the Meridian Sand Member of the Tallahatta Formation or beds higher in the Tallahatta. In western Alabama, the Meridian Sand Member generally overlies the interlaminated silt and clay of the uppermost part of the Hatchetigbee; where the Meridian is not present, siliceous claystones of the Tallahatta rest upon the top of the Hatchetigbee. In eastern Alabama, a basal sand unit of the Tallahatta overlies the Hatchetigbee. The contact usually is a channeled surface with 5-20 ft of relief. The basal sand of the Tallahatta here is gravelly, is medium to coarse grained, is commonly crossbedded, and contains clay clasts as large as 3 to 4 in in diameter. The underlying Hatchetigbee is a massively bedded very fine grained sand showing indistinct lamination and moderate bioturbation fabric.

### Geographic extent and thickness

The Hatchetigbee Formation has been recognized in outcrops as far east as Plains, Sumter County, Ga. (fig. 10), where it is an interlaminated sequence approximately 20 ft thick. The laminated facies dominates updip

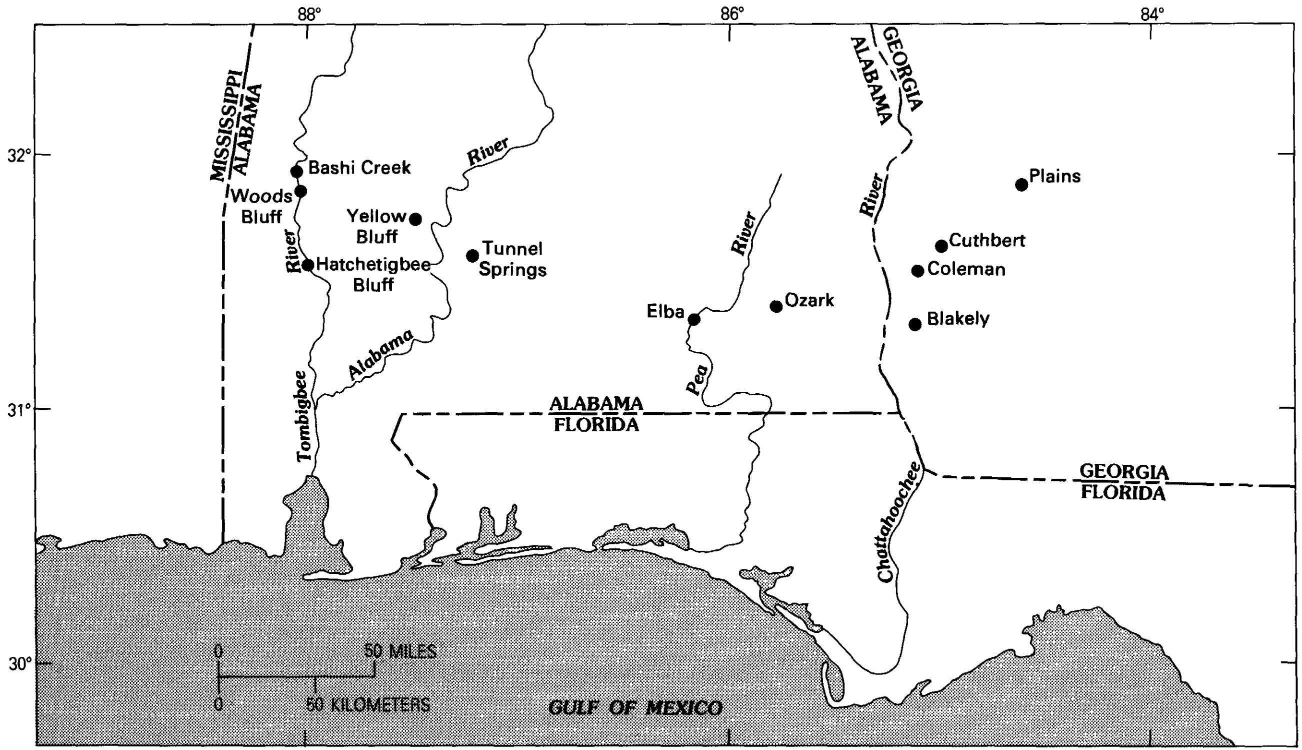


FIGURE 10.–Locations of Hatchetigbee and Bashi sections, eastern Gulf Coastal Plain.

outcrops westward through Cuthbert, Randolph County, Ga. (fig. 10). In the more downdip exposures and coreholes in westernmost Georgia and eastern Alabama, the Hatchetigbee is mostly massively bedded sand with lesser amounts of discontinuously laminated sand and clay and massively bedded clay.

Sand with interlaminated silt and clay continues across central Alabama, as seen at localities along the Pea River south of Elba in Coffee County and at Ozark in Dale County (fig. 10). In these sections, the interfingering of the Hatchetigbee with the Bashi is conspicuous. The interfingering of the two formations is noticeable also in the more downdip localities in western Alabama, particularly at the type locality at Hatchetigbee Bluff (fig. 10). The updip localities in northern Monroe and Wilcox Counties are dominated by thick sequences of interlaminated sand, silt, and clay beds, often highly carbonaceous, of the Hatchetigbee with only a thin tongue of the Bashi at the base. The Hatchetigbee Formation has been recognized westward into Winston County, Miss. (Mellen, 1939).

In western Georgia, the Hatchetigbee Formation is from 20-30 ft thick updip but thins downdip to only several feet thick with the remainder of the sections consisting of the Bashi Formation. In the Chattahoochee River area, the thickness is 56 ft updip and thins downdip to 1 ft or less as shown by Gibson and Bybell (1981). This westward thickening trend continues across Ala. Sixty-five ft of laminated beds and sand of the Hatchetigbee Formation occur along the Pea River (Smith, Johnson, and Langdon, 1894), and the section thickens to greater than 200 ft in west-central and western Alabama as seen in Monroe County (Scott, 1972) and Choctaw County (Mancini, 1981).

### Age

As shown by Gibson and Bybell (1981), the "unnamed upper member" of the Hatchetigbee is an updip facies and time equivalent of its more downdip former Bashi Marl Member, herein raised in rank to Bashi Formation. Palynomorphs in the revised Hatchetigbee Formation establish its age as early Eocene (N. O. Frederiksen, in Reinhardt and Gibson, 1980, p. 422-423), but the absence of calcareous nannofossils and planktonic foraminifers does not allow a more precise placement for these beds. However, the time equivalency with the well-dated Bashi Formation gives the age of the Hatchetigbee. The age of the Bashi is in the lower part of calcareous nannoplankton zone NP10 and middle of planktonic foraminiferal zone P6 (Gibson and Bybell, 1981), placing the Hatchetigbee also in the earliest 500,000 years of the Eocene (fig. 11).

### Environments of deposition

The Hatchetigbee Formation represents deposition in estuarine, tidal flat, shoreface, and very shallow inner neritic environments during the marine transgressive cycle in which the Bashi Formation was deposited in inner to middle neritic environments to the south. The sediments of the Hatchetigbee are interpreted as having been deposited in shallow marine and marginal marine environments. The massively bedded sand that dominates the downdip exposures in eastern Alabama and western Georgia and occurs less frequently in western Alabama is considered to have been deposited in

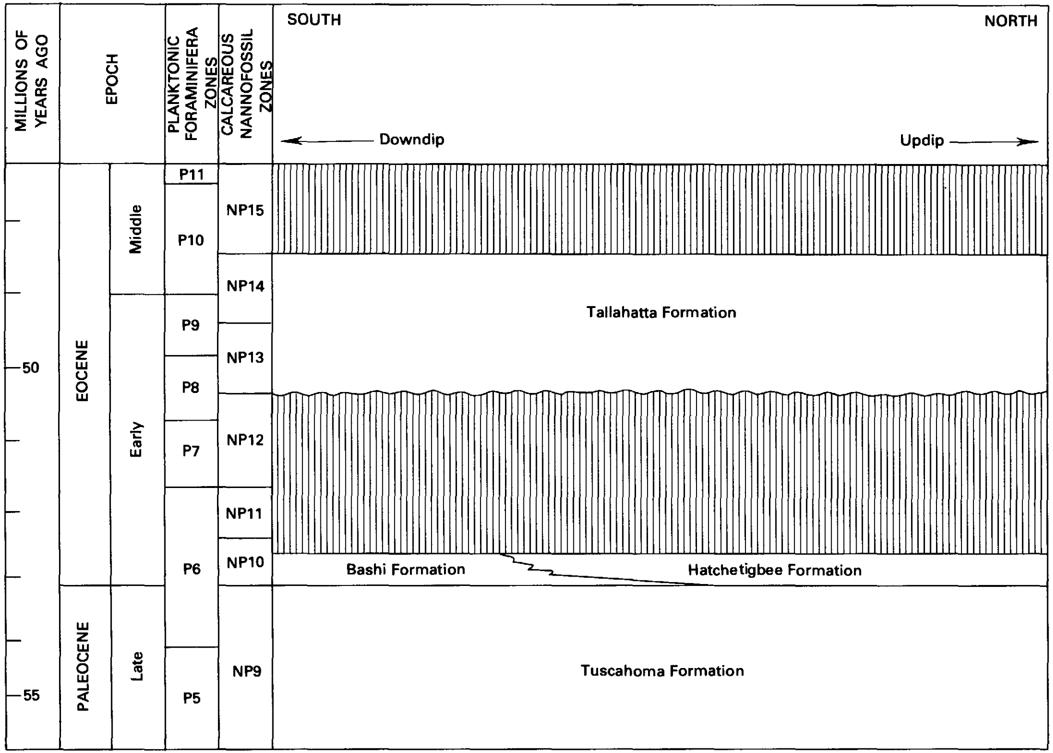


FIGURE 11.—Contact relations of Hatchetigbee and Bashi Formations, eastern Gulf Coastal Plain. Patterned area indicates nondeposition. Biostratigraphic zonation from Gibson and Bybell (1981); ages from Vail and Mitchum (1979).

shallow inner neritic environments. The "sawdust sand" and the discontinuously laminated strata containing flaser bedding occur updip from the massively bedded sand and are interpreted as having been formed in shallow marine environments, ranging from very shallow neritic through lower shoreface to tidal flat. The highly crossbedded sand, characteristic of the farthest updip basal sequences, particularly in eastern Alabama, suggests a variety of shallow environments including offshore bars, tidal inlets, and possibly nearshore bars. The continuously laminated beds, which are commonly carbonaceous, are found throughout the geographic range of the Hatchetigbee but are dominant especially in western Alabama and suggest deposition in various tidal flat, lagoonal, and deltaic environments.

### Reference sections

In addition to the type and reference sections in western Alabama mentioned previously, the following reference sections are included for the Hatchetigbee in eastern Alabama and western Georgia where it has not been as commonly recognized.

In Barbour County, Ala., a roadcut on north side of County Highway 57, 100 yd east of intersection with U.S. Highway 431, sec. 29, T. 9 N., R. 28 E. (Lawrenceville, Alabama, 7 1/2-min quadrangle); exposes lower beds of Hatchetigbee Formation and contact with underlying Tuscahoma Formation.

In Henry County, Ala., the west face of Mathison mine, 1 mi east of Screamer, 100 yd north of County Highway 57, NW 1/4 sec. 13, T. 8 N., R. 28 E. (Fort Gaines NW, Alabama, 7 1/2-min quadrangle); exposes lower beds of Hatchetigbee Formation and contact with underlying Tuscahoma Formation.

In Randolph County, Ga., a roadcut on south side of County Highway 152, 3.1 mi northeast of Cuthbert, 100 yd southwest of Pachitla Creek; exposes middle and upper parts of Hatchetigbee Formation and contact with overlying Tallahatta Formation.

In Dale County, Ala., a roadcut on both sides of road, 3/4 mi north of Bells Crossroads, NE 1/4 sec. 5, T. 6 N., R. 26 E. (Clopton, Alabamam 7 1/2-min quadrangle); exposes complete Hatchetigbee Formation and contacts with underlying Tuscahoma and overlying Tallahatta Formations.

## BASHI FORMATION (HERE RAISED IN RANK)

### Stratigraphy

The Bashi Formation, here raised in rank, consists of the lithologies previously called the Bashi Marl Member of the Hatchetigbee Formation. The general sedimentary nature of the formation is of a fine- to very fine-grained sand, often clayey, silty, and calcareous, massively bedded, with abundant glauconite (percentages commonly of 10-25 percent), and with abundant calcareous fossils, commonly in relatively densely packed layers. Rounded calcareous concretions are common. In the farthest downdip exposures, as along the Chattahoochee River, the formation becomes a clayey silt to very fine-grained sand, and shells are more disseminated throughout the beds.

The type section for the Bashi Formation is the type of the former Bashi Marl Member, along the banks of Bashi Creek, northwestern Clarke County (fig. 10). A description of this outcrop is found in Smith, Johnson, and Langdon (1894). Other reference sections in western Alabama are at Hatchetigbee Bluff and Woods Bluff (Smith, Johnson, and Langdon, 1894), where the Bashi interfingers with the Hatchetigbee Formation, and at Tunnel Springs (Scott, 1972) where the upper part is exposed (fig. 10). Farther east in Alabama, good exposures are present at the dam site on the Pea River south of Elba (lower part of formation) and in the railroad cut north of Ozark where the upper part is exposed. The entire formation was exposed along the Chattahoochee River northwest of Blakely, Ga. (Toulmin and LaMoreaux, 1963), although much of the section is now covered by slumps. A corehole at Hutchins Landing on the Chattahoochee River northwest of Blakely is given in Gibson and Bybell (1981) (see locality 99, figs. 2 and 3). The farthest known eastward outcrop is at Greens Branch near Coleman, Ga. (fig. 10); a description of this exposure is given by Gibson (in Reinhardt and Gibson, 1980).

## Contact relations

The Bashi Formation is usually underlain by the Tuscahoma Formation in Alabama and westernmost Georgia. The carbonaceous interlaminated silt and clay beds of the Tuscahoma are easily differentiated from the fossiliferous glauconitic sand of the Bashi.

In western and central Alabama, the Bashi usually is overlain conformably by the Hatchetigbee Formation; there may be only one thin to thick tongue of the Hatchetigbee that overlies the Bashi or there may be an overlying tongue and several lower tongues that interfinger with the Bashi. In some places in western Alabama, the Tallahatta Formation overlies the Bashi, and here the basal Tallahatta deposits consist of siliceous claystone or, less frequently, of medium- to coarse-grained sand. In eastern Alabama, shelly medium-grained sand of the Tallahatta Formation overlies the clayey silt of the Bashi in downdip exposures. The contact is usually straight because of marine planation by the transgressing Tallahatta sea.

## Geographic extent and thickness

The Bashi Formation has been recognized in surface outcrops from Coleman, Ga., on the east (fig. 10), westward across Alabama and into eastern Mississippi. In updip areas where the Bashi Formation is largely replaced by the Hatchetigbee, thicknesses of 5 to 10 ft for the Bashi are common. The thickness of the Bashi increases downdip and reaches 35 feet along the Chattahoochee River where the Bashi comprises the entire section. Similar thicknesses are found in Alabama in the more downdip localities as at Ozark, along the Pea River south of Elba, and at Hatchetigbee Bluff (fig. 10); at these localities, the Bashi still has some interfingering with the Hatchetigbee.

## Age

The Bashi Formation is of earliest Eocene age, belonging to the lower portion of calcareous nannoplankton zone NP10 and the middle of planktonic foraminiferal zone P6 (Gibson and Bybell, 1981). This places the formation in the first 500,000 years of the Eocene (fig. 11).

**Environments of deposition**

The Bashi Formation is interpreted as having been deposited in inner to middle neritic environments. The more updip localities, especially those in which it interfingers with the Hatchetigbee Formation, contain deposits from inner neritic environments. The more downdip localities, best exemplified by the exposures and corehole along the Chattahoochee River northwest of Blakely, Ga., (Gibson and Bybell, 1981), probably approach middle neritic environments of deposition.

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# FORMATION NAMES IN THE WORCESTER AREA, MASSACHUSETTS

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## ABSTRACT

The Worcester Formation of Emerson (1917) on the east flank of the Merrimack synclinorium, east-central Massachusetts, is redefined, and the fossiliferous rocks of Middle Pennsylvanian age at Worcester are here renamed the Coal Mine Brook Formation. Other rock units formerly included in Emerson's Worcester Formation and in his Brimfield Schist, Oakdale Quartzite, and Paxton Quartz Schist are renamed or redefined and reassigned new ages on the basis of recent quadrangle mapping, radiometric age determinations of intrusive rocks, and regional correlation. Changes also have been made in the nomenclature of the rocks intrusive into the mostly Silurian section. These rocks are the Ayer Granite, Chelmsford Granite, Fitchburg Complex, Newburyport Complex, and Dracut and Exeter Diorites.

## INTRODUCTION

Recent mapping in east-central Massachusetts by geologists of the U.S. Geological Survey in cooperation with the Massachusetts Department of Public Works and by faculty and graduate students from Harvard University, Boston College, and the University of Massachusetts, has led to a better understanding of the stratigraphic and structural relations in the Worcester area of Massachusetts.

In addition, a number of people, principally R. H. Zartman of the U.S. Geological Survey and R. E. Naylor of Northeastern University, have recently made radiometric age determinations on intrusive and volcanic rocks in the area. These determinations have enabled a closer approximation of the age of the layered rocks. The purpose of this paper is to redefine and rename some of the stratigraphic units in the Worcester area and to reassign ages to these units on the basis of the above work. The units discussed are all on the east flank of the Merrimack synclinorium and lie in or west of the Clinton-Newbury fault zone (fig. 12).

Redefinition of many of the stratigraphic units in the Worcester area depends on prior redefinition of the Worcester Formation. Hence, units in this paper will not be discussed in order of decreasing age but will be discussed as names develop from the redefined Worcester. Layered rock units renamed and redefined fall into eastern and western sectors separated by the Wekepeke fault. Because correlation across this fault cannot be made with certainty, layered rock units in the two sectors are discussed separately.

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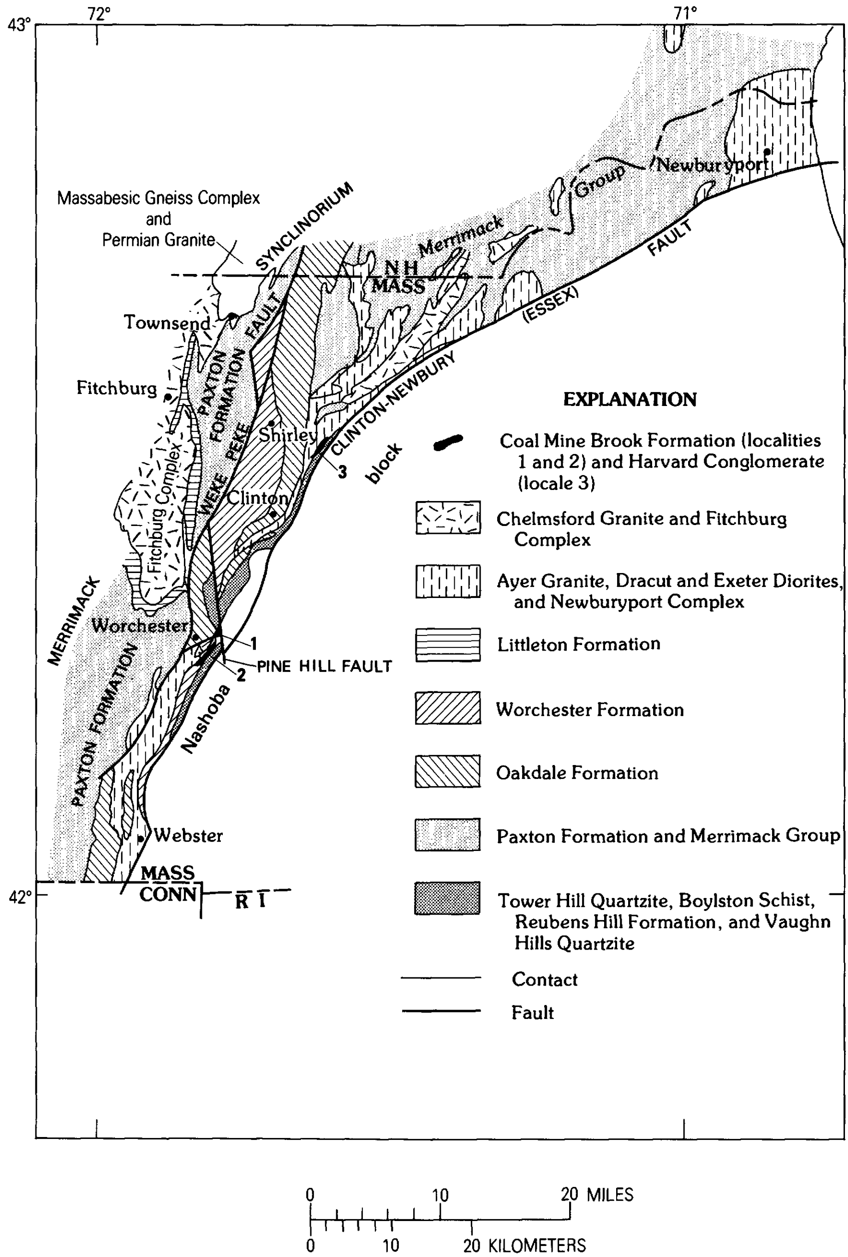


FIGURE 12.—Generalized geology of the east flank of the Merrimack synclinorium, Massachusetts and New Hampshire.

## LAYERED ROCKS, EASTERN SECTOR

### Worcester Formation (here redefined)

The name "Worcester" was first used by Emerson (1889, p. 560; 1898, p. 17) to refer to an extensive belt of argillite-, phyllite-, and chiastolite-bearing mica schist in Worcester and Middlesex Counties, Mass. The formation was described in detail as the Worcester Phyllite by Perry and Emerson (1903) and Emerson (1917) and assigned a Carboniferous age on the basis of plant fossils found in coal-bearing beds at the Worcester "coal mine" (White, 1912). The Pennsylvanian age of the coal-bearing rocks has been recently confirmed by Grew and others (1970) and P. C. Lyons (in Grew, 1976, p. 395). However, Grew (1970, 1973, 1976) has shown that the "coal mine" strata containing the fossils is actually separated by faults and an unconformity from the rest of the pelitic rocks that Emerson called Worcester Phyllite.

Hansen (1956), mapping in the Hudson area, changed the name to Worcester Formation and redefined the unit to include other rock types in addition to phyllite (fig. 13). The belt of Hansen's Worcester Formation in the Hudson quadrangle is separated by the Oakdale Formation from the main belt of Worcester Phyllite as mapped by Emerson (fig. 12) and lies within a fault zone (Skehan, 1968; Peck, 1976; Gore, 1976b).

The Worcester Formation is here redefined as the slate, phyllite, and metasiltstone coinciding approximately with Emerson's (1898) belt of Worcester Phyllite extending from the Worcester area northward into the towns of Clinton and Shirley, excluding the fossiliferous strata and related rocks of Pennsylvanian age mapped by Grew at Worcester. As so redefined, the Worcester no longer includes the mica schist facies, the Vaughn Hills Member, the phyllite facies, and the Harvard Conglomerate Lentil of Hansen (1956) (see fig. 13) and is geographically and stratigraphically restricted to exclude those units.

The Worcester Formation thus redefined is equivalent to units 3 and 4 of Peck (1976) (see fig. 13) and the eastern belt of Grew's (1970) Holden Formation (unit De of Grew, 1973). Its type area would be in the Clinton quadrangle where the rocks were described in detail by Peck (1976). As redefined, the formation north of Worcester lies east of the western belt of the Oakdale Quartzite of Emerson (1917; Peck's unit 5) and west of the eastern belt of Emerson's Oakdale Quartzite, equivalent to Peck's unit 2 (not shown on fig. 12, south of Worcester). The Worcester Formation is offset by the Pine Hill fault (Grew, 1970) at Worcester and is cut off by faults to the north near the Massachusetts-New Hampshire border and to the south in the Webster area near the Massachusetts-Connecticut border. Peck (1976, p. 244) (see fig. 13, units 3 and 4) estimated the thickness to be between 3,050 and 4,270 m in the Clinton quadrangle. Grew (1970, p. 157) estimated the thickness of his "Holden Formation" to be greater than 3,000 m in the Worcester area.

In the Worcester area, the Worcester Formation is in an apparent conformable sequence with the Oakdale Quartzite of Emerson (Peck, 1976; Grew, 1970), which is considered to be a probable Silurian age. The Worcester is intruded by granite of the Fitchburg Complex, which has an

EASTERN SECTION

| AGE | EMERSON (1917) | HANSEN (1956) | SKEHAN (1967) | GREW (1970) | PECK (1975, 1976) | THIS PAPER | |
|---|---|---|---|---|---|---|---|
| PENNSYLVANIAN | Worcester Phyllite | Worcester Formation<br>Harvard Conglomerate Lentil of Worcester Formation | | Worcester Formation<br>Harvard Conglomerate<br>UNCONFORMITY(?) | | Coal Mine Brook Formation<br>Harvard Conglomerate<br>UNCONFORMITY(?) | |
| EARLY DEVONIAN AND SILURIAN | Worcester Phyllite | | Oakdale and Worcester Formations | Holden Formation (part) | Units 3 and 4 | Worcester Formation | |
| SILURIAN | Oakdale Quartzite Worcester Phyllite | | | Oakdale Formation (part) | Unit 2 | Oakdale Formation | Merrimack Group |
| | Oakdale Quartzite | | | Tower Hill Quartzite Member of the Boylston Formation | Unit 1 | Tower Hill Quartzite | |
| SILURIAN OR ORDOVICIAN | Boylston Schist Worcester Phyllite | Worcester Formation (Phyllite facies) | FAULT | Boylston Formation | FAULT | Boylston Schist<br>FAULT | |
| | | | Reubens Hill amphibolite | | Reubens Hill Igneous Complex | Reubens Hill Formation | ? |
| | | Vaughn Hills Member of Worcester Formation | Vaughn Hills Formation<br>UNCONFORMITY(?) | | Vaughn Hills Member of the Tadmuck Brook Schist | Vaughn Hills Quartzite | ? |
| SILURIAN(?)<br>?<br>ORDOVICIAN OR PROTEROZOIC Z | Brimfield Schist | Worcester Formation (mica schist facies)<br>Nashoba Formation | Nashoba Formation | | Tadmuck Brook Schist | Tadmuck Brook Schist<br>UNCONFORMITY(?)<br>Nashoba Formation | |

WESTERN SECTION

| AGE | EMERSON (1917) | GREW (1970) | PECK (1976) | PEASE (1972), PEPER AND OTHERS (1976) | THIS PAPER | | |
|---|---|---|---|---|---|---|---|
| EARLY DEVONIAN | Brimfield Schist Worcester Phyllite | Holden Formation (part) | | | Littleton Formation | | |
| SILURIAN | Oakdale Quartzite Paxton Quartz Schist | Oakdale Formation (part) | UNIT 5 | Southbridge Formation of "Paxton Group"<br>FAULT<br>Bigelow Brook Formation of Brimfield Group | Southbridge Member<br>Bigelow Brook Member | Paxton Formation | Merrimack Group |

FIGURE 13.—Layered-rock units in the east flank of the Merrimack synclinorium, Massachusetts and New Hampshire. Query indicates equivalent units not recognized in the Merrimack Group.

Early Devonian age (R. E. Zartman and R. S. Naylor, written commun., 1978). In composition and structural position, the Worcester is similar to the Lower Devonian Shapleigh Group of Hussey (1968) in southern Maine. Hussey correlated this group with the Littleton Formation of Early Devonian age in its type area in New Hampshire. Robinson (1981) has evidence from graded bedding, however, that indicates that the Worcester underlies the Oakdale. Because the Worcester is in a stratigraphic sequence with probable Silurian rocks and is intruded by the Lower Devonian Fitchburg Complex, the Worcester is most likely Silurian to Early Devonian in age. However, the stratigraphic sequence that includes the Worcester may be older than Silurian and Devonian if some radiometric ages obtained from the Ayer Granite and Newburyport Complex are correct (see below). The Worcester is shown on figure 13, however, in its structural and possible stratigraphic position above the Oakdale, rather than in a possible stratigraphic position beneath the Oakdale. On the basis of the evidence presented above, the Worcester is assigned a Silurian and probable Early Devonian age.

### Coal Mine Brook Formation (here named)

The fossiliferous strata of Middle Pennsylvanian age at the Worcester coal mine on the property of Notre Dame Institute, Worcester (loc. 1, fig. 12), and garnetiferous phyllite, arkose, and conglomerate containing pebbles of Devonian granite near Franklin and Shrewsbury Streets, Worcester (loc. 2, fig. 12), described by Grew (1970, 1973, 1976) are herein named the Coal Mine Brook Formation after the name of the stream near the coal mine. According to Lyons and others (1976), the Coal Mine Brook Formation is correlative with the lower part of the Rhode Island Formation of Middle and Late Pennsylvanian age in the Narragansett basin on the basis of fossil flora (Westphalian C of the Canadian maritime provinces and Europe) found at the coal mine. Grew (1973) gives minimum thicknesses of the Coal Mine Brook as 330 m of conglomerate, arkose, and phyllite in downtown Worcester and 50 m of slate and phyllite at the coal mine, the composite type localities. The Coal Mine Brook Formation at the mine is less metamorphosed than are the adjacent regionally metamorphosed Oakdale and Worcester Formations.

### Harvard Conglomerate (here redefined)

The nonfossiliferous Harvard Conglomerate (Crosby, 1876) considered by Emerson (1917) and Hansen (1956) to be a lentil in their Worcester Phyllite or Formation (fig. 13) is located northeast of the Coal Mine Brook Formation (loc. 3, fig. 12) in the same northeast-trending belt as the Coal Mine Brook and is involved in the same fault system (Skehan, 1968; Grew, 1976). The Harvard also has been considered to be Pennsylvanian in age (Hansen, 1956; Grew, 1970, 1973, 1976; Thompson and Robinson, 1976). It rests unconformably on the Ayer Granite at Pin Hill, near Harvard, Mass., providing a lower limit for the age of the Harvard. Some uncertainty exists as to the age of the Ayer at this locality; the Ayer could be Silurian or Late Ordovician according to R. E. Zartman and R. S. Naylor (written commun., 1978). The Harvard Conglomerate remains a valid name coequal in rank with the Coal Mine Brook Formation, but the age of the Harvard is considered to be probably Pennsylvanian.

**Oakdale Formation**

The Oakdale Quartzite of Emerson (1917) is a calcareous metasiltstone and pelite that is part of a sequence described by Peck (1976), Grew (1970, 1973), and Robinson (1978) in the Clinton, Worcester, and Shirley area that also includes the Worcester Formation and orthoquartzite (Tower Hill Quartzite, Vaughn Hills Quartzite). This sequence is interpreted by Peck (1976) as a mostly distal-turbidite deposit. The Oakdale of Emerson is separated by the Wekepeke fault (Novotny, 1961; Peck, 1975; Pine Hill fault of Castle and others, 1976) into east and west belts. West of the Wekepeke fault, Emerson's Oakdale is equivalent to unit 5 of Peck; east of the Wekepeke fault, his Oakdale is equivalent to unit 2 of Peck and unit DSd of Grew (1973). Following Grew (1970) and Pease (1981), Emerson's Oakdale Quartzite is renamed the Oakdale Formation because it is not a quartzite.

The stratigraphic order of the units in the turbidite section and their equivalency to other units in the region have been difficult to establish because of faulting and thrusting. The sequence is separated from the Ordovician or Proterozoic Z Nashoba Formation to the east by the Clinton-Newbury fault zone, but some evidence suggests that the section lies unconformably on the Nashoba (Alvord and others, 1976). The Oakdale is continuous with the Merrimack Group of southeastern New Hampshire and southern Maine (Hitchcock, 1877; Katz, 1917; Billings, 1956; Hussey, 1968), which has been correlated with fossiliferous strata of Silurian age in central Main (Osberg, 1968). The Oakdale is also continuous with part of the Hebron Formation of Connecticut (Rodgers and others, 1959). Both the Oakdale and the Hebron are intruded by granite dated as Silurian (R. E. Zartman and R. S. Naylor, written commun., 1978); and, as Dixon (1976, p. 282) has pointed out, the Hebron (and Oakdale) could be older than Silurian. The Oakdale Formation is assigned a probable Silurian age.

**Tower Hill Quartzite (name adopted)**

Orthoquartzite with interlayered pelite lying between the Oakdale Formation and the Boylston Schist (unit B of Grew, 1973) was informally named the Tower Hill Quartzite Member of the Boylston Formation by Grew (1970), and this name is formally adopted as the Tower Hill Quartzite for use in the Worcester area. The Tower Hill is assigned a probable Silurian age based on the tentative correlation of this turbidite section with the rocks in central Maine. G. R. Robinson, Jr., one of the authors, believes the Tower Hill represents a proximal facies laterally equivalent to the distal deposits forming the Worcester Formation.

**Vaughn Hills Quartzite (here redefined)**

A quartzite with associated pelite, conglomerate, and chlorite schist that appears to lie at the base of the turbidite section and that may or may not correlate with the Tower Hill Quartzite is the Vaughn Hills Member of the Worcester Formation of Hansen (1956). Peck (1975) considered the Vaughn Hills to be the upper member of the Tadmuck Brook Schist (Bell and Alvord, 1976), which lies on top of the Nashoba Formation east of the Clinton-Newbury fault. The Vaughn Hills, raised to formation rank as Vaughn Hills Quartzite, is herein removed from the Worcester and assigned a probable Silurian or Ordovician age.

### Reubens Hill Formation (name adopted)

The Reubens Hill Formation was named by Skehan (1967), who applied the name to metavolcanic rocks overlying the Vaughn Hills Quartzite and Nashoba Formation of Ordovician or Proterozoic Z age and underlying the Oakdale-Worcester sequence in the Clinton-Newbury fault zone in the Wachusett-Marlborough tunnel. Peck (1975) changed the name to "Reubens Hill Igneous Complex" because it contained metadiorite. We prefer to use the simpler Reubens Hill Formation for this small unit. Its type area is at Reubens Hill east of the Wachusett Reservoir, Clinton quadrangle, Worcester County, Mass. In the tunnel, the Reubens Hill is 590 m thick. On the basis of its uncertain stratigraphic position, the Reubens Hill is assigned a Silurian or Ordovician age.

### Boylston Schist (here restricted)

The Boylston Schist of Emerson (1917) and Grew (1970) is restricted in areal usage to apply to partly graphitic and pyritic pelitic, psammitic, and calcarenitic rocks between the Oakdale Formation and the Ordovician or Proterozoic Z Nashoba Formation (fig. 12). It may be equivalent in part to the Tadmuck Brook Schist. The Boylston is probably Silurian or Ordovician.

### Merrimack Group

The Kittery Quartzite and Eliot and Berwick Formations of the Merrimack Group of New Hampshire and southern Maine (Katz, 1917; Billings, 1956; Hussey, 1968) are here recognized as valid units in northeastern Massachusetts. The stratigraphic order of the units is uncertain, but the Kittery appears to lie at the base of the group. The name Kittery Quartzite is changed to Kittery Formation following Novotny (1963) and Hussey (1968). Its age is Silurian or Ordovician because it is intruded by granodiorite of the Newburyport Complex dated as Silurian and Ordovican(?) (R. E. Zartman and R. S. Naylor, written commun., 1981). The Eliot and Berwick Formations remain as probable Silurian in age on the basis of tentative correlation with similar rocks in central Maine (Osberg, 1968).

## LAYERED ROCKS, WESTERN SECTOR

### Paxton Formation

Emerson's (1917) name Paxton Quartz Schist (Paxton Schist of Perry and Emerson, 1903) is herewith changed to Paxton Formation because the unit contains a variety of rock types although it is principally a calcareous metasiltstone. Emerson considered the Paxton and the Oakdale to grade into one another, and some believe the Paxton to be a higher metamorphic grade equivalent to the Oakdale (see Billings, 1956). Pease (1972) and Barosh (1976), on the other hand, have subdivided the Paxton into the Southbridge Formation and an unnamed unit equivalent to part of the Hebron Formation of Connecticut. The Southbridge is here considered to be a member of the Paxton Formation. Peter Robinson, (oral commun., 1978) mapping north of the type area of the Brimfield Schist near Brimfield, Mass., believes that the Bigelow Brook Formation of the Brimfield Group (Peper and others, 1976) is more appropriately a member of the Paxton Formation (fig. 13) because it projects into Emerson's type Paxton Quartz Schist. The Paxton is continuous with part of the Merrimack Group of southeastern New Hampshire and probably is equivalent to the Berwick Formation of that group. The Paxton Formation is assigned a probable Silurian age based on the correlation with the Merrimack Group (Berwick part). It is intruded by granite of Silurian age (R. E. Zartman and R. S. Naylor, written commun., 1978), however, and could be older than Silurian.

## INTRUSIVE ROCKS

### Ayer Granite

The Ayer Granite (Emerson, 1917), called the "Ayer crystalline complex" by Gore (1976a), has been divided by Gore into two facies: a gneissic, partly porphyritic biotite quartz monzonite (granite) to quartz diorite--the Devens-Long Pond facies--and a porphyritic biotite quartz monzonite (granite)--the Clinton facies. Earlier U-Pb isotopic age determinations had shown that the Devens-Long Pond facies could not be dated reliably (419 m.y.-462 m.y.) (R. E. Zartman and R. S. Naylor, written commun., 1978). Radiometric age determinations of the Clinton facies indicated an Early Silurian age (about 436 m.y.). These ages were used in compiling a new bedrock geologic map of Massachusetts (Zen and others, 1981). Re-evaluation of their earlier data and application of a new time scale has indicated a Late Ordovician or possibly an Early Silurian age (433 m.y. $\pm$ 5 m.y.) for both facies (R. E. Zartman, written commun., 1981).

### Chelmsford Granite

Gore (1976b) added a hydrous two-mica granite, the commercial "Chelmsford" of Currier (1937) and Currier and Jahns (1952) and the Chelmsford Granite of Lyons and Faul (1968), to his "Ayer crystalline complex." The Chelmsford is Early Devonian in age (about 383 m.y.) (R. E. Zartman and R. S. Naylor, written commun., 1978) and cuts the Devens-Long Pond facies of the Ayer. Although the Chelmsford and the Clinton facies of the Ayer might be related genetically as Gore suggests, we believe the Chelmsford should be a unit in its own right because of the age designation and composition, and it has thus been so adopted by Robinson (1978).

### Fitchburg Complex (here redefined)

The Fitchburg Granite (Emerson, 1917), mostly two-mica granite with subordinate granodiorite, has been subdivided by various workers into subunits including gneissic and nongneissic phases (Peper and Wilson, 1978; J. C. Hepburn, unpub. data, 1976) and as mapped may contain rocks of several ages (see Aleinikoff and others, 1980). The name is changed, therefore, to Fitchburg Complex. U-Pb and Rb-Sr radiometric determinations by R. E. Zartman and R. S. Naylor (about 395 m.y.-400 m.y.) (written commun., 1978) indicate that granite of the Fitchburg is Early Devonian or younger in age. Not all subunits have been dated, however.

### Massabesic Gneiss Complex (here adopted)

Pink microcline gneiss, white oligoclase gneiss, and amphibolite formerly included in the Fitchburg Granite (now Complex) pluton (Emerson, 1917; Billings and others, 1952) were separated out from the Fitchburg in the

Manchester, N.H., area by Sriramadas (1966) and called the Massabesic Gneiss. The Massabesic Gneiss, as currently mapped, is a mixed assortment of rocks that includes rock resembling Monson Gneiss of the Bronson Hill anticlinorium and rock resembling parts of the Nashoba Formation of eastern Massachusetts. The Massabesic has not been mapped in detail, however. U-Pb ages on zircons from the Massabesic range from 475 m.y. to 646 m.y. (Besancon and others, 1977; Aleinikoff and others, 1980). Its age is thus uncertain, and the Massabesic may contain rocks of Ordovician or Proterozoic Z age or both. It is herein renamed the Massabesic Gneiss Complex for U.S. Geological Survey usage. The Massabesic extends into the Townsend area, Massachusetts, north of the main mass of the Fitchburg Granite (fig. 12) in an area occupied also by granite of Permian age centered on Milford, N.H. (Aleinikoff, 1978; Aleinikoff and others, 1980).

### Newburyport Complex (here redefined)

The Newburyport Quartz Diorite of Emerson (1917) is associated with porphyritic quartz monzonite in the Newburyport area of Massachusetts (Novotny, 1969). Under the International Union of Geological Sciences (Streckeisen, 1973) classification, the quartz diorite is a tonalite and granodiorite, and the porphyritic quartz monzonite is a porphyritic granite. These two rocks are abruptly gradational into each other (Shride, 1971) and are here grouped as the renamed Newburyport Complex. Earlier U-Pb radiometric age determinations on zircons (R. E. Zartman and R. S. Naylor, written commun., 1978) indicated that the tonalite and granodiorite are Ordovician (about 465 m.y.) in age, and the porphyritic quartz monzonite is Silurian (about 437 m.y.) in age. These ages were used in compiling a new geologic map of Massachusetts (Zen and others, 1981). Reevaluation of the isotopic data from the porphyritic quartz monzonite by Zartman and Naylor (R. E. Zartman, written commun., 1981) has indicated an Ordovician age (455 m.y. $\pm$ 15 m.y.) for it. The Newburyport Complex is believed to intrude the Kittery Formation of the Merrimack Group. The age of the Kittery is unknown but could be either Silurian or Ordovician or both. The possibility exists, then, that some of the Newburyport is Silurian in age.

### Dracut and Exeter Diorites

Plutons of diorite, quartz diorite, granodiorite, and gabbro and norite associated with the Ayer Granite in east-central and northeastern Massachusetts and southeastern New Hampshire include the Dracut Diorite (Emerson, 1917) and Exeter Diorite (Hitchcock, 1877; Billings, 1956). All these rocks are assigned a Devonian and Silurian age on the basis of their spatial association with the Devonian Chelmsford and the probable Silurian Ayer Granites (see Sundeen, 1971; Robinson, 1978), but some masses could be older, as suggested by the uncertainty as to the ages of the tonalite and granodiorite of the Newburyport Complex and the Devens-Long Pond facies of the Ayer. Gaudette and others (1975) suggest that the Exeter is the same age as the older part of the "Hillsboro Plutonic Series" of southeastern New Hampshire and southwestern Maine, about 440 m.y., placing it in the Early Silurian or Late Ordovician.

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# STRATIGRAPHIC NAMES IN EASTERN MASSACHUSETTS AND ADJACENT STATES

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## ABSTRACT

Changes in name and age assignment and statement of geologic distribution for stratified and intrusive rock units in the Nashoba block of eastern Massachusetts and the Putnam block of eastern Connecticut and for stratified and intrusive rock units in the Milford-Dedham terrane of eastern Massachusetts and adjacent States are described and listed. These changes are necessary because of new radiometric age determinations on rocks in eastern Massachusetts, Rhode Island, and Connecticut and reconnaissance mapping in preparation for a new geologic map of Massachusetts.

## INTRODUCTION

New radiometric age determinations on rocks in eastern Massachusetts, Rhode Island, and Connecticut and reconnaissance mapping for a new geologic map of Massachusetts have resulted in a need for reassignment of ages, revision of names, and restatement of geographic distribution for many rock units. Eastern Massachusetts is divided into two structural blocks by the Bloody Bluff and Lake Char fault zones: the narrow Nashoba and Putnam blocks on the west and the extensive Milford-Dedham terrane or "Eastern basement" (Avalonian) on the east (fig. 14).

Brief descriptions for changes in nomenclature are presented for rock units in each of two blocks. Only rock units for which changes in name, age, or geographic distribution are required are discussed. Changes are summarized on table 1.

## NASHOBA AND PUTNAM BLOCKS

The Nashoba block (Skehan, 1969) is bounded by the Bloody Bluff fault zone on the east and the Clinton-Newbury (Essex) fault zone on the west (fig. 14). Within the block, the stratified units are metamorphosed eugeosynclinal rocks of Proterozoic Z or early Paleozoic age, and the intrusive rocks are early to middle Paleozoic in age (table 1). The stratified rocks do not correlate with rock units in either the Merrimack synclinorium to the west or the rocks of the Milford-Dedham terrane to the east. The stratified rocks are the Marlboro Formation, Shawsheen Gneiss, Fish Brook Gneiss, Nashoba Formation, and Tadmuck Brook Schist. Equivalent units in Connecticut are the Tatnic Hill and Quinebaug Formations of the Putnam Group. None of the stratified rocks are fossiliferous. Intrusive rocks in the Nashoba block are the Andover Granite and related rocks and the Sharpners Pond Diorite and related rocks.

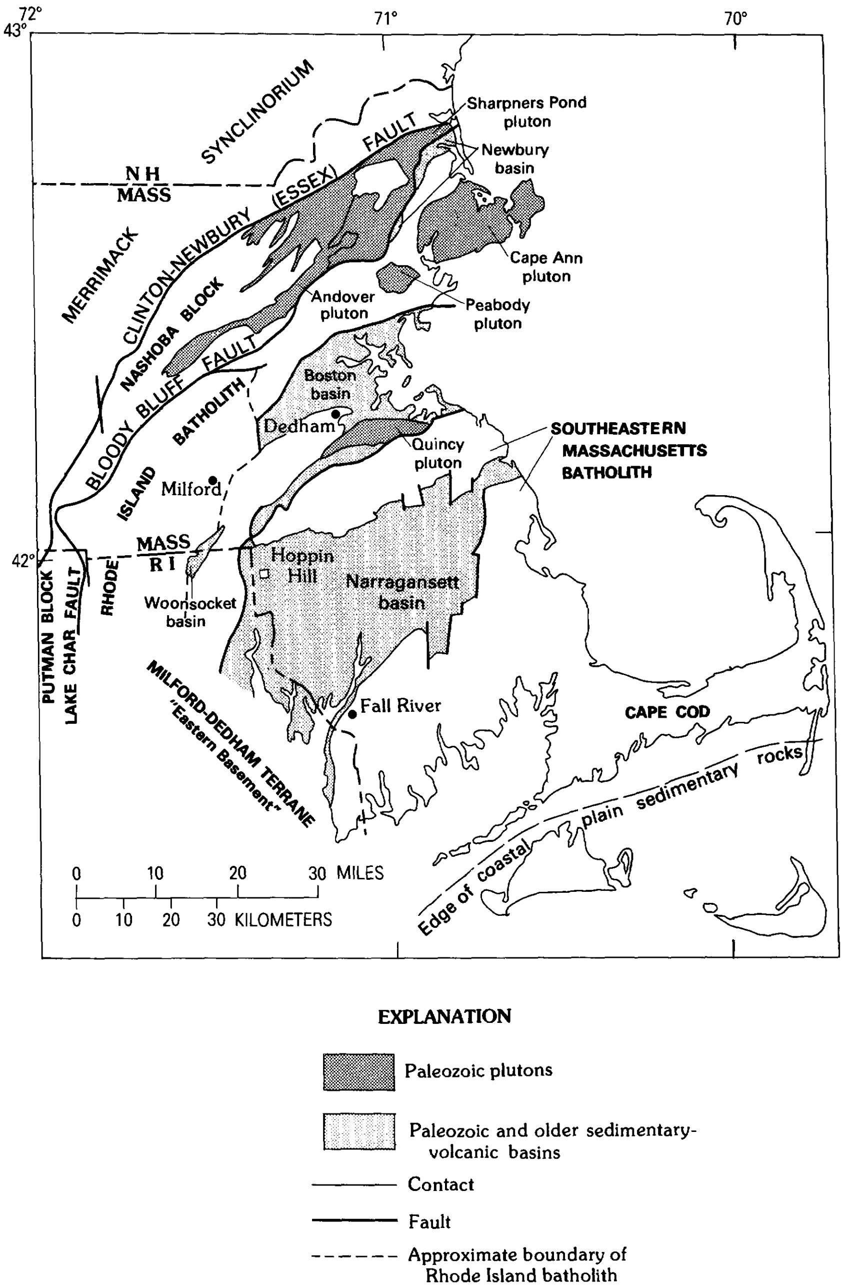


FIGURE 14.—Major geologic features of eastern Massachusetts.

Table 1.--Summary list of present age assignments and present known geographic distribution of selected rock units mapped in the Nashoba block of eastern Massachusetts and the Putnam block of eastern Connecticut

| NAME | AGE | DISTRIBUTION |
| --- | --- | --- |
| | **Stratified rocks** | |
| Fish Brook Gneiss | Ordovician or Proterozoic Z | Massachusetts |
| Marlboro Formation | Ordovician, Cambrian, or Proterozoic Z | Massachusetts |
| Nashoba Formation | Ordovician or Proterozoic Z | Massachusetts |
| Quinebaug Formation | Ordovician, Cambrian, or Proterozoic Z | Connecticut and Massachusetts |
| Shawsheen Gneiss | Ordovician or Proterozoic Z | Massachusetts |
| Tadmuck Brook Schist | Silurian(?), Ordovician, or Proterozoic Z | Massachusetts |
| Tatnic Hill Formation | Ordovician or Proterozoic Z | Connecticut and Massachusetts |
| Yantic Member | Ordovician or Proterozoic Z | Connecticut and Massachusetts |
| Fly Pond Member | Ordovician or Proterozoic Z | Connecticut and Massachusetts |
| | **Intrusive rocks** | |
| Acton Granite | Devonian or Silurian and Ordovician(?) | Massachusetts |
| Andover Granite | Devonian or Silurian and Ordovician(?) | Massachusetts |
| Assabet Quartz Diorite | Silurian | Massachusetts |
| Sharpners Pond Diorite | Silurian | Massachusetts |
| Straw Hollow Diorite | Silurian | Massachusetts |

## Stratified rocks

**Nashoba Formation, Fish Brook Gneiss, and Shawsheen Gneiss.**--The Nashoba Formation (Hansen, 1956), the Fish Brook Gneiss (Castle, 1965b), and the Shawsheen Gneiss (Bell and Alvord, 1976) are intruded by the Silurian Sharpners Pond Diorite and the Devonian or Silurian and Ordovician(?) Andover Granite and generally have been considered to be Ordovician or older. Olszewski (1978) obtained a 742-m.y.±91-m.y. age on a concordia plot of zircons from the Fish Brook and Shawsheen Gneisses, units considered to be volcanic and volcaniclastic by Bell and Alvord. This suggests the possibility of a Proterozoic Z age. The stratified rocks in the Nashoba block, with the exception of the Tadmuck Brook Schist, are assigned an Ordovician or Proterozoic Z age.

**Marlboro Formation.**--The Marlboro Formation (Emerson, 1917) was redefined by Bell and Alvord (1976). The Marlboro is considered here to be Ordovician, Cambrian, or Proterozoic Z in age because it underlies the Nashoba Formation and Fish Brook and Shawsheen Gneisses.

**Tadmuck Brook Schist.**--Bell and Alvord (1976) placed the Tadmuck Brook Schist above the Nashoba Formation. The Tadmuck Brook is truncated at the top by the Clinton-Newbury fault. The Tadmuck Brook was considered to underlie the Silurian rocks of the Merrimack synclinorium in and west of the fault zone. In places in the fault zone, however, the Tadmuck Brook is difficult to distinguish from schist and phyllite of the synclinorium sequence. In addition, Alvord and others (1976, p. 327) suggested that the Tadmuck Brook might lie unconformably on the Nashoba Formation because it truncates units of the Nashoba. The age of the Tadmuck Brook, therefore, is considered to be Silurian(?), Ordovician, or Proterozoic Z.

**Tatnic Hill and Quinebaug Formations.**--The Tatnic Hill Formation, with its members the Yantic and the Fly Pond, and the underlying Quinebaug Formation form the Putnam Group (Dixon, 1964), which lies in a structural block south of the Nashoba block (fig. 14). The units in the Putnam block are considered to be equivalent to those in the Nashoba block (Barosh, 1977; Barosh and others, 1977; Goldsmith, 1980). Their age is also Ordovician or Proterozoic Z, with the Quinebaug also possibly being of Cambrian age (equal to the Marlboro). These formations project into Massachusetts in the Webster area (Barosh, 1977; H. R. Dixon, unpub. data, 1978).

## Intrusive rocks

**Acton Granite.**--The Acton Granite (Hansen, 1956) is considered to be the same general age as the Andover, Devonian or Silurian and Ordovician(?) because its texture, mineralogy, and field relations are similar to those of the Andover (Castle, 1965a, p. 79).

**Andover Granite.**--The Andover Granite (Clapp, 1910) is a mass of syntectonic to late-tectonic granite that consists of several phases that have not been mapped separately. Rb-Sr age determinations by R. E. Zartman and R. S. Naylor (written commun., 1981) and by Handford (1965) indicate that rocks of more than one age may be present. Masses of undeformed pegmatite and aplite give an Early Devonian or Late Silurian age. A gneissic phase gives an Early Silurian or Late Ordovician age. According to Castle (1965a), some of the Andover is gradational with phases of the Sharpners Pond Diorite; if so, a Silurian age is indicated for parts of the Andover. Until the different phases of the Andover are sorted out, a Devonian or Silurian and Ordovician(?) age is indicated for the Andover.

**Sharpners Pond Diorite.**--The Sharpners Pond Tonalite of Castle (1965a) is here renamed the Sharpners Pond Diorite because its compositional range is from quartz diorite to diorite according to the Streckeisen (1973) classification. R. E. Zartman and R. S. Naylor (written commun., 1981) have obtained a concordant U-Th-Pb age of about 430 m.y.$\pm$5 m.y. on zircons from the Sharpners Pond, placing it in the Silurian.

**Straw Hollow Diorite and Assabet Quartz Diorite.**--The Straw Hollow Diorite (Emerson, 1917) and Assabet Quartz Diorite (Hansen, 1956) are considered to be roughly equivalent in age to the Sharpners Pond on the basis of similar compositions and field relations (see Castle, 1965a, p. 79).

## MILFORD-DEDHAM TERRANE

Eastern Massachusetts is part of the "Eastern basement" or Avalonian terrane of southeastern New England (Lilly, 1966; Skehan, 1969; Rodgers, 1970; Naylor, 1975; Osberg, 1978) (see fig. 14). This largely Proterozoic Z terrane (Zartman and Naylor, 1972; Kovach and others, 1977; Smith and Giletti, 1978) extends from near New Haven, Conn., through southeastern Connecticut, western Rhode Island, and into Massachusetts east of the Lake Char and Bloody Bluff faults. In places, basins of Paleozoic and Mesozoic sedimentary and volcanic rocks lie in the Proterozoic Z rocks, and on Cape Cod a cover of Tertiary and Quaternary sediments overlies the basement. The older rocks of the terrane are intruded by Paleozoic granitoid rocks, mostly alkalic, of several ages and by mafic dikes of Mesozoic age.

Granitoid rocks, both gneissic and nongneissic, comprise a large part of the terrane. The widely distributed Proterozoic Z Dedham Granite and the commercial Proterozoic Z Milford Granite give their names to the terrane. Metasedimentary and metavolcanic rocks, including mafic plutonic-volcanic complexes, form marginal belts, enclaves, and roof pendants within the granitoid rocks. An episode of Proterozoic Z volcanism seems to have occurred (Kaye and Zartman, 1980).

Changes in age designations, nomenclature, and geographic distribution are discussed for the units listed in table 2. Where fossils are lacking, radiometric age determinations are assumed to be definitive in determining ages.

### Stratified rocks

**Bellingham Conglomerate.**--The age of the Bellingham Conglomerate is changed from Pennsylvanian to Pennsylvanian, Cambrian, or Proterozoic Z because the formation is not fossiliferous and lithologies in part resemble those in the Boston basin to the north, which, according to Kaye and Zartman (1980), could be as old as Proterozoic Z. The rocks, in part, are similar also to the Lower Pennsylvanian Pondville Conglomerate at the base of the Pennsylvanian sequence in the Northfolk and the Narragansett basins.

Table 2.--Summary list of present age assignments and present known geographic distribution of selected units mapped in the Milford-Dedham terrane of eastern Massachusetts and adjacent States

| NAME | AGE | DISTRIBUTION |
|---|---|---|
| | **Stratified rocks** | |
| Bellingham Conglomerate | Pennsylvanian, Cambrian, or Proterozoic Z | Massachusetts and Rhode Island |
| Blackstone Group (with its Mussey Brook Schist, Quinnville Quartzite, Sneech Pond Schist, and Hunting Hill Greenstone) | Proterozoic Z | Massachusetts and Rhode Island |
| Cambridge Argillite (of Boston Bay Group) | Cambrian(?) and Proterozoic Z | Massachusetts |
| Hoppin Formation | Early and Middle Cambrian | Massachusetts |
| Lynn Volcanic Complex | Early Devonian, Silurian, or Proterozoic Z | Massachusetts |
| Mattapan Volcanic Complex | Proterozoic Z or younger | Massachusetts |
| Plainfield Formation | Proterozoic Z | Connecticut, Rhode Island, and Massachusetts |
| Roxbury Conglomerate (of Boston Bay Group) (with its Brookline, Dorchester, and Squantum Members) | Cambrian(?) and Proterozoic Z | Massachusetts |
| Westboro Formation | Proterozoic Z | Massachusetts |
| | **Intrusive rocks** | |
| Beverly Syenite (of Cape Ann Complex) | Early Silurian or Late Ordovician | Massachusetts |
| Blue Hills Granite Porphyry | Early Silurian or Late Ordovician | Massachusetts |
| Brighton Melaphyre (of Boston Bay Group) | Pennsylvanian, Cambrian, or Proterozoic Z | Massachusetts |

| | | |
|---|---|---|
| Bulgarmarsh Granite | Proterozoic Z | Rhode Island and Massachusetts |
| Cape Ann Complex | Early Silurian or Late Ordovician | Massachusetts |
| Cherry Hill Granite | Devonian | Massachusetts |
| Dedham Granite (includes Barefoot Hills Quartz Monzonite of Lyons, 1969, and Lyons and Wolfe, 1971) | Proterozoic Z | Massachusetts |
| Esmond Granite | Proterozoic Z | Rhode Island and Massachusetts |
| Grant Mills Granodiorite | Proterozoic Z | Rhode Island |
| Hope Valley Alaskite Gneiss (of Sterling Plutonic Group) | Proterozoic Z | Rhode Island, Connecticut, and Massachusetts |
| Milford Granite | Proterozoic Z | Massachusetts |
| Peabody Granite | Middle Devonian | Massachusetts |
| Ponaganset Gneiss (of Sterling Plutonic Group) | Proterozoic Z | Rhode Island, Connecticut, and Massachusetts |
| Scituate Granite Gneiss (of Sterling Plutonic Group) | Proterozoic Z | Rhode Island, Connecticut, and Massachusetts |
| Sharon Syenite | Proterozoic Z | Massachusetts |
| Squam Granite (of Cape Ann Complex) | Early Silurian or Late Ordovician | Massachusetts |
| Topsfield Granodiorite | Proterozoic Z | Massachusetts |
| Wenham Monzonite | Early Devonian | Massachusetts |

**Blackstone Group.**--The Blackstone Series of Rhode Island (Woodworth, in Shaler and others, 1899) was subdivided by Quinn and others (1948, 1949) into four formations: the Hunting Hill Greenstone, the Mussey Brook Schist, the Quinnville Quartzite, and the Sneech Pond Schist and assigned a probable Proterozoic Z age. The Blackstone is now reduced to group status (Goldsmith, 1980), geographically extended into eastern Massachusetts, and assigned a definite Proterozoic Z age because units of the group are intruded by Proterozoic Z plutonic rocks and are more metamorphosed than the nearby Cambrian Hoppin Formation.

**Cambridge Argillite and Roxbury Conglomerate.**--The Cambridge Argillite; the Roxbury Conglomerate with its Brookline, Dorchester, and Squantum Members as defined by Billings (1976); and associated mafic and felsic volcanic rocks, the Brighton Melaphyre, comprise the Boston Bay Group. The Boston Bay Group has been considered to be Pennsylvanian in age on the basis of now discredited plant fossils and by correlation with rocks of the Narragansett basin (see Billings, 1976, and Billings, in Skehan and others, 1979). P. C. Lyons (written commun., 1981) suggests that some or all of the Boston Bay Group could be Devonian or Mississippian because the Cambridge Argillite lacks the plant fossils so abundant in the Rhode Island Formation of the Narragansett basin. Cameron and Jeanne (1976) suggest an Ordovician age for the Boston Bay Group. Recent isotopic dating of the Mattapan Volcanic Complex and field observations by Kaye (Kaye and Zartman, 1980) indicate that the Boston Bay Group underlies fossiliferous Cambrian strata and, thus, the Boston Bay Group may be as old as Proterozoic Z. In support of Kaye and Zartman's thesis, acritarchs of Proterozoic Z to Early Cambrian age have been identified recently in the Cambridge Argillite (Lenk and others, 1982). Accordingly, the age of the Boston Bay Group could be limited to, and is most likely, Proterozoic Z and possibly Early Cambrian. The Boston Bay Group is shown as Paleozoic or Proterozoic Z on a new bedrock geologic map of Massachusetts (Zen and others, 1981) that was compiled before the fossil discovery.

**Hoppin Formation.**--The name Hoppin Slate of Emerson (1917; see Foerste, in Shaler and others, 1899) is changed herein to Hoppin Formation because the rock is not strictly a slate and contains limestone, shaley limestone, sandstone, and conglomeratic quartzite.

**Lynn Volcanic Complex.**--The Lynn Volcanic Complex (LaForge, 1932) has been correlated with the Upper Silurian and Lower Devonian(?) Newbury Volcanic Complex (Shride, 1976; Billings, 1976) and with the Proterozoic Z or younger Mattapan Volcanic Complex (see discussions by Shride, 1976, p. 173-174, and Billings, in Skehan and others, 1979, p. A17-A18). The Lynn is intruded by alkalic granite of the Quincy type (LaForge, 1932, p. 33), but, according to R. L. Zartman (oral commun., 1980), two groups of alkalic rocks exist in the area, one of Ordovician age (Quincy type) and one of Devonian age (Peabody type). We do not know which of the two intrude the Lynn at LaForge's locality. The Lynn could range in age from Devonian to Proterozoic Z and, therefore, is assigned an Early Devonian, Silurian, or Proterozoic Z age.

**Mattapan Volcanic Complex.**--The Mattapan Volcanic Complex (LaForge, 1932) unconformably underlies the Roxbury Conglomerate of the Boston Bay Group, according to Billings (1976). Correlation with volcanic rocks of Pennsylvanian age adjacent to the Norfolk basin and in the Narragansett basin would permit a Pennsylvanian age. Correlation with the adjacent Lynn Volcanic Complex, on the other hand, would permit an age as young as Early Devonian. Field relations indicate the Mattapan to be younger than the Dedham Granite. U-Th-Pb radiometric ages on zircons from the Mattapan by Zartman (Kaye and Zartman, 1980) give a 602-m.y.±3-m.y. age for the Mattapan. If the zircon age is accepted, the Mattapan would be Proterozoic Z or younger (earliest Cambrian?). The Mattapan, therefore, is assigned a Proterozoic Z or younger age.

**Westboro and Plainfield Formations.**--The Plainfield Formation of Connecticut and southern Rhode Island and the Westboro Formation of Massachusetts are designated Proterozoic Z because they are intruded by Proterozoic Z granitoid rocks of the Rhode Island batholith (Zartman and Naylor, 1972, and written commun., 1978; Nelson, 1975; Goldsmith, 1980). The Plainfield and the Westboro are similar in composition and occupy a similar structural-stratigraphic position. They are correlated with part or all of the Blackstone Group. The Plainfield extends into Massachusetts in and along the western flank of the Rhode Island batholith.

**Intrusive rocks**

**Barefoot Hills Quartz Monzonite.**--The Barefoot Hills Quartz Monzonite of Lyons (1969) and Lyons and Wolfe (1971) is considered to be coeval with (Lyons, 1969), or a phase of (Wones, 1978), the Dedham Granite and, therefore, is included in the Dedham and assigned a Proterozoic Z age.

**Beverly Syenite.**--The Beverly Syenite is a facies of the Cape Ann Complex according to Dennen (1976). Its age designation, thus, is changed from late(?) Paleozoic to: Early Silurian or Late Ordovician (Zartman and Marvin, 1971).

**Blue Hills Granite Porphyry.**--The name Blue Hills Granite Porphyry of Chute (1966, 1969) is retained although the term Blue Hills Porphyry has been used by Naylor and Sayer (1976). Naylor and Sayer believe the Blue Hills to be the same age as the Quincy Granite, which Zartman and Marvin (1971) and Zartman (1977) have dated with the Cape Ann Complex as Early Silurian or Late Ordovician.

**Bulgarmarsh Granite.**--The Bulgarmarsh Granite (Pollock, 1964) of southeastern Rhode Island and adjacent Massachusetts south of Fall River has been dated as early Paleozoic (Galloway, 1970), but it cannot be separated readily in the field from granite north of Fall River from which R. L. Zartman (written commun., 1979) has obtained a Proterozoic Z age similar to others in the southeastern Massachusetts batholith (see Kovach and others, 1977). The Bulgarmarsh, thus, is assigned a Proterozoic Z age.

**Cape Ann Complex.**--The Cape Ann Granite is here renamed the Cape Ann Complex for the alkalic plutonic rocks (facies of Dennen, 1976) comprising the Cape Ann pluton (fig. 14). These are an unnamed alkalic granite and quartz syenite unit (Cape Ann Granite of Warren and McKinstry, 1924, and of Toulmin, 1964), the Beverly Syenite, and the Squam Granite. Zartman and Marvin (1971) derived a radiometric age of 450 m.y.$\pm$25 m.y. from a concordia plot of U-Pb analyses of zircons from granite of the Cape Ann Complex and consider the Cape Ann to be Late Ordovician in age. A possibility exists, however, that because of the analytical uncertainty all or part of the complex might be Early Silurian (table 2); the Cape Ann, therefore, is considered to be Early Silurian or Late Ordovician in age.

**Cherry Hill Granite.**--The Cherry Hill Granite, according to Toulmin (1964), who named it, intrudes the Wenham Monzonite and is, thus, at least slightly younger than the Wenham. The latter is considered with the Peabody Granite to be part of the suite of alkalic plutons of eastern Massachusetts having Devonian radiometric ages (R. E. Zartman, written commun., 1981; Lyons and Kreuger, 1976, p. 94).

**Dedham Granite.**--The name Dedham Granodiorite (Emerson, 1917) was changed to Dedham Granite (Wones, 1978) because most of the rock mapped as Dedham is granite according to the Streckeisen (1973) classification.

**Esmond Granite.**--The Esmond Granite (Quinn and others, 1948) is Proterozoic Z in age (Hermes and others, 1981). It has similar spatial and textural similarities to the Proterozoic Z granites of southeastern Massachusetts and northern Rhode Island. Its age, thus, is changed from Late Ordovician or older to Proterozoic Z. The Esmond extends from Rhode Island into Massachusetts in the Blackstone area.

**Grant Mills Granodiorite.**--The Grant Mills Granodiorite (Warren and Powers, 1914) is gradational into the Esmond according to Quinn (1971, p. 29) and here is assigned a Proterozoic Z age (Hermes and others, 1981). The Grant Mills does not project into Massachusetts, although it is shown in Rhode Island on the bedrock geologic map of Massachusetts (Zen and others, 1981).

**Hope Valley Alaskite Gneiss.**--The Hope Valley Alaskite Gneiss (Moore, 1958) is continuous from Rhode Island and Connecticut into Massachusetts in the Rhode Island batholith (anticlinorium of Rodgers, 1970). Recent age determinations (Day and others, 1980; Hermes and others, 1981) indicate a Proterozoic Z rather than Mississippian(?) or older age for this unit.

**Milford Granite.**--U-Th-Pb radiometric age determinations on zircons from the Milford Granite (Emerson and Perry, 1907) by R. E. Zartman and R. S. Naylor (written commun., 1978) indicate a Proterozoic Z age. Its age designation of Proterozoic Z to early Paleozoic, thus, is changed to: Proterozoic Z. Although Emerson and Perry originally extended the Milford Granite into Rhode Island, Quinn (1971) does not recognize the Milford there. Granite that Emerson mapped as Milford in Rhode Island is mapped now mostly as Esmond Granite (see Quinn, 1971, p. 29). The Milford Granite, thus, is restricted geographically to Massachusetts.

**Northbridge Granite Gneiss.**--Rocks formerly mapped as Northbridge Granite Gneiss (Emerson, 1917) have been found to be part of either the

Scituate Granite Gneiss or the Ponaganset Gneiss (H. R. Dixon, unpub. data, 1978), and the name Northbridge herewith is abandoned.

**Peabody Granite.**--Zartman and Marvin (1971) considered the Peabody Granite (Clapp, 1910; Toulmin, 1964) to be the same age as the Quincy and former Cape Ann Granites on the basis of radiometric age determinations by several methods. However, subsequent work by R. E. Zartman (oral commun., 1979 and cited in Lyons and Kreuger, 1976, p. 94) has indicated that the Peabody is approximately 370 m.y. old. The Peabody, therefore, is assigned a Middle Devonian age.

**Ponaganset Gneiss.**--The Ponaganset Gneiss (Quinn, 1967) is part of the granite gneiss terrane of the Rhode Island batholith that extends from Rhode Island and Connecticut into Massachusetts. Its age is considered to be Proterozoic Z, the same as the Hope Valley Alaskite Gneiss and Scituate Granite Gneiss, because of its similar gneissic character and intimate field relations with those rocks.

**Scituate Granite Gneiss.**--The Scituate Granite Gneiss (Quinn, 1951) is Proterozoic Z in age on the basis of radiometric age determinations mentioned by Day and others (1980) and determinations by R. E. Zartman and R. S. Naylor (written commun., 1978). Hermes and others (1981), however, have shown that rock mapped as Scituate in central Rhode Island is Devonian. The Scituate, like the Hope Valley Alaskite Gneiss and the Ponagansett Gneiss, extends into Massachusetts in the Rhode Island batholith.

**Sharon Syenite.**--The Sharon Syenite (Emerson, 1917) appears to have been produced by interaction between Dedham Granite and older gabbro and diorite. Its age, thus, is changed from Devonian or older to: Proterozoic Z.

**Squam Granite.**--The Squam Granite (Clapp, 1910, 1921) is a facies of the Cape Ann Complex (Dennen, 1976) and, thus, is Early Silurian or Late Ordovician in age (Zartman and Marvin, 1971).

**Topsfield Granodiorite.**--The age designation for the Topsfield Granodiorite (Toulmin, 1964) is changed from middle Paleozoic to Proterozoic Z on the basis of a K-Ar age of 640 m.y. on hornblende obtained from a dioritic phase of the Topsfield by Zartman and Naylor (1972; see also Shride, 1976, p. 151).

**Wenham Monzonite.**--The Wenham Monzonite (Toulmin, 1964) is considered to be part of the suite of alkalic plutonic rocks of eastern Massachusetts having Devonian radiometric ages (Zartman, 1977; Lyons and Kreuger, 1976, p. 94). R. E. Zartman (written commun., 1981) obtained a Pb-U age of 395 m.y.$\pm$20 m.y. on zircon from the Wenham Monzonite, and it is thus assigned an Early Devonian age.

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# MAUZY FORMATION, A NEW STRATIGRAPHIC UNIT OF PERMIAN AGE IN WESTERN KENTUCKY

By Thomas M. Kehn, John G. Beard[1], and
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## ABSTRACT

The Mauzy Formation herein is named and defined as a rock-stratigraphic unit in the western Kentucky coal field. Paleontological studies of the uppermost strata of the Sturgis Formation indicate that they are, at least in part, of Early Permian age, and they are reassigned to the Mauzy Formation. These strata are preserved in a fault block in the Sturgis, Union County, Ky., area. Lower Permian strata are unknown elsewhere in the Eastern Interior (Illinois basin) coal field area but probably covered part or all of this area prior to being eroded. The type section of the Mauzy Formation consists of interbedded shale, siltstone, and limestone, with minor amounts of sandstone and coal. It has a known thickness of about 390 ft but may be as much as 1,300 ft thick, as suggested by projected structural data. The Mauzy Formation conformably overlies the Sturgis Formation; the contact between the formations is at the base of a predominantly limestone sequence, in which Permian-age fusulinids have been identified, and above a sequence of shale, siltstone, sandstone, and coal that contains spores and pollen of Pennsylvanian age.

## INTRODUCTION

Detailed geologic investigations supported by core drilling in the Bordley 7 1/2-min quadrangle (Kehn, 1975a) and the Sturgis 7 1/2-min quadrangle (Kehn, 1975b) have revealed a small down-faulted structural block of the Rough Creek fault system that contains a stratigraphic section more than 3,600 ft thick that was assigned to the Pennsylvanian System by Kehn (1973). Glenn (1912a,b; 1922) and Lee (1916) reported the Pennsylvanian section to be much thinner (fig. 15) and, because of the paucity of stratigraphic and structural control, did not recognize these younger strata in their reports of this area. Dunbar and Henbest (1942) summarized information on fusulinid-bearing horizons in the Illinois portion of the Eastern Interior coal field area and reported that fusulinids from the uppermost marine limestone beds were of Late Pennsylvanian age. They also stated that equivalent strata might be present in western Kentucky.

A core drilling program in the western Kentucky coal field conducted by the Kentucky Geological Survey was concluded recently. The core hole Gil-30 was taken from the fault block in which the thickest Pennsylvanian section had been identified. Samples of limestone from this core were obtained for paleontological studies, and this report is based in part on the result of that investigation and in part on a revision of structural control from data obtained in the core drilling program and reinterpretation of the previous mapping.

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The rocks formerly assigned to the uppermost part of the Sturgis Formation herein are reassigned to the Mauzy Formation, a new stratigraphic unit that is all, or in part, of Early Permian age. A summary of the nomenclature is given in figure 15.

## PREVIOUS NOMENCLATURE

### Sturgis Formation

Recently, the nomenclature of the uppermost part of the rock-stratigraphic section of Pennsylvanian age in western Kentucky was modified and renamed the Sturgis Formation (Kehn, 1973), (see fig. 15). Later mapping, in the Bordley and Sturgis quadrangles (Kehn, 1975a,b) (see fig. 16), revealed about 470 ft of younger strata not previously described. These rocks, although containing a much greater percentage of limestone and calcareous siltstone than previously described, were included as part of the Sturgis Formation. The base of the limestone and calcareous siltstone sequence is proposed as the base of the Mauzy Formation and the top of the Sturgis Formation, which, as redefined, is about 2,039 ft thick.

The Sulphur Springs coal bed, about 200 ft below the top of the Sturgis Formation, is equivalent to or slightly younger than the Pittsburgh coal bed of the Appalachian area (R. A. Peppers, Illinois State Geological Survey, written commun., 1978). It is the youngest stratum of Late Pennsylvanian age for which paleontological data are available in the western Kentucky coal field.

## NEW STRATIGRAPHIC UNIT

### Mauzy Formation (here named)

The area underlain by the Mauzy Formation largely is concealed by loess and alluvium. Hence, surface criteria for differentiating the Mauzy Formation from the Sturgis Formation for stratigraphic and paleontological purposes could not be found. The core from core hole Gil-30 (fig. 17) confirmed the existence of the thick limestone and calcareous siltstone beds previously reported by Kehn (1975a, b) and also provided material for paleontological studies of the contained fusulinids. Therefore, the composite type section for the formation is designated as the 340 ft of measured section from core hole Gil-30 drilled by the Kentucky Geological Survey and at least 50 ft of younger strata inferred to exist in a nearby ridge about 1 mi west of Cap Mauzy Lake (fig. 17), from which the name Mauzy is taken. The core hole was about 7 1/2 mi northeast of Sturgis, Ky., in the northwestern part of the Bordley quadrangle (see figs. 16 and 17 and p. H79 for additional location data). The core is on file at the core library of the Kentucky Geological Survey, Lexington, Ky.

A generalized description of the composite type section of the Mauzy is included in this report. A thin bed of limestone at a depth of 193.9-195.8 ft (243.9-245.8 ft, p. H81) contains a few large Triticites sp. of Early Permian age (Douglass, 1979). About 145 ft of strata below the fusulinid-bearing limestone is assigned also to the Mauzy Formation because the strata are lithologically similar to rocks above the limestone.

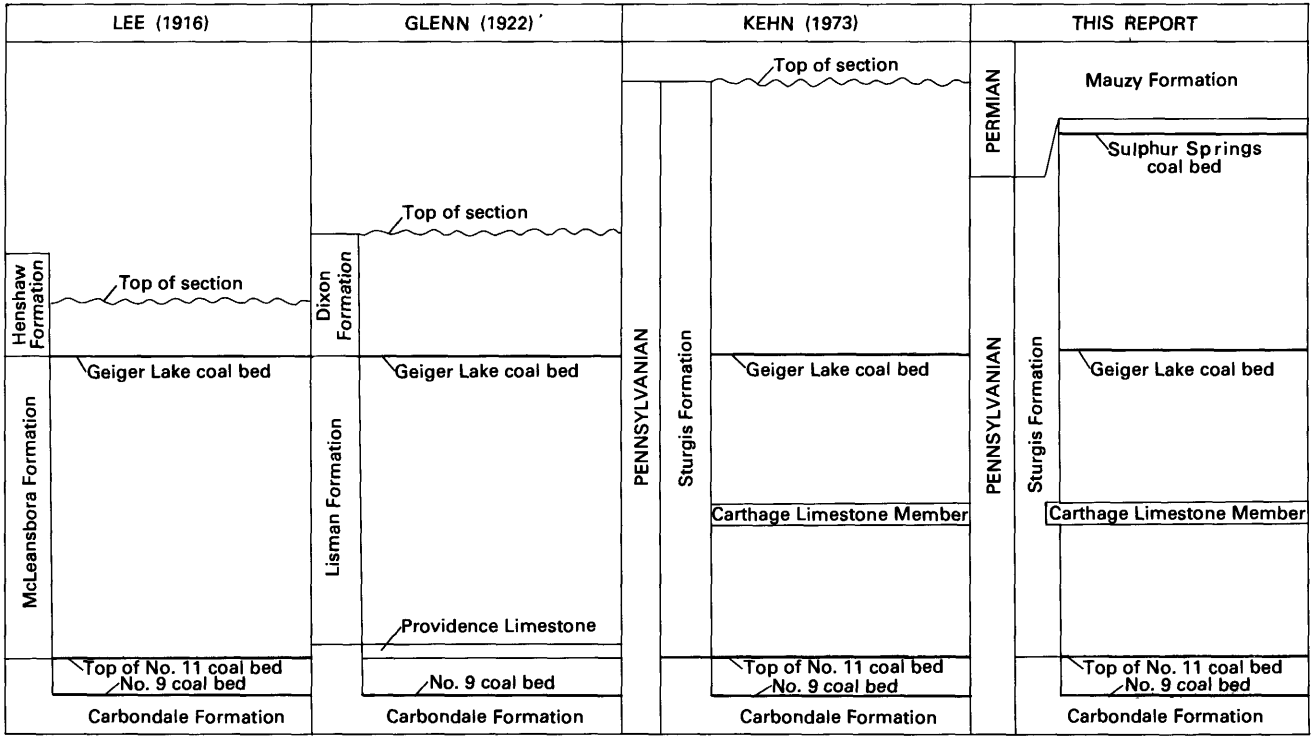


FIGURE 15.—Nomenclature of Kehn and others (this report) and of previous reports. Lee, Glenn, and Kehn nomenclature modified from Kehn (1973).

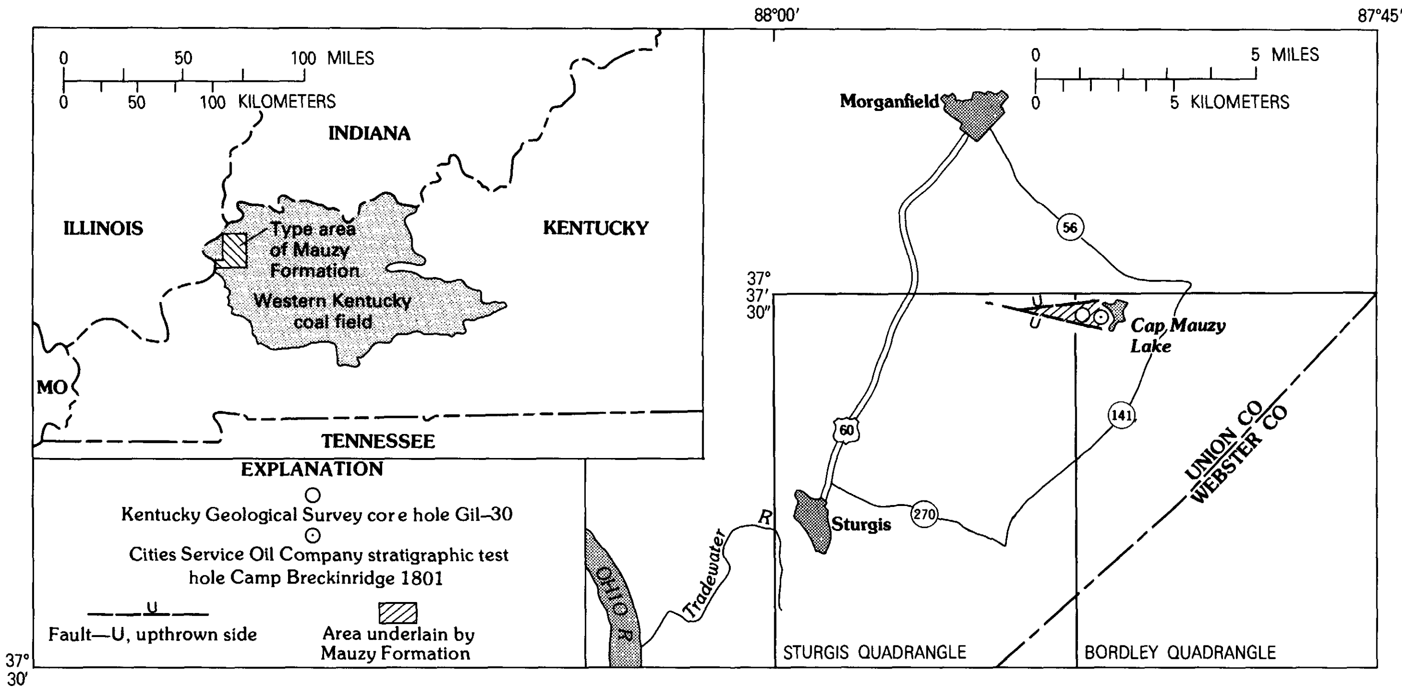


FIGURE 16.—Western Kentucky coal field, type area of the Mauzy Formation, and location of the stratigraphic test holes.

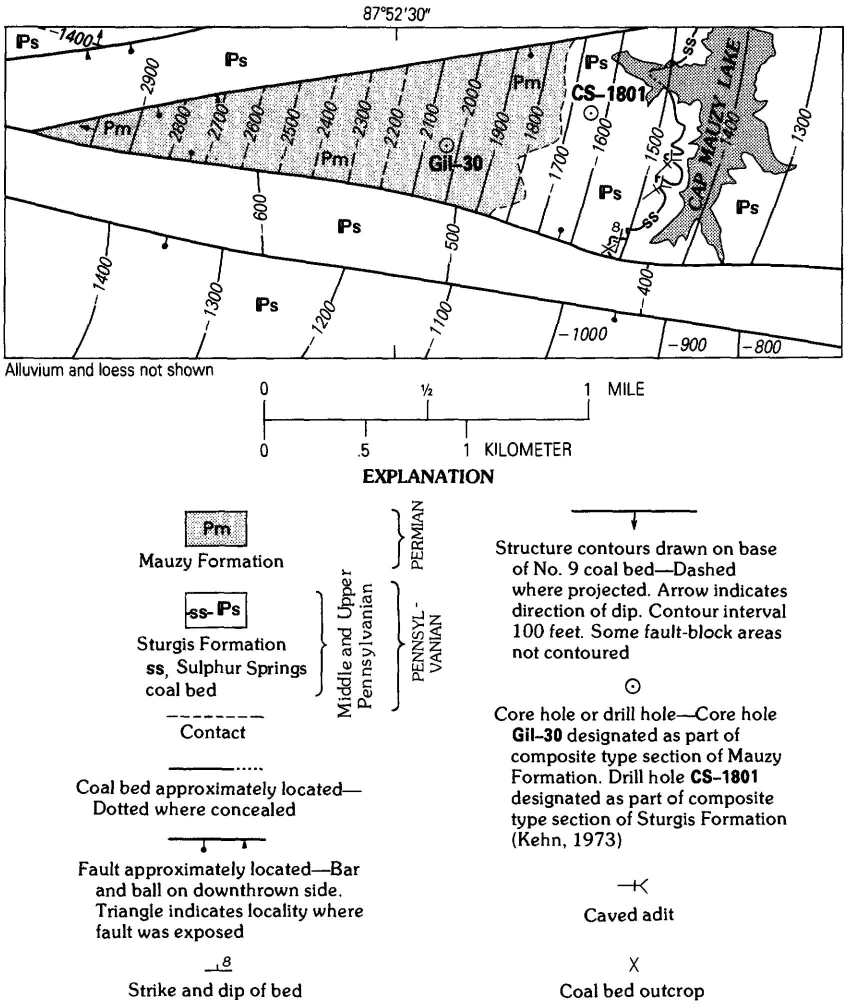


FIGURE 17.—Area underlain by the Mauzy Formation, western Kentucky. Modified from Kehn (1975a, b).

The Mauzy Formation is composed of interbedded shale, siltstone, limestone, and sandstone. Shale and siltstone, the dominant rock types in the cored section, make up about 70 percent of the formation. The shale and siltstone, generally interlaminated, are commonly light gray to black or green to greenish gray. It is poorly to well laminated in even to wavy beds, except for steep crossbeds at a depth of 69.0-80.7 ft in the cored section (see p. H80). It also contains a few thin limestone and sandstone beds. The shale and siltstone generally are calcareous and micaceous along bedding surfaces. A coal bed and two thin carbonaceous shale beds are reported in the lower part of the cored section.

Limestone makes up about 25 percent of the Mauzy Formation. In the Sturgis Formation, limestone makes up less than 5 percent of the formation (Kehn, 1973, p. B8). The limestone of the Mauzy Formation is generally light gray to tan to buff. It is mostly very fine to fine grained and clayey to silty where it grades into or is interbedded with shale or siltstone. These calcareous sequences have been described as a pelmicrite by Dever and Macucuown (personal commun., 1978), a lithology not common to the western Kentucky coal field. The limestone is mostly even bedded and is as much as 10 ft thick. Marine fossils have not been recognized except for the fusilinids in the thin limestone bed near the middle of the cored section.

Sandstone makes up less than 5 percent of the formation. It is medium gray and fine grained. The sandstones are generally thin bedded; crossbedding, cut-and-fill structures, and interlaminations of shale and siltstone are common.

Coal is a minor constituent of the Mauzy Formation, as only three very thin coaly or carbonaceous shale beds were found in the lower part of the cored section. Additional coal beds could be in the loess-covered 50 ft of the type section lying above core hole Gil-30.

The total thickness of the formation is not known because of structural uncertainties and probable erosion of the top part of the section within the fault block. The formation may be as much as 1,300 ft thick as indicated by projected structure contours drawn on the No. 9 coal bed (fig. 17), which is widely used in western Kentucky as a structural datum. A total thickness of 1,300 ft or more for the Permian in western Kentucky would not be improbable as shown by McKee and others (1967, pl. 7). However, the structure is probably more complex than that shown (fig. 17). The strata may have dips more or less than those projected, or there could be a reversal of dip direction, and the 390 ft described here might be the maximum thickness preserved in the fault block and be essentially the maximum thickness deposited in the Eastern Interior region.

## CONTACTS

### Lower boundary of Mauzy Formation

The Mauzy Formation appears to overlie the Sturgis Formation conformably and may intergrade with it. Deposition is assumed to have been continuous from one formation to the other as no evidence of a disconformity is seen in the core. This contact is here arbitrarily placed at the base of the limestone sequence in the uppermost 340 ft (390.0 ft of the described section, p. H80-H82) of strata cored at the core hole Gil-30 locality. Such placement of the contact was made because the core contains a much greater percentage of limestone and calcareous shale above and a much greater percentage of sandstone and coal below.

The upper limit of the Mauzy Formation is not defined because it has been removed by erosion or is present only in the western portion of the fault block (fig. 17) where stratigraphic and structural control is not available to determine its placement. Younger rocks, if present, have been removed by erosion or are covered by Quaternary alluvium or Pleistocene loess in the western portion of the fault block.

### Lower boundary of the Permian System

The boundary between the Pennsylvanian and Permian Systems cannot be defined precisely because sufficient paleobotanical and paleontological data are lacking. For convenience, the systemic boundary is tentatively placed at the proposed contact between the Mauzy and Sturgis Formations in core hole Gil-30. This differentation is based in part on gross lithologic character and on age differences of beds above and below the boundary.

The Sulphur Springs coal bed, about 200 ft below the boundary, is reported (R. A. Peppers, written commun., 1978) to be equivalent to or younger than the Pittsburgh coal bed and is presently the youngest coal bed of Late Pennsylvanian age identified in the western Kentucky section. The unnamed limestone, about 140 ft above the boundary, contains a few fusulinids of Early Permian age. Because deposition appears to have been continuous, all or part of the 340 ft of stratigraphic section between the Sulphur Springs coal bed and the fusulinid-bearing limestone bed mentioned earlier could be assigned to a transition zone or the boundary placed elsewhere within this interval of the section. Similarly, rocks at the Pennsylvanian-Permian boundary in the Appalachian area are gradational, and the contact is arbitrarily chosen (McKee and others, 1967, p. 36), or it is not precisely defined (Henry and others, 1979, p. 85).

## LOCATION OF DRILL HOLES

(fig. 17)

[Core hole Gil-30; 300 ft from south line and 1,650 ft from east line of Carter coordinate section 13, N-20, or 2,800 ft from north boundary and 700 ft from west boundary of Bordley 7 1/2-min quadrangle, Union and Webster Counties, Ky. Surface elevation 455 ft above mean sea level. Core log datum 455 ft above mean sea level. Drilled 1976]

[Cities Service Oil Company, stratigraphic test hole, drill hole CS-1801, 700 ft from south line and 600 ft from west line of Carter coordinate section 12, N-20, or 2,150 ft from north boundary and 3,150 ft from west boundary of Bordley 7 1/2-min quadrangle, Union and Webster Counties, Ky. Surface elevation 520 ft above mean sea level. Driller's log and geophysical electric log datum 526 ft above mean sea level. Drilled 1965]

## COMPOSITE TYPE SECTION OF THE MAUZY FORMATION

[This composite type section includes 50 ft of strata that is inferred above the top of the cored section at the drill hole locality. Approximately 1,000 ft of additional strata may be present as shown by the structure (fig. 17). However, no lithologic data are available, and structural control is uncertain; therefore, no description can be given for this part of the section. The following descriptive log is generalized from the core description by personnel of the Kentucky Geological Survey. The detailed descriptive log and core can be studied at that Survey, in Lexington, Ky.]

| Unit | Depth (feet) |
|---|---|
| **Permian System** | |
| **Mauzy Formation:** | |
| Surface, includes loess, soil and bedrock probably similar to rock types described below | 0-69.0 |
| Shale, medium-gray to greenish-gray, slightly silty with siltstone laminations dipping 10-20° | 69.0-80.7 |
| Sandstone, medium-gray, fine-grained; thin dark shale laminations at top, thicker toward base; base sharp | 80.7-106.1 |
| Shale, gray to dark-gray, calcareous | 106.1-107.9 |
| Limestone, gray to buff, finely crystalline to dense, slightly clayey, calcite-filled fractures | 107.9-113.3 |
| Shale, medium-gray, calcareous | 113.3-113.8 |
| Limestone, gray to buff | 113.8-114.8 |
| Shale, medium-gray to greenish-gray, clayey at top, thin beds of limestone at base | 114.8-120.6 |
| Shale, medium-gray, clayey, calcareous toward base | 120.6-124.1 |
| Limestone, buff to gray, finely crystalline to dense; clayey toward base | 124.1-126.7 |
| Shale, medium-gray, clayey | 126.7-128.9 |
| Limestone, light-gray to tan, dense; calcite-filled fractures near top; green to gray shale parting 0.25 ft thick in middle of unit | 128.9-132.7 |
| Shale, light-gray to greenish-gray | 132.7-133.5 |
| Limestone, buff to gray, clayey; becoming more shaley toward base | 133.5-139.0 |
| Shale, dark-gray, calcareous | 139.0-140.0 |

| | |
|---|---|
| Limestone, light-gray to tan to buff, dense; scattered fractures, argillaceous in part | 140.0-147.9 |
| Shale, dark-gray, calcareous; few thin lenses of medium-gray limestone in basal part | 147.9-150.0 |
| Limestone, medium-gray; argillaceous, dense | 150.0-159.9 |
| Shale, gray to greenish-gray, calcareous; scattered limestone nodules | 159.9-170.8 |
| Limestone, medium-gray; argillaceous, dense | 170.8-172.2 |
| Shale, dark-gray, calcareous | 172.2-176.1 |
| Limestone, light-gray to tan to buff; argillaceous, dense; dark-gray to greenish-gray shale partings | 176.1-202.8 |
| Shale, greenish-gray to dark-gray, thinly laminated in part, silty; few thin laminations of very fine grained sandstone in middle of unit | 202.8-243.9 |
| Limestone, light- to medium-gray; very argillaceous; fossiliferous with a few fusulinids of a large Triticites sp. of Early Permian age | 243.9-245.8 |
| Shale, dark-gray, thinly laminated; few siltstone laminations near base | 245.8-252.8 |
| Sandstone, light-gray, very fine grained, calcareous; scattered clay and shale nodules and laminations | 252.8-253.5 |
| Shale, very dark gray, clayey, carbonaceous | 253.5-266.5 |
| Limestone, tan to buff, finely crystalline to dense; fractured, nodular appearance, interbedded with greenish-gray to gray shale | 266.5-290.9 |
| Shale, gray to greenish-gray, poorly bedded | 290.9-295.0 |
| Shale, black, carbonaceous | 295.0-295.1 |
| Shale, gray to greenish-gray; occasional silt laminations, limestone nodules at top and base | 295.1-325.0 |
| Shale, dark-gray; occasional silt laminations | 325.0-373.1 |
| Coal, bony shaly near top | 373.1-374.1 |
| Shale, dark-gray to green, calcareous | 374.1-375.0 |
| Limestone, brown to tan, finely crystalline to dense | 375.0-376.8 |
| Shale, dark-gray | 376.8-377.2 |

Limestone, brown to tan; argillaceous toward base ........ 377.2-379.1

Shale, dark-gray .................................. 379.1-380.0

Limestone, light-gray to brown; dense, argillaceous ....... 380.0-388.2

Shale, dark-gray, calcareous ........................ 388.2-389.5

Limestone, light-gray to brown; dense, argillaceous. The contact between the Mauzy and the Sturgis Formations is arbitrarily placed at base of this unit. The boundary between the Permian and the Pennsylvanian Systems also is placed tentatively at this contact ........................ 389.5-390.0

## SECTIONS OF THE STURGIS FORMATION

### Upper part of the section

[This part of the composite section includes 200 ft of strata described from the core taken at the core hole Gil-30 locality and at the stratigraphic test hole, drill hole CS-1801, locality. It corresponds to the 70- to 270-ft portion of the upper part of the section of the Sturgis Formation (Kehn, 1973, p. B11-B12). The systemic boundary between the Permian and the Pennsylvanian Systems may be within this unit]

| Unit | Depth (feet) |
|---|---|
| **Pennsylvanian System** | |
| **Upper Pennsylvanian Series** | |
| **Sturgis Formation (in part):** | |
| Shale, green to gray, calcareous; clayey and brown to reddish near middle and silty at base | 390.0-395.9 |
| Sandstone, medium-gray, very fine grained; argillaceous | 395.9-402.4 |
| Shale, dark-gray | 402.4-410.0 |
| Shale, very dark gray to black; argillaceous; scattered limestone bands and nodules; carbonaceous in upper part | 410.0-440.5 |
| Coal, bright- and dull- banded; fine pyrite and calcite-filled veins at top | 440.5-442.3 |
| Shale, medium-gray, clayey, non bedded | 442.3-443.3 |
| Coal, dull to bright, fusain partings; bony with carbonaceous shale partings at top and base; calcite laminations in bony coal at base | 443.3-454.3 |
| Shale, medium-gray, clayey, non bedded; plant impressions at top; limestone nodules in lower part | 454.3-468.8 |

Shale, medium-gray, and limestone, tan, dense; silty toward base . . . . . . . . . . 468.8-480.1

Shale, dark-gray; with laminations of light-gray siltstone and light-gray, fine-grained sandstone at base . . . . . . . . . . 480.1-490.0

Sandstone, light- to medium-gray, fine-grained . . . . . . . . . . 490.0-493.3

Shale, greenish-gray to black; carbonaceous at base; pyrite nodules and laminations . . . . . . . . . . 493.3-500.8

Coal, dull to bright, much fusain; pyrite bands on pyrite or on cleat; much bony coal and carbonaceous shale bands . . . . . . . . . . 500.8-505.8

Shale, medium-gray, non bedded; calcareous, with white limestone nodules . . . . . . . . . . 505.8-520.4

Limestone, light-gray to gray, finely crystalline to dense . . . . . . . . . . 520.4-526.0

Shale, medium-gray; scattered silty and calcareous laminations and nodules . . . . . . . . . . 526.0-538.0

Shale, black, carbonaceous; few pyrite nodules . . . . . . . . . . 538.0-541.1

Sandstone, medium-gray, fine- to medium-grained; scattered thin, carbonaceous shale laminations, argillaceous; sandy at top . . . . . . . . . . 541.1-556.5

Shale, dark-gray; scattered silty laminations . . . . . . . . . . 556.5-565.1

Shale, black, carbonaceous, fissile . . . . . . . . . . 565.1-567.1

Limestone, brown to gray, finely crystalline to dense; in beds 1-2 ft thick separated by dark-gray to black carbonaceous shale beds 0.2-0.9 ft thick . . . . . . . . . . 567.1-572.8

Shale, dark-gray to black; calcareous with limestone laminations at top, carbonaceous and bony coal at base . . . . . . . . . . 572.8-589.0

**Lower part of the section**

[This part of the composite section includes a modified log description of about 1,839 ft of the Sturgis Formation (Kehn, 1973, p. B12-B23). Depth to described units modified to provide continuity]

| Unit | Depth (feet) |
|---|---|
| **Pennsylvanian System** | |
| **Middle Pennsylvanian Series** | |
| **Sturgis Formation (in part):** | |
| Coal, dull to brightly banded. The Sulphur Springs coal bed (Kehn, 1973, p. B12, 268-270 ft) contains a spore assemblage that is equivalent to or younger than the Pittsburgh coal bed (R. A. Peppers, written commun., 1978) | 589-591 |
| Siltstone, sandstone, shale, limestone, coal, and underclay: siltstone, light- to dark-gray; interbedded with sandstone and shale; sandstone, light-gray, very fine to fine-grained; shale, light-gray to black; carbonaceous; limestone, brown and gray, dense; silty and shaley; coal, impure, generally in thin beds; underclay, light- to dark-gray | 591-1,416 |
| Coal (Geiger Lake coal bed), bright | 1,416-1,417 |
| Siltstone, shale, sandstone, limestone, and coal: siltstone, light- to dark-gray; shale, light-gray to black, clayey to sandy; sandstone, light-gray, interbedded with shale and siltstone; limestone, medium-gray; coal, thin, generally impure. Lisman coal bed, 6 in thick at base | 1,417-1,776 |
| Siltstone, shale, sandstone, limestone, and coal: rock types similar to those described above. No. 18 coal bed, thin, in middle of unit | 1,776-1,957 |
| Limestone (Carthage Limestone Member), light-olive-gray, dense, fossiliferous | 1,957-1,964 |
| Shale, siltstone, sandstone, limestone, coal, and underclay: rock types similar to those described above. Unit includes three named coal beds and the Madisonville Limestone Member, a single bed 5 ft thick at base | 1,964-2,248 |
| Siltstone, sandstone, shale, limestone, coal, and underclay: rock types similar to those described above. Unit includes four named coal beds and the Providence Limestone Member at base. Basal part of claystone of the Providence Limestone Member in contact with top of the No. 11 coal bed of the Carbondale Formation | 2,248-2,429 |

## SECTION OF THE CARBONDALE FORMATION

[This section includes a modified log description of about 124 ft of the Carbondale Formation penetrated by stratigraphic test hole, drill hole CS-1801, Cities Service Oil Company. Depth to described units modified to provide continuity]

| Unit | Depth (feet) |
|---|---|
| **Pennsylvanian System** | |
| **Middle Pennsylvanian Series** | |
| **Carbondale Formation (in part):** | |
| Coal, hard, bright; No. 11 coal bed. No. 11 coal bed at depth of 2,047-2,051 ft | 2,429-2,433 |
| Claystone, light-gray; interbedded with light-gray siltstone in lower part | 2,433-2,436 |
| Sandstone, light-gray, fine-grained | 2,436-2,440 |
| Siltstone, light- to dark-gray, shaley | 2,440-2,464 |
| Shale, light-gray to black at base | 2,464-2,473 |
| Coal, No. 10 coal bed | 2,473-2,474 |
| Shale and siltstone, dark-gray | 2,474-2,487 |
| Sandstone, medium-gray, fine-grained | 2,487-2,522 |
| Shale, dark-gray, silty; carbonaceous in lower part | 2,522-2,548 |
| Coal, bright, hard; No. 9 coal bed | 2,548-2,553 |

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# SIGNAL GRANITE (PRECAMBRIAN), WEST-CENTRAL ARIZONA

By Ivo Lucchitta and Neil Suneson

## INTRODUCTION

The name Signal Granite is given here to the granitic batholith that is at least 30 km in diameter and is best exposed in the northern part of the Artillery Mountains in the Artillery Peak 15-min quadrangle in west-central Arizona, between the ghost town of Signal (SW 1/4 sec. 9, T. 13 N., R. 13 W.) and Eagle Point (E 1/2 sec. 7, T. 12 N., R. 13 W.) (fig. 18). These exposures, considered its type locality, can be reached by jeep trail from near Signal and by the poorly maintained Old Alamo dirt road. Formalizing the name of the Signal Granite and describing its occurrence, geologic relations, lithology, and assigned age will make future references to this body simpler and more convenient.

## GEOLOGIC SETTING AND LITHOLOGY

Geologic studies and field mapping on which this report is based indicate that the northern contact of the pluton is intrusive and occurs in the northwestern part of the Artillery Peak 15-min quadrangle and in the northeastern part of the Castaneda Hills 15-min quadrangle (fig. 18). The country rock consists of gneiss, schist, and plutonic rocks typical of the Precambrian terrane of the Hualapai Mountains of the southwestern edge of the Colorado Plateau. The eastern extent of the pluton is unknown because no mapping has been done in that direction. The western contact is poorly exposed and structurally complex and occurs approximately along the 113° 50' west meridian in the Castaneda Hills quadrangle. Southward, the granite becomes increasingly sheared, altered, and mylonitized and is believed to form the protolith of the mylonitic quartz-feldspar gneiss of the Rawhide Mountains (Lucchitta and Suneson, unpub. data, 1978).

The granite is cut by veins of quartz, pegmatite, and aplite and by dikes of diabase, metarhyolite porphyry, and leucogranite. The Signal Granite is a light-gray to medium-brownish-gray, leucocratic to mesocratic, unfoliated to weakly foliated, typically massive, locally jointed and spheroidally weathered, medium- to coarse-grained porphyritic biotite granite or monzogranite. Phenocrysts are subequant, typically 2-3 cm long but as much as 5 cm in places, and composed of twinned potassium feldspar. The matrix consists of 0.5 cm plagioclase, gray quartz, and biotite that commonly is oxidized. Modal analysis of one stained slab has yielded the following mineral percentages: potassium feldspar, 36.2 percent; plagioclase, 25.7 percent; quartz, 27.5 percent; and biotite, 10.5 percent (Suneson, 1980). A coarse-grained equigranular phase occurs locally. In places, the granite contains sparse xenoliths of melanocratic diorite, biotite schist, and leuco- to mesocratic gneiss. Leucocratic and mesocratic phases of the granite form bodies whose dimensions measure hundreds of meters. The granite is cut by numerous shear zones that are mineralized locally and show weak recrystallization. Zones of closely spaced subparallel joints are present in places.

U.S. Geol. Survey Bull. 1529-H, 1982, p. H87-H90

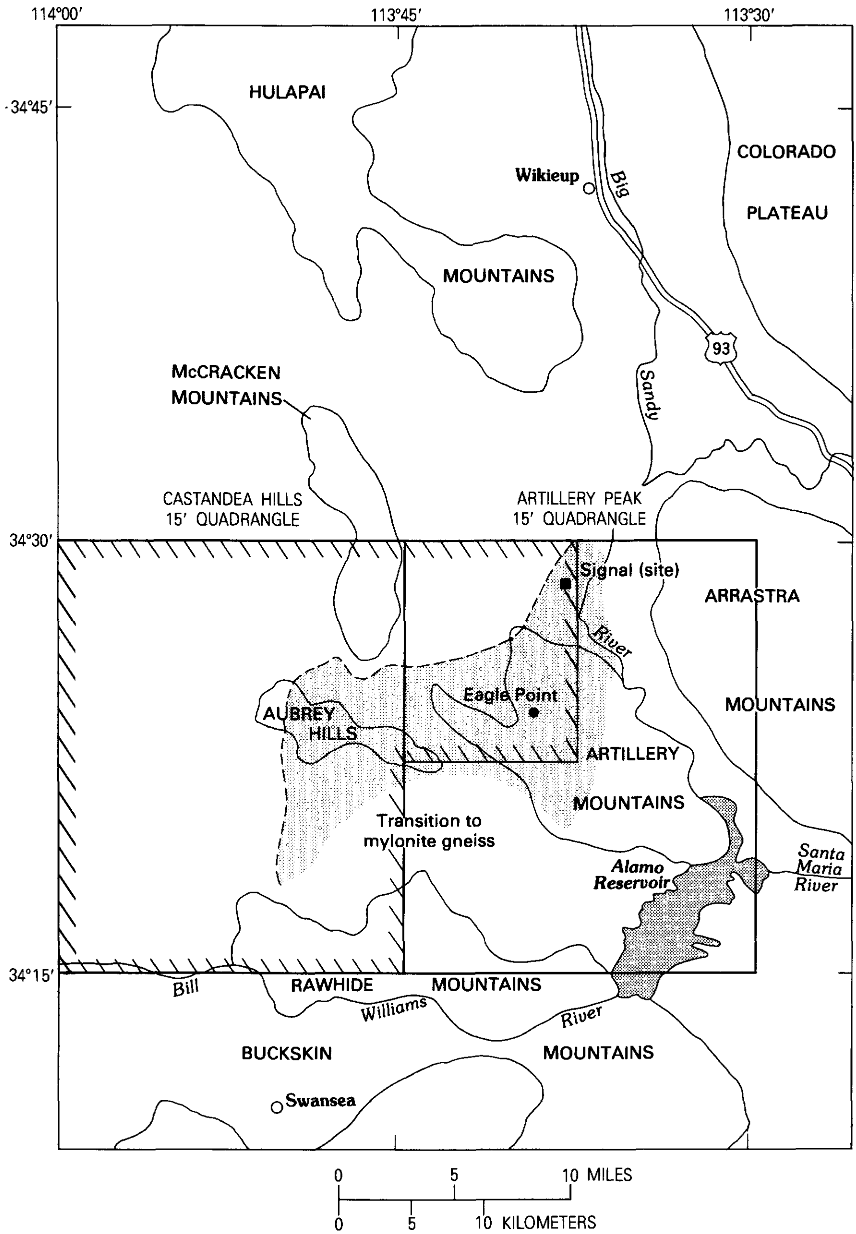


FIGURE 18.—Known outcrop areas of Signal Granite, west-central Arizona (stippled area). Area outlined with hachures has been mapped (Lucchitta and Suneson, unpub. data).

## AGE

Rocks here defined as the Signal Granite were assigned to the Precambrian by Lasky and Webber (1949) and by Wilson and others (1969). In both cases, however, the age assignment probably was made primarily on the grounds that basement plutonic and metamorphic rocks are Precambrian because a Precambrian age was assigned also to mylonite gneisses that are very different from those of the typical Precambrian basement of the area. These gneisses are known now to be of late Mesozoic or Tertiary age. Before the advent of routine radiometric dating, assignment of a Precambrian age on such assumptions was customary.

The Signal Granite still has not been dated radiometrically, although such dating is planned. Nevertheless, we assign it a Precambrian age on the following grounds:

1. It is lithologically and structurally similar to granites of known Precambrian age in Arizona (Hualpai [sic][1] Granite of Putnam and Burnham (1963); the Proterozoic Y Dells Granite, dated at 1.3 b.y. by Marvin and Cole (1978); granites near Bagdad, Ariz., described by Anderson and others (1955); the Oracle Granite of Peterson (1938); the Proterozoic Y Ruin Granite, which has a 1.4-b.y. age given by Ludwig and Silver (1977); and the quartz monzonite or granodiorite of Shride (1967)).

2. It is part of a basement terrane continuously traceable, without significant change in lithology, to rocks that underlie the Tapeats Sandstone (Lower and Middle Cambrian) of the Colorado Plateau margin, less than 30 km away.

3. It is cut by diabase and metarhyolite porphyry dikes that resemble similar rocks unconformably overlain by Cambrian strata in many parts of Arizona.

These criteria do not permit an unequivocal age assignment; the granite could be younger than Precambrian. In this unlikely case, however, the granite probably would be Late Cretaceous and (or) early Tertiary ("Laramide") in age because that is the only known episode of post-Precambrian age that has produced large, deep-seated plutons in northwestern Arizona. Middle Tertiary plutons generally are small and hypabyssal.

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# ADOPTION OF NAMES OF CERTAIN MEMBERS OF FORMATIONS IN THE HAMILTON GROUP (MIDDLE DEVONIAN) OF NEW YORK

By William A. Oliver, Jr.

## INTRODUCTION

Present understanding and nomenclature of Hamilton Group (Middle Devonian) rocks in New York date from Cooper (1930). The marine parts of the group in western and central New York are divided into four formations; in ascending order these are the Marcellus Shale, Skaneateles Shale, Ludlowville Shale (now changed to Formation), and Moscow Shale. These formations have long been used in U.S. Geological Survey publications as defined by Cooper. The history of usage through 1960 is outlined in the lexicons of Wilmarth (1938) and Keroher and others (1966).

Cooper (1930) subdivided each of the Hamilton formations into numerous members, many of which have been adopted for U.S. Geological Survey usage as the need arose. Cooper's members have stood the test of time very well, but later workers have redefined a few units, and additional members have been named. Currently used member nomenclature is shown on the correlation chart of Rickard (1975), although Baird (1979) recently has modified the formational assignment of some members in western New York.

Recent work on coral biostratigraphy in New York (Oliver and Sorauf, 1981) has indicated the need for U.S. Geological Survey adoption of additional member names as summarized in the following section.

## NEW ADOPTIONS

### Kashong Shale Member (of Moscow Shale)

This name is adopted as originally defined by Cooper (1930, p. 231-232) and as used by subsequent workers (for example, Boardman, 1960; p. 7; Rickard, 1964, 1975; Oliver and others, 1969; Baird, 1979, p. 12-14). Baird's recent description and interpretation is the most complete. The Kashong is underlain by the Menteth Limestone Member and the Portland Point Limestone Member of Cooper (1930) and is overlain by the Windom Member, all of the Moscow Shale.

The type section is in Kashong Creek on the west side of Seneca Lake, Yates County, where it is 24 ft thick (Cooper, 1930, p. 231).

### Otisco Shale Member (of Ludlowville Formation)

This member is adopted as originally defined by Smith (1935, p. 45-47) and accepted by Cooper (in Cooper and others, 1942, chart). It was used with the same meaning by Oliver (1951), Rickard (1964, 1975), and Oliver and others (1969). The most recent description and discussion is by Grasso (1978, p. 144-146). The member is underlain by the Centerfield Limestone Member and overlain by the Ivy Point Member of Smith (1935), both of the Ludlowville Formation.

The type section is in Millers Place ravine on the west side of Otisco Lake, 1 mi northwest of the former causeway, Cayuga County, where it is 150 ft thick (Smith, 1935, p. 45).

**Pompey Shale Member (of Skaneateles Shale)**

This member is adopted as originally defined by Cooper (1930, p. 220-221) and as subsequently used by Rickard (1964, 1975) and by Oliver and others (1969). The member is underlain by the Delphi Station Member of Cooper (in Cooper and others, 1942, chart; name replacing Delphi Member of Cooper, 1930) and is overlain by the Butternut Shale Member of Cooper (in Cooper and others, 1942, chart; name replacing Berwyn Member of Cooper, 1930), all of the Skaneateles Formation.

The type section is at the top of Pratt Falls, Onondaga County (Oran 7 1/2-min quadrangle), where the member is 60 ft thick (Cooper, 1930, p. 220).

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# CORRELATION OF EASTFORD GNEISS WITH CANTERBURY GNEISS, EASTERN CONNECTICUT

By Maurice H. Pease, Jr.

Gregory (in Rice and Gregory, 1906) assigned the name Eastford granite gneiss to a body of orthogneiss exposed mostly in the town of Eastford, Conn. He also mapped as Canterbury granite gneiss (in Gregory and Robinson, 1907) several bodies of similar granitic orthogneiss, the largest of which underlies the western part of the town of Canterbury.

Dixon (1968) gives a comprehensive review of the historical background and geographic distribution of the Eastford Gneiss and Canterbury Gneiss, as they were termed at that time. She restricted use of the term Eastford Gneiss to the orthogneiss exposed in the Eastford quadrangle and northern part of the Hampton quadrangle and mapped (in Dixon and Pessl, 1966) these rocks as a separate body not contiguous with exposures of similar orthogneiss to the south.

Dixon also recognized minor differences in mineral composition between this restricted body of Eastford Gneiss and typical Canterbury Gneiss stating that, "In general, the Eastford Gneiss contains more microcline and slightly more muscovite and less biotite and calcium-bearing accessory minerals such as epidote, allanite and sphene" (Dixon, 1968, p. 175) than the Canterbury. In addition to this mineralogical difference, it was postulated that the Eastford Gneiss has a relict primary foliation not present in the Canterbury Gneiss and that, on this basis, the Eastford is a separate and slightly older pluton (Dixon, 1968; Pease, 1972). Dixon (1968, p. 109), however, did recognize that the composition and mode of occurrence of the two gneisses were similar, and stated further that whether they constituted a single body was still not certain.

More recent detailed mapping in this area (fig. 19; Pease and Fahey, 1978) has shown that the orthogneiss mapped as Eastford Gneiss in the Eastford quadrangle (Pease, 1972) and in the northwestern part of the Hampton quadrangle (Dixon and Pessl, 1966) extends into the Spring Hill quadrangle where it is contiguous with rocks mapped as Canterbury Gneiss. These two rock units cannot be distinguished reliably in outcrop, and studies of core samples from drill holes in the critical area and of samples from a gas-transmission-line trench have shown no persistent differences between samples of Eastford and Canterbury (Pease, 1980). Furthermore, although no statistical assessment has been made, greater differences in mineral composition and degree of foliation have been observed within the various exposures of orthogneiss than the slight differences observed between the Eastford and the Canterbury.

Radiometric age determinations also indicate that the Canterbury and the Eastford are probably one and the same. Zartman (in Zartman and others, 1965) originally reported Rb-Sr whole-rock isochron ages for the Canterbury as 430 m.y.+20 m.y. and for the Eastford as approximately 380 m.y. More recently, Zartman (personal commun., 1979), using zircons, derived $^{207}Pb/^{206}Pb$ data from Canterbury and Eastford samples that plot

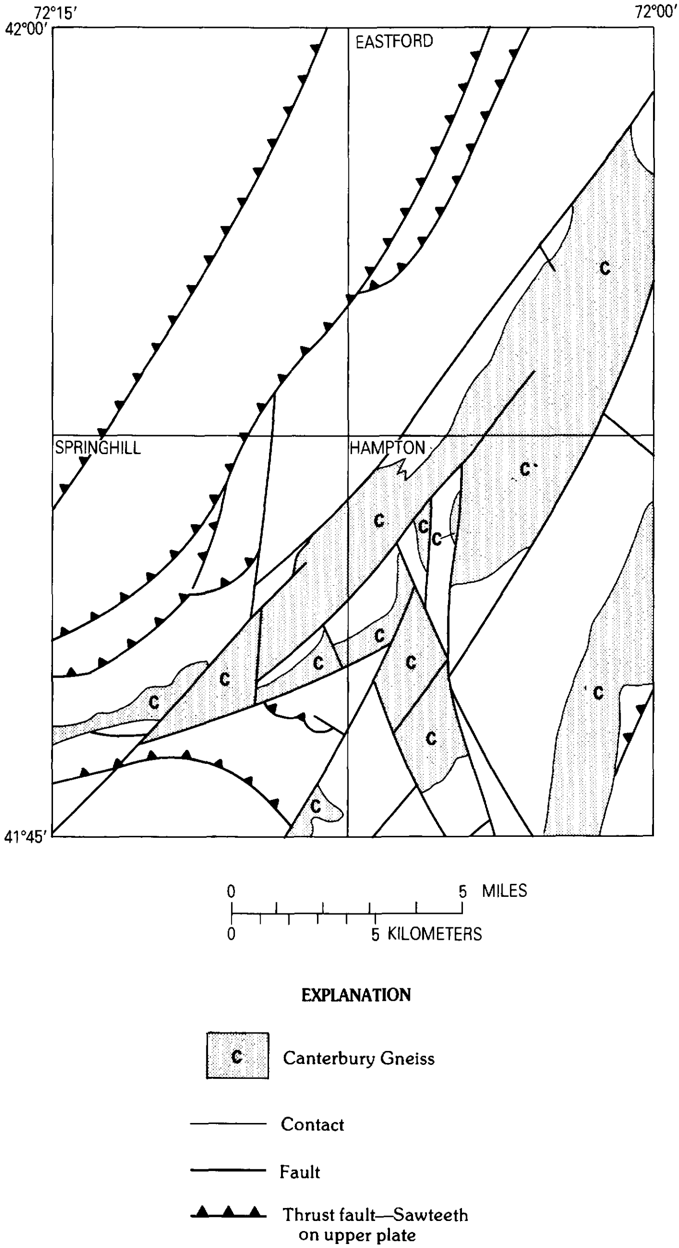


FIGURE 19.—Distribution of Canterbury Gneiss in the Eastford, Hampton, and Spring Hill quadrangles, New York. Geology modified from Pease (1978).

reasonably well as a linear array on a concordia diagram yielding an age of 400 m.y.±10 m.y. Zartman (personal commun., 1979) also redetermined the Rb-Sr whole-rock isochron ages by using more samples and more precise analytical techniques. He was able to derive a colinear single isochron diagram of 392 m.y.±9 m.y. for all of the samples that is in close agreement with the $^{207}Pb/^{206}Pb$ age. Independent isochrons for the Canterbury and the Eastford yielded values of 403 m.y.±28 m.y. and 389 m.y.±14 m.y., respectively. Each separate isochron is less colinear than the combined isochron, and they lie well within the margin of analytical error for a single isochron age. Zartman (personal commun., 1979) concluded that the Canterbury and the Eastford probably were intruded at the same time.

Thus, the criteria used to separate the body of orthogneiss mapped as Eastford from the several bodies of orthogneiss mapped as Canterbury appear to be insufficient to warrant the distinction. Use of the name Eastford Gneiss, therefore, is abandoned herein; and use of the name Canterbury Gneiss herein is extended geographically to include those rocks formerly mapped as Eastford.

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# REVISION OF UPPER CRETACEOUS NOMENCLATURE IN MONTANA AND SOUTH DAKOTA

By Dudley D. Rice, George W. Shurr, and Donald L. Gautier

Changes in Upper Cretaceous nomenclature are necessary to facilitate the presentation of results of recent and future stratigraphic and sedimentological studies in Montana and South Dakota. The changes pertain to the Belle Fourche Shale, Greenhorn Formation, and Mosby Sandstone Member of Cenomanian and Turonian Age and with the Groat Sandstone Bed of Campanian Age (fig. 20).

In central Montana, the Belle Fourche Shale and Greenhorn Formation are redefined. Their contact is moved upward from the base of the Mosby Sandstone Member where the Mosby occurs (fig. 21) to the top of noncalcareous shale that overlies the Mosby. The change is based on lithology. Thus, the Mosby is removed from the Greenhorn and reassigned to the Belle Fourche. The Greenhorn Formation, as redefined, is composed primarily of calcareous shale and limestone, whereas the Belle Fourche Shale is mainly noncalcareous shale with minor siltstone and sandstone, particularly in the upper part.

In outcrop, the Mosby Sandstone Member commonly consists chiefly of two sandstone beds that contain a distinct gastropod fauna developed in the uppermost 10 m of the Belle Fourche (Cobban, 1953). The upper contact of the Mosby is the top of a unit of noncalcareous shale that overlies the sandstone beds and coincides with the top of the Belle Fourche Shale. As a result of detailed outcrop and subsurface studies, the Mosby hereafter is extended to include all sandstone beds developed in the Belle Fourche that are approximately confined to the upper 60 m. If more than one sandstone is present, the member also includes noncalcareous siltstone and shale between and above the sandstones. Logs in figure 22 show the presence of several sandstones in the Mosby in north-central Montana.

Sandstone that crops out 45 m below the top of the Gammon Ferruginous Member of Pierre Shale along the northern flank of the Black Hills has been named the Groat Sandstone Bed (Rubey, 1930). The term is not widely used and is hereby restricted to outcrop. The name Shannon Sandstone Member is hereby extended from Wyoming to the subsurface of east-central Montana and northwestern South Dakota as a member of the Gammon Shale and applied to sandstone equivalent to the Groat in the subsurface. This extension is justified because (1) the name Shannon was adopted earlier than the Groat (Wegemann, 1911), (2) the term Shannon commonly is used in the northern Great Plains, and (3) the sandstone in the subsurface of Montana and South Dakota correlates with the type Shannon of Wyoming (Parker, 1958) and lies within the same ammonite zones (Gill and Cobban, 1973). The name Shannon Sandstone Member in Montana and South Dakota is restricted to discontinuous but widespread sandstone bodies that are equivalent to and lie eastward (seaward) of more continuous sandstone beds in the Eagle Sandstone. The Shannon, as now used, is laterally continuous and was deposited in the same depositional environment as the

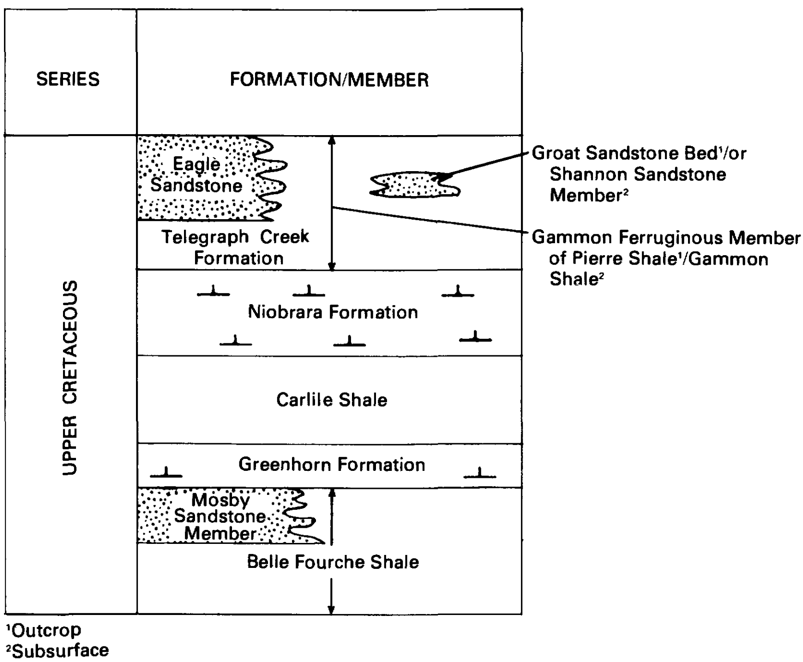


FIGURE 20.–Upper Cretaceous rocks in east-central Montana and northwestern South Dakota.

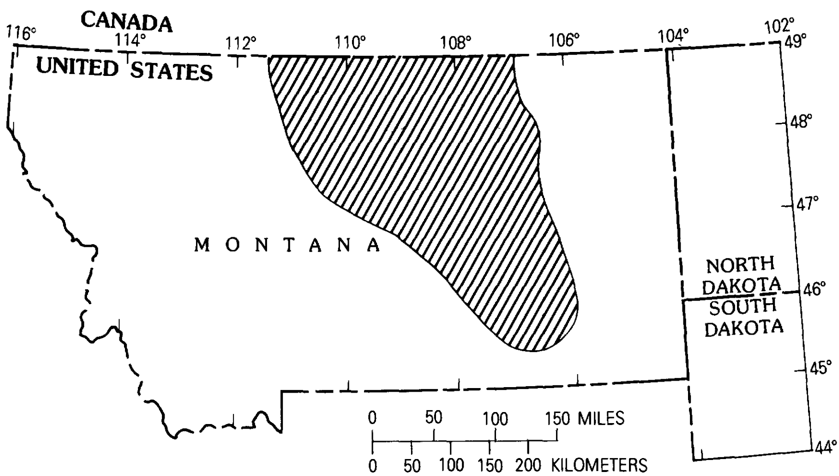


FIGURE 21.—Areal distribution of Mosby Sandstone Member (lined pattern), north-central Montana.

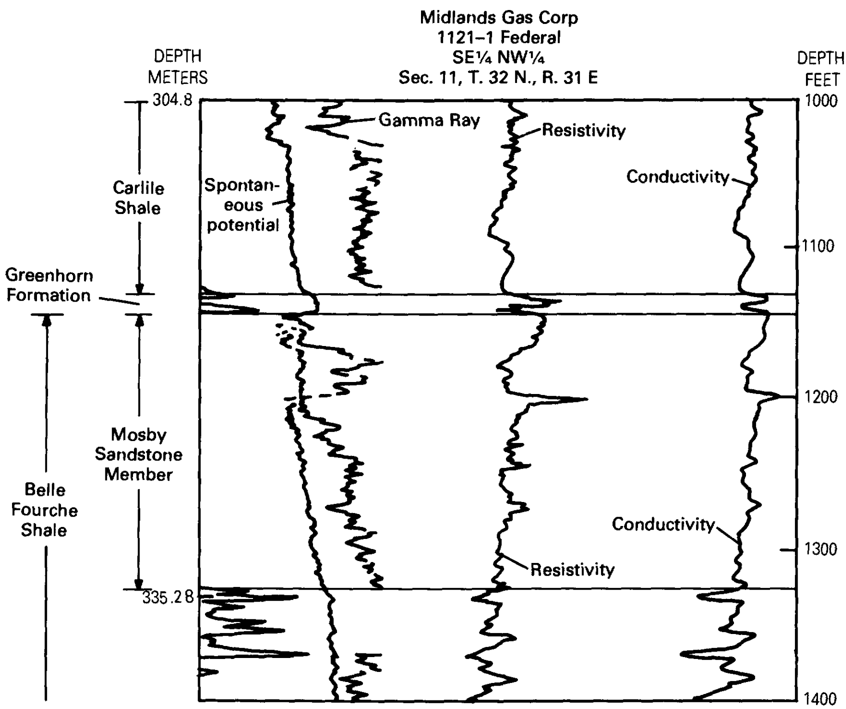


FIGURE 22.–Electric log showing presence of several sandstones in Mosby Sandstone Member, north-central Montana.

Groat Sandstone Bed in outcrop. The approximate areal distribution of the Shannon Sandstone Member and Groat Sandstone Bed in Montana and South Dakota is shown in figure 23. Where the term Shannon Sandstone Member is applied in Montana and South Dakota, the Gammon is raised to the rank of formation and named the Gammon Shale. The Shannon remains a member of the Cody and Steele Shales in the Big Horn and Powder River Basins of Wyoming, respectively.

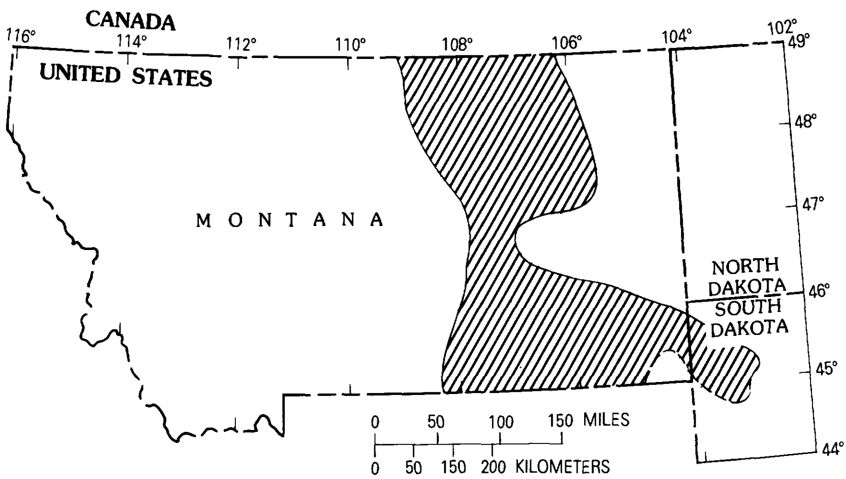


FIGURE 23. – Areal distribution of Shannon Sandstone Member (lined pattern) and Groat Sandstone Bed (dashed line), Montana and South Dakota.

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# CHANDLER BRIDGE FORMATION—A NEW OLIGOCENE STRATIGRAPHIC UNIT IN THE LOWER COASTAL PLAIN OF SOUTH CAROLINA

By Albert E. Sanders[1], Robert E. Weems, and Earl M. Lemon, Jr.

## ABSTRACT

The name Chandler Bridge Formation is introduced herein for a thin sequence of noncalcareous arenaceous beds that lies disconformably above the Ashley Member (Oligocene) of the Cooper Formation and disconformably below post-Oligocene deposits northwest of Charleston, S.C. A diverse fauna of well-preserved whales indicates that the Chandler Bridge is of late Oligocene (Chattian) age.

## INTRODUCTION

The name Chandler Bridge Formation is introduced here for a sequence of noncalcareous arenaceous beds as much as 5 m thick that disconformably overlies the Ashley Member (Oligocene) of the Cooper Formation (Ward and others, 1979) in the Stallsville, Ladson, Johns Island, and Mount Holly 7 1/2-min quadrangles, Dorchester, Berkeley, and Charleston Counties, S.C. (fig. 24). The formation is disconformably overlain throughout most of this area by surficial Pleistocene deposits.

The Chandler Bridge Formation can be divided into three conformable beds: bed 1, a basal, dark-yellowish-brown (10YR 4/2)[2], clayey, slightly calcareous, fine-grained quartz-phosphate sand; bed 2, a middle, medium-gray (N5), clayey, noncalcareous, poorly compacted, fine-grained quartz-phosphate sand; and bed 3, an upper, medium-gray (N5), clayey, noncalcareous, well-compacted, medium-grained quartz-phosphate sand. Phosphate nodules ranging from 10 to 50 mm in diameter are abundant in beds 2 and 3. Bed 1 occurs throughout the type area, but it is only sporadic elsewhere in the subcrop belt and may be discontinuous even within a single subcrop. Its maximum known thickness is 30 cm. Bed 2 ranges in thickness from 30 to 60 cm, whereas bed 3 may be as thick as 4 m. The maximum known thickness for the entire formation is about 5 m.

The beds described here as the Chandler Bridge Formation possibly were included in the Ladson Formation as defined by Malde (1959), but reevaluation of his detailed sections suggests that he did not actually penetrate the Chandler Bridge in any of his augered holes. Because the Chandler Bridge is preserved sporadically in low areas on the surface of the Cooper Formation, it is not surprising that these beds were not reported. Although generally similar in gross appearance to some of the Pleistocene sediments that overlie it, the Chandler Bridge Formation, unlike the Pleistocene sediments, contains abundant sand-sized rounded phosphate grains and has a much finer quartz-sand fraction. Its diverse fauna of

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[2]Color designations are based on the "Rock-Color Chart" of the National Research Council (Goddard and others, 1948).

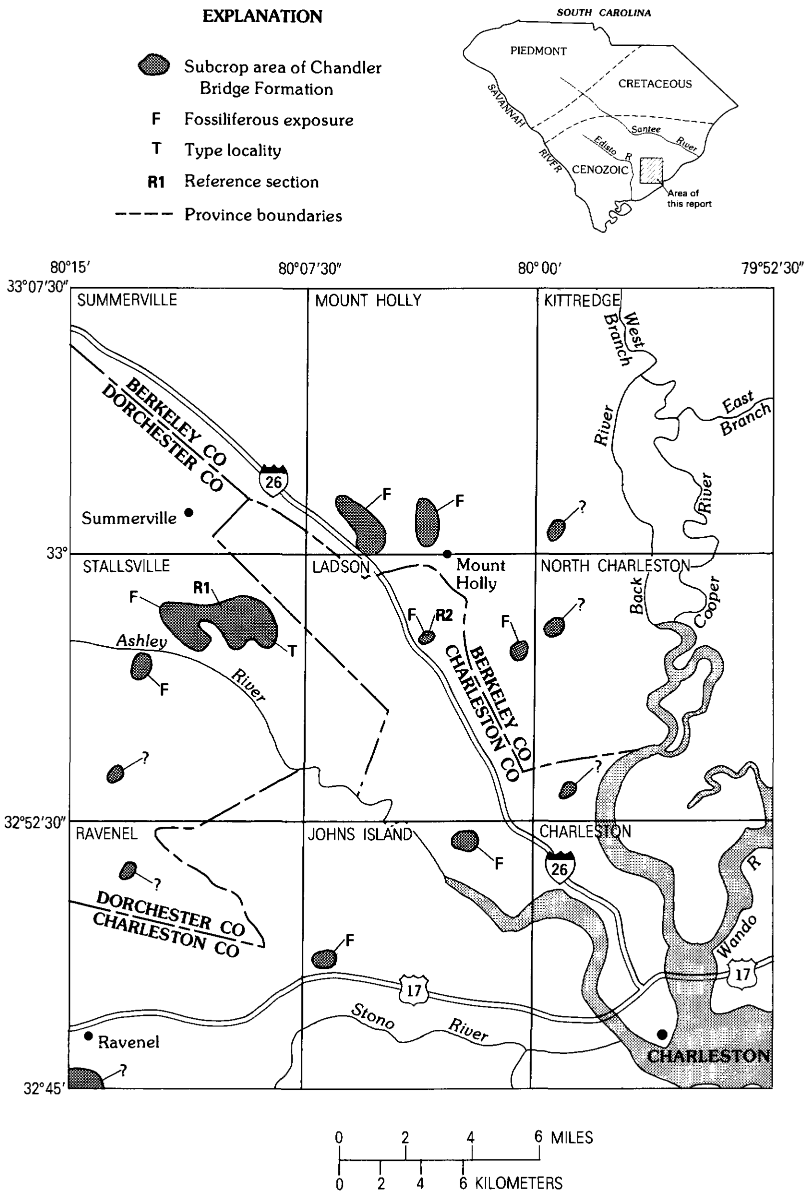


FIGURE 24.—Report area, eastern South Carolina, including 7 ½-minute quadrangles and known subcrop areas of the Chandler Bridge Formation. At the locality in the northwestern North Charleston quadrangle labeled with a question mark, Malde (1959) reported sand beneath the "upper Miocene Duplin Marl" (equivalent to lower Pliocene "Goose Creek Marl" of Sloan, 1908) that possibly belongs with the Chandler Bridge Formation. Other unfossiliferous but lithologically identical subcrops, penetrated during drilling in this area, are labeled with question marks and probably belong to this unit.

archaeocete and odontocete whales establishes the age of this unit as Oligocene (Sanders and others, 1979). The occurrence in the Chandler Bridge of only one genus and none of the species found in the underlying Ashley Member of the Cooper Formation (listed in Whitmore and Sanders, 1976) strongly suggests that a significant unconformity exists between these two units.

## TYPE SECTION AND REFERENCE SECTIONS

The Chandler Bridge Formation is named here for exposures in a Charleston Museum paleontological excavation along Chandler Bridge Creek, 0.7 km northwest of of the confluence of that creek with Eagle Creek, in the northeast quarter of the Stallsville 7 1/2-min quadrangle, Dorchester County, S.C. (figs. 24, 25). At the time of excavation, these beds were informally designated as the "Eagle Creek beds" (Whitmore and Sanders, 1976), but this name has been abandoned as it has been used widely elsewhere in the geological literature and is preempted as a formal geologic name. Also, at the time of excavation, the recognizable lithologies were field designated "zone 1" (equivalent to Ashley Member, Cooper Formation), "zone 2" (equivalent to bed 1 of this report), "zone 3" (equivalent to bed 2), and "zone 4" (equivalent to bed 3). This terminology appears in figure 26, a photograph taken at the time of excavation. These zones do not have formal status.

Nowhere does this unit crop out naturally, but data from channelization ditches and borings indicate that the unit is more or less continuous throughout the northeast quarter of the Stallsville quadrangle (fig. 24). Because all these manmade outcrops are evanescent, the northeast quarter is designated as a type area containing the type section. A sandpit southeast of the Creekside Trailer Park in the north-central part of the Stallsville quadrangle and an excavation site in the northeast quarter of the Ladson 7 1/2-min quadrangle are designated here as reference sections (fig. 24), although at these localities all the constituent beds are not present.

## TYPE SECTION OF THE CHANDLER BRIDGE FORMATION

[Measured at the Charleston Museum paleontologic excavation site 0.7 km northwest of the confluence of Chandler Bridge Creek and Eagle Creek, 30 m northeast of Chandler Bridge Creek, Stallsville 7 1/2-min quadrangle (northwest of Charleston), Dorchester County, S.C. Surface altitude is about 6 m].

| Unit | Description | Thickness (meters) |
|---|---|---|
| Pleistocene deposits: | | |
| 7. | Topsoil | 0.15 |
| 6. | Hardpan, clayey subsoil | 0.35 |
| 5. | Quartz sand, medium-gray (N5), medium-grained, angular, massive, clayey, noncalcareous, well-compacted; basal quartz and phosphate pebble bed 0.05 m thick | 0.65 |

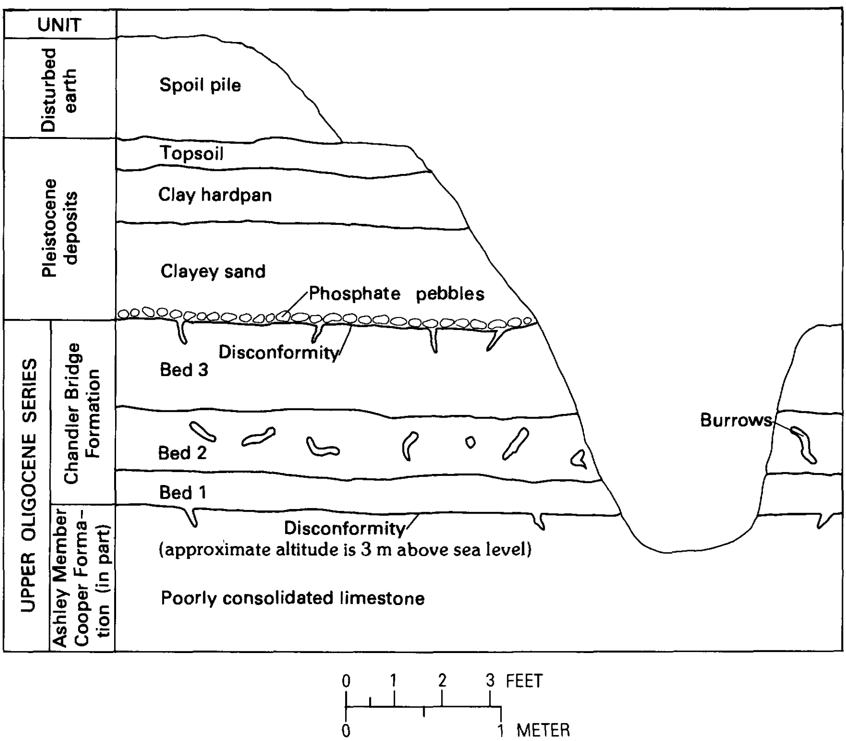


FIGURE 25.–Paleontologic excavation site designated as the type section of the Chandler Bridge Formation, lower coastal plain of South Carolina. No vertical exaggeration.

Chandler Bridge Formation (Oligocene):

4. Quartz-phosphate sand (bed 3 = "zone 4"), medium-gray (N5), medium-grained, massive, clayey, noncalcareous, well-compacted; abundant subrounded to angular lumps of phosphate . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.55

3. Quartz-phosphate sand (bed 2 = "zone 3"), medium-gray (N5), fine-grained, bioturbated, clayey, noncalcareous, poorly compacted; abundant subrounded to angular lumps of phosphate . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.40

2. Quartz-phosphate sand (bed 1 = "zone 2"), dark-yellowish-brown (10 YR 4/2), fine-grained, massive, clayey, slightly calcareous, poorly compacted; rests on a sharply defined, undulatory surface . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.20-0.30

**Total thickness of the Chandler Bridge Formation** . . . . . . . . . . . . . 1.15-1.25

Cooper Formation, Ashley Member (Oligocene):

1. Limestone ("zone 1"), light-olive-brown (5 Y 5/6), fine-grained, bioturbated, well-compacted but poorly consolidated; contains fine-grained phosphate and quartz sand; top surface burrowed and filled with matrix of overlying unit . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.30-0.40

## REFERENCE SECTION 1 OF THE CHANDLER BRIDGE FORMATION

[Measured 0.8 km southeast of S.C. Route 165 bridge over Sawmill Branch-Dorchester Creek in bank of borrow pit southeast of Creekside Trailer Park, in the Stallsville 7 1/2-min quadrangle (northwest of Charleston), Dorchester County, S.C. Surface altitude about 6 m .]

| Unit | Description | Thickness (meters) |
|---|---|---|

Pleistocene deposits:

6. Topsoil . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.15

5. Quartz sand, medium-gray (N5), medium-grained, angular, massive, clayey, noncalcareous, well-compacted; basal quartz and phosphate pebble bed 0.05 m thick . . . . . . . . . . . . . . 0.60

Chandler Bridge Formation (Oligocene):

4. Absent (bed 3) . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.00

3. Quartz-phosphate sand (bed 2), medium-gray (N5), fine-grained, bioturbated, clayey, noncalcareous, poorly compacted; abundant subrounded to angular lumps of phosphate . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . . 0.70

2. Quartz-phosphate sand (bed 1), dark-yellowish-brown (10 YR 4/2), fine-grained, massive, clayey, slightly calcareous, poorly compacted; rests on a sharply defined, undulatory surface .................................. 0.00-0.10

**Total thickness of the Chandler Bridge Formation** ............. 0.70-0.80

Cooper Formation, Ashley Member (Oligocene):

1. Limestone, light-olive-brown (5 Y 5/6), fine-grained, bioturbated, well compacted but poorly consolidated; contains fine-grained phosphate and quartz sand ........ 0.20-0.30

## REFERENCE SECTION 2 OF THE CHANDLER BRIDGE FORMATION

[Measured 1.6 km north-northwest of junction of Ashley Phosphate Road and Interstate 26, 0.25 km northeast of Interstate 26 in sewerline pit at Northwoods Estates, in the Ladson 7 1/2-min quadrangle (northwest of Charleston), Charleston County, S.C. Altitude about 6 m.]

| Unit | Description | Thickness (meters) |
|---|---|---|
| | Pleistocene deposits: | |
| 6. | Topsoil | 0.15 |
| 5. | Quartz sand, medium-gray (N5), medium-grained, angular, massive, clayey, noncalcareous, well compacted; basal quartz and phosphate pebble bed 0.10 m thick | 1.65 |
| | Chandler Bridge Formation (Oligocene): | |
| 4. | Quartz-phosphate sand (bed 3), medium-gray (N5), medium-grained, massive, clayey, noncalcareous, well compacted; abundant subrounded to angular lumps of phosphate | 0.60 |
| 3. | Quartz-phosphate silt (bed 2), medium-gray (N5), sandy and clayey, bioturbated, noncalcareous, poorly compacted; sand content decreases downward so that basal 0.10 m is a sandy clay; abundant subrounded to angular lumps of phosphate; rests on a sharply defined, nearly planar surface | 0.30 |
| 2. | Absent (bed 1) | 0.00 |
| | **Total thickness of the Chandler Bridge Formation** | 0.90 |
| | Cooper Formation, Ashley Member (Oligocene): | |
| 1. | Limestone, light-olive-brown (5 Y 5/6), micritic, bioturbated, well compacted but poorly consolidated; contains fine-grained phosphate and quartz sand | 0.90 |

## CONTACT RELATIONS

The basal contact of the Chandler Bridge Formation is a sharp, rolling (amplitude about 5 cm), sparsely burrowed surface separating sparsely calcareous or noncalcareous, dark-yellowish-brown (bed 1) or medium-gray (bed 2) sand from the underlying light-olive-brown, highly calcareous sand of the Ashley Member of the Cooper (fig. 26, 27). This contact is interpreted as a disconformity. This interpretation is supported by the lack of any whale species (and only one genus) common to the two units, which suggests a significant time gap between the units.

The top of the Chandler Bridge Formation is disconformably overlain in all studied exposures by surficial Pleistocene deposits. This contact is marked by a basal Pleistocene bed of coarse-grained quartz sand containing many subrounded to discoidal quartz pebbles and cobbles and many rounded to angular phosphate pebbles, cobbles, and boulders. At some localities, the top of the Chandler Bridge Formation contains burrows filled with Pleistocene sediment and mollusk shells. Textural analysis of Chandler Bridge sediments from near the upper contact reveals a significant percentage of sand that is typical of the modal size range of the overlying Pleistocene sediments and that is coarser than the sand found elsewhere (downward) in the Chandler Bridge beds. The presence of this sand suggests that bioturbation of the Chandler Bridge-Pleistocene contact has taken place even where burrows are not readily visible near the upper contact.

Malde (1959) reported a thin sand unit similar to the Chandler Bridge Formation beneath the "upper Miocene Duplin marl" (equivalent to the lower Pliocene "Goose Creek marl" of Sloan, 1908) along the route of the F. B. McDowell, Jr., Tunnel in the western part of the North Charleston 7 1/2-min quadrangle. Because this bed is only 1.4 km northeast of a fossiliferous Chandler Bridge locality of comparable altitude in the eastern Ladson 7 1/2-min quadrangle, the unit mentioned may be assignable to the Chandler Bridge Formation. If so, at least locally in the North Charleston 7 1/2-min quadrangle, the Chandler Bridge Formation may be overlain by the "Goose Creek marl" of Sloan. Though not exposed, the contact between the Chandler Bridge Formation and the "Goose Creek marl" is presumed to be a disconformity.

## LITHOLOGIC FEATURES

In contrast to the Pleistocene sediments, which have a median grain size of 2 to 2.2 $\phi$ for the quartz-sand fraction, the sediments of the Chandler Bridge Formation have a median grain size of 3.3 to 3.5 $\phi$ for the quartz-sand fraction. Generally, the phosphate-sand fraction of the Chandler Bridge is coarser than the quartz-sand fraction and consists of foraminifer internal molds and bone fragments. The quartz- and phosphate-sand fractions of the Cooper Formation (Ashley Member) are similar to those of the Chandler Bridge. Both the upper bed and the lower bed of the Chandler Bridge Formation are massive and show no obvious evidence of bioturbation or crossbedding. In contrast, the middle bed shows extensive evidence of *Thallasanoides*-type burrowing. The lower and middle beds are soft and crumble readily when dug, but the upper bed is dense and cannot be excavated as easily. The clay content of the upper bed is not obviously higher than that of the other beds (table 3), so the difference in the cohesion

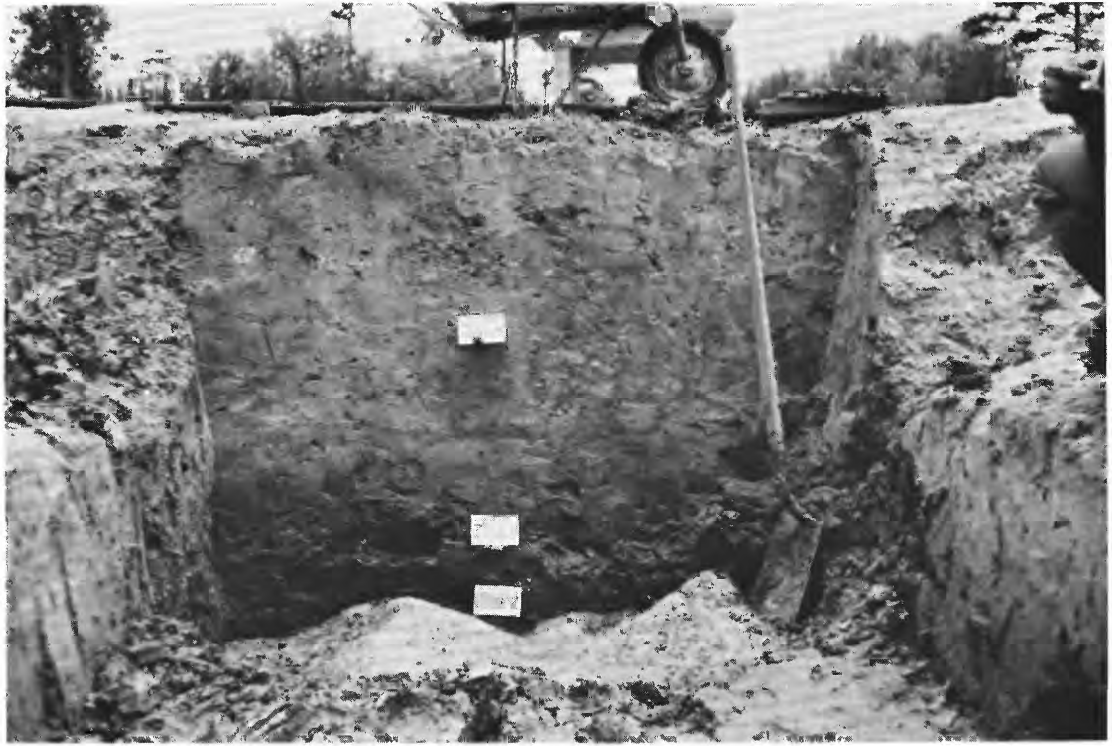

FIGURE 26. – Type section of the Chandler Bridge Formation showing the basal contact with the Ashley Member of the Cooper Formation ("zone 1," not labeled) and the three beds (beds 1–3 equivalent to "zones 2–4") composing the Chandler Bridge Formation. The bottom of each "zone" card marks the basal contact of the corresponding bed.

FIGURE 27. – View from above of the section shown in figure 26, showing the top surface of the Ashley Member of the Cooper Formation. Note large burrow next to shovel.

of the beds is attributed here tentatively to tighter grain packing in the upper bed. The upper bed also differs from the lower two in that it contains a dominantly medium-grained phosphate-sand fraction as opposed to a dominantly fine-grained phosphate-sand fraction in the lower two beds.

## MINERALOGY

The Chandler Bridge Formation is noncalcareous except for the basal bed, which is slightly calcareous. Because the basal bed is thin and directly overlies the highly calcareous Cooper Formation, the calcite content of the basal bed is possibly the result of direct reworking of Cooper sediments. The sand fraction of the Chandler Bridge Formation consists almost exclusively of quartz and phosphate but contains some very stable detrital heavy minerals (ilmenite and sillimanite) or their alteration products. In all beds, quartz predominates by weight over phosphate, but generally, within any one bed, phosphate is dominant in the relatively coarser fractions (except where weathering has selectively leached the phosphate), and quartz is dominant in the relatively finer fractions. The relative proportions of quartz and phosphate in the Chandler Bridge Formation vary widely with stratigraphic position; the quartz fraction ranges from 15 to 50 percent dry weight, and the phosphate fraction, from 5 to 35 percent. The quartz-sand fraction in the underlying Cooper sediments generally ranges from 10 to 25 percent.

The Chandler Bridge silt fraction also is dominated by quartz and phosphate, but the clay fraction consists mostly of interlayered illite-smectite having high percentages of expandable layers and small amounts of kaolinite and phosphate. In contrast, the clay-sized fraction of the Cooper sediments contains sepiolite, palygorskite, calcite, apatite, kaolinite, and illite-smectite. Bed 1 is exceptional in that it contains less kaolinite than either the beds above it or the Cooper below it. Clay minerals in the Pleistocene sediments are typically kaolinite, illite, gibbsite, and vermiculite (Lemon, 1979). The mineralogy of the Chandler Bridge Formation is summarized in table 3.

## FAUNA AND FLORA

Because the Chandler Bridge Formation is thin and permeable and thus typically leached of carbonate, attempts to recover a calcareous microfauna or microflora have been either unsuccessful or equivocal in that the few specimens obtained may represent material reworked from the underlying Ashley Member of the Cooper into bed 1. Unworn phosphatized molds of solitary corals (Balanophyllia sp.), snails, and pelecypods (some of the last still retaining original but decayed shell material around them) were found at the type section. Specimens of these three groups appear to be grossly similar to forms found in the Ashley Member (Druid Wilson, oral commun., 1970) but have not yet been studied in detail.

By far the most abundant and best preserved components of the fauna are vertebrates. Ray plates (Aetobatis, Myliobatis), sharks' teeth (Carcharodon, Galeocerdo, Odontaspis, and others), scombroid skeletons, and other unidentified bony-fish remains are abundant as are skeletal remains representing at least four genera of sea turtles. A single femur of a crocodilian is the only indication thus far of these animals in the Chandler Bridge. Bones of birds also were recovered at the type section, but studies

Table 3.--Texture and mineralogy of selected samples of the Pleistocene deposits and type and reference sections of the Chandler Bridge Formation and the Ashley Member of the Cooper Formation

[Values are in weight percent]

| Bed and section | Sand-sized fraction: Mineralogy: Quartz | Sand-sized fraction: Mineralogy: Apatite | Sand-sized fraction: Texture[1]: VF-F | Sand-sized fraction: Texture[1]: Med | Sand-sized fraction: Texture[1]: C-VC | Sand-sized fraction: Texture[1]: Total | Silt-sized fraction | Clay-sized fraction: Mineralogy[2]: V | Clay-sized fraction: Mineralogy[2]: K | Clay-sized fraction: Mineralogy[2]: I | Clay-sized fraction: Mineralogy[2]: I/S | Clay-sized fraction: Percent expandable | Clay-sized fraction: Percent clay in sample |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| **Pleistocene deposits** | | | | | | | | | | | | | |
| Reference 1 | 100 | 0 | 8 | 44 | 10 | 62 | 5 | 5 | 80 | 0 | 15 | 61 | 33 |
| Reference 2 | 98 | 2 | 10 | 42 | 6 | 58 | 21 | 44 | 44 | 12 | 0 | 0 | 21 |
| **Chandler Bridge Formation** | | | | | | | | | | | | | |
| Bed 3 Type | 67 | 33 | 30 | 35 | 4 | 69 | 15 | 0 | 45 | 8 | 47 | 62 | 16 |
| Reference 2 | 60 | 40 | 7 | 38 | 8 | 53 | 27 | 0 | 47 | 11 | 42 | 65 | 20 |
| Bed 2 Type | 73 | 27 | 52 | 13 | 3 | 68 | 25 | 0 | 19 | 30 | 51 | 64 | 7 |
| Reference 1 | 79 | 21 | 54 | 9 | 0 | 63 | 8 | 0 | 47 | 9 | 44 | 56 | 29 |
| Reference 2 (upper) | 70 | 30 | 18 | 14 | 2 | 34 | 46 | 0 | 30 | 18 | 52 | 67 | 20 |
| Reference 2 (lower) | 75 | 25 | 22 | 1 | 0 | 23 | 24 | 0 | 22 | 23 | 55 | 64 | 53 |
| Bed 1 Type | 63 | 37 | 48 | 10 | 1 | 59 | 28 | 0 | 14 | 29 | 57 | 68 | 13 |
| Reference 1 | 54 | 46 | 56 | 16 | 2 | 74 | 13 | 0 | 19 | 26 | 55 | 62 | 13 |
| **Cooper Formation, Ashley Member** | | | | | | | | | | | | | |
| Type | 23 | -- | 32 | 13 | 2 | 47 | 41 | 0 | 0 | 0 | 0 | 0 | 3/ 12 |
| Reference 2 | 30 | -- | 30 | 6 | 0 | 36 | 7 | 0 | 27 | 20 | 53 | 65 | 57 |

1/ VF = very fine, F = fine, Med = medium, C = coarse, VC = very coarse

2/ V = vermiculite, K = kaolinite, I = illite, I/S = illite-smectite; determined from relative peak heights of X-ray traces.

3/ Clay-sized fraction is calcite.

of these specimens have not been completed. Among mammals, sirenian bones are common; some isolated elements of these animals can be referred to Halitherium and Metaxytherium. Whales are abundantly represented; taxa include at least two species of archaeocetes, four genera of squalodontoids, and a species of the taxonomically problematical genus Xenorophus, which is not the same as X. sloanii found in the underlying Ashley Member of the Cooper Formation (Sanders, 1980).

Plant remains are also present in the Chandler Bridge Formation. At the type section, only remains of acritarchs were recovered from samples submitted for microfloral analysis (F. E. May, written commun., 1975). The presence of small (as much as 18 microns in diameter) sphaeromorphs and small (as much as 28 microns in diameter) acanthomorphs bearing very short (as much as 3 microns) spines was noted. The megafossil flora includes 20 to 30 seed cases of hickory, several acorns, and a single seed of a grape (Vitis sp.) (J. A. Wolfe, written commun., 1971), all recovered at the type section. At first, we were concerned that these seed remains might have been introduced by Pleistocene burrowing rodents, but careful observation during the excavation of the type locality showed no sign of such burrowing near the seeds. Many acorns and hickory nuts were recovered from bed 1, in which they were completely surrounded by normal bed 1 matrix; burrowing in this distinctively colored and textured unit would have been obvious, but no evidence was found of it.

## ORIGIN OF THE CHANDLER BRIDGE FORMATION

The presence of a vertebrate fauna dominated by whales and, to a lesser extent, by sea turtles, sharks, rays, and scombroid bony fishes indicates that the Chandler Bridge Formation was deposited under marine to marginal-marine conditions. The presence of abundant hickory nuts, acorns, and a grape seed further restricts the likely area of deposition to nearshore marine or lagoonal environments. The acritarch assemblage suggests an inshore, basinal, turbulent environment. Thus, a back-bay to lagoonal environment seems best to accommodate the requirements of all the above faunal and floral components. The finding of many of the whale skeletons in nearly parallel alinement in bed 3 indicates that the whales may have become stranded along some sort of intertidal bar within the marginal-marine environment. The probable tighter grain packing in bed 3, mentioned earlier, might be explained also by accumulation along a wave-sorted intertidal bar.

The modal quartz-grain size found in the Chandler Bridge Formation (3.3 to 3.5 $\phi$) is the same as that in the quartz fraction of the underlying Cooper Formation. This suggests that the Chandler Bridge was derived either from the same source as the Cooper or by reworking of the Cooper from local sources. Because the Chandler Bridge is preserved only in low areas in the Cooper, reworking may be a more likely explanation, unless tectonic downwarping can be demonstrated to account for the present geometry and distribution of the Chandler Bridge beds. Modern rivers from the Piedmont carry large quantities of kaolinite and (or) its feldspar precursors, whereas local coastal rivers (for example, the Ashley River) do not. Therefore, the very low kaolinite content of bed 1 suggests that only a local coastal drainage system was present in early Chandler Bridge time.

## AGE AND CORRELATIONS

Because the Chandler Bridge Formation lacks well-preserved macroinvertebrate, microfaunal, or microfloral assemblages, biostratigraphic correlations cannot be made. Therefore, the age of the Chandler Bridge must be deduced from determinations of the age of the immediately underlying Ashley Member of the Cooper Formation and from the evolutionary grades represented in the rich cetacean fauna of the Chandler Bridge beds. The youngest beds of the Ashley have yielded a well-preserved planktonic foraminiferal fauna (Hazel and others, 1977; C. W. Poag, written commun., 1974) which includes Globigerina angulisuturalis, Turborotalia mendacis, and T. siakensis (all first appearing in zone P21) as well as G. gortanii and T. nana (both disappearing in P22). Thus, the Ashley definitely ranges into zone P21 and cannot be younger than zone P22. In the biostratigraphic correlation charts of Berggren (1972), Steininger and others (1976), and Curry and others (1978), zone P21 extends from uppermost Rupelian (middle Oligocene) to the middle Chattian (upper Oligocene), the base of this interval being marked by the appearance of G. angulisuturalis (Berggren, 1972). The additional presence of Globigerinoides primordius in the Ashley assemblage would have suggested a latest Oligocene (P22) or earliest Miocene (N4) age for the youngest beds of the Cooper Formation (Berggren, 1972), but Steininger and others (1976, p. 179) found the first occurrence of this species "to be worthless as an index of the Oligocene/Miocene boundary," noting that G. primordius "is found in Trinidad and Venezuela within the Globigerinoides ciperoensis ciperoensis zone" (P22) and "in the western Carpathians even together with Globorotalia opima opima (Cicha and others, 1971)." Because the horizon of disappearance of G. o. opima is currently used as the boundary between P21 and P22 (Berggren, 1972; Steininger and others, 1976; Curry and others, 1978) and the because foraminifers reported from the uppermost beds of the Ashley range through P21 and possibly into P22, the uppermost part of the Ashley must at least be referable to P21. It cannot as yet be proven to range into P22 because the absence of G. o. opima may be an artifact of collecting or of paleoenvironmental conditions rather than an indication that this species was extinct by latest Ashley time.

The cetacean fauna of the Chandler Bridge Formation is composed of toothed whales (suborder Odontoceti) referable to the super-family Squalodontoidea Simpson, 1945, and of undescribed species of archaeocetes (suborder Archaeoceti). The odontocete assemblage consists of cranial elements and partial skeletons documenting at least five genera. Three are undescribed and have been designated temporarily as Genus X, Genus Y, and Genus Z (Whitmore and Sanders, 1976); one partial skull is tentatively regarded as a new species of the genus Squalodon Grateloup, 1840; and a new species of the genus Xenorophus Kellogg, 1923, is abundantly represented by cranial and postcranial elements. The type of the last genus, Xenorophus sloanii Kellogg (1923), was described from a partial skull found in the underlying Ashley Member of the Cooper Formation near Woodstock, Charleston County, S.C. Because of the primitive features in the holotype, Xenorophus was placed in the Agorophiidae by Miller (1923, p. 40). However, the more complete cranial material of the new species of Xenorophus has proved that assignment to be incorrect, and, as a result, Xenorophus was placed in incertae sedis by Whitmore and Sanders (1976). The other four genera (Squalodon, Genus X, Genus Y, Genus Z) represent the family

Squalodontidae Brandt, 1873. Virtually all of the odontocete material was found during the Charleston Museum excavation of the Chandler Bridge beds conducted by Sanders (1980) during the summers of 1970-72. Supported by research grants from the National Geographic Society (1971-72) and the Charleston Scientific and Cultural Educational Fund (1970), the project yielded the remains of at least 17 associated partial skeletons (one from bed 2, 16 from bed 3) within the 21- by 21-m area of excavation. Some of the forms represented have been discussed and figured by Whitmore and Sanders (1976).

Archaeocete remains were not found during the excavation at the type section of the Chandler Bridge Formation, but skulls and partial skeletons of three individuals have since been discovered in this unit at other localities near Charleston. One specimen was obtained from bed 3, 2.2 km north-northwest of the type section in 1975, and two were found in bed 2 at separate locations during the summer of 1978. The specimen from bed 3 has not been prepared and thus awaits taxonomic evaluation. Preparation of the two specimens from bed 2 has not been completed, but we already know that they represent two undescribed species that share similarities in dental morphology but differ in size and in details of the auditory bones. The relations between these two forms at the generic level and their affinities with other archaeocetes at the familial level are not clear at present. The most primitive of the three recognized suborders of the Cetacea (Archaeoceti, Mysticeti, Odontoceti), the Archaeoceti are best represented in Eocene beds and do not appear to have been validly recorded previously above the middle Rupelian (middle Oligocene), with the possible exception of Mammalodon colliveri[3].

As one of the largest and most diverse assemblages of cetacean remains yet recorded from an Oligocene formation, the Chandler Bridge material has furnished important new information about the telescoping of the cranial elements in early odontocetes (Whitmore and Sanders, 1976, p. 312-314) (see figs. 24-27). The degrees of telescoping (that is, the progressive movement of the nasal opening and contiguous bones backward to a position at the vertex of the skull) represented in the three new

[3]The archaeocete Phococetus vasconum (Delfortrie) was described from a single tooth from beds in France that were thought to be of Burdigalian (early Miocene) Age (Kellogg, 1936, p. 230, citing Dollfus, 1909, p. 385, 397) but which have since been referred to the middle Rupelian (middle Oligocene) (Richard, 1946, p. 136, 342). Kekenodon onamata Hector, from New Zealand, was originally reported from the Eocene (Hector, 1881) and was later referred to the early Miocene (Kellogg, 1936) but has since been placed in the middle Oligocene (Keyes, 1973). Chonecetus sookensis Russell, from the Sooke Formation (upper Oligocene) of Vancouver Island, British Columbia, was originally described as an archaeocete (Russell, 1968), but the propriety of this assignment has been questioned by Mitchell (Whitmore and Sanders, 1976, p. 305) and by Barnes (1976, p. 324). Aetiocetus cotylalveus Emlong (1966), from the Yaquina Formation (upper Oligocene part) of Oregon, Ferecetotherium kelloggi Mchedlidze (1970), from upper Oligocene beds in the Caucasus region of the Soviet Union, and Mirocetus riabinini Mchedlidze (1970), from upper Oligocene beds of Azerbaijan, were originally described as archaeocetes but were referred to incertae sedis by Whitmore and Sanders (1976, p. 305, 317).

squalodontids (Genus X, Genus Y, Genus Z) from the Chandler Bridge Formation compare most closely with those of squalodontids from the Chattian (upper Oligocene) of Europe (Whitmore and Sanders, 1976; Sanders, 1980).

Genus X represents the same stage of telescoping as that of Eosqualodon langewieschei Rothausen (1968a) (Whitmore and Sanders, 1976, figs. 2b, 5; Sanders, 1980) from Eochattian beds at Doberg bei Bunde (Westfalen) in northwestern Germany (Karlheinz Rothausen, written commun., 1979). The Eochattian sands (Chattian A) at Doberg have been referred to nannoplankton zone NP24 (Martini and Muller, 1975) and are considered to be of early Chattian Age (Curry and others, 1978, p. 46). The genus is also represented by E. latirostris (Capellini, 1904) (Rothausen, 1968b) from the "arenaria calcarifera" (Capellini, 1904, p. 441) at Schio, northern Italy. Formerly considered to be of Miocene age (Capellini, 1904), these deposits since have been placed in the late Oligocene (Rothausen, 1968b, p. 4). Genus X is a smaller form than either of these two species and differs also in having teeth that are similar to those of taxa assigned to the genus Microcetus, the type-genus of which, M. ambiguus (Meyer, 1840), was founded only upon a series of teeth from Doberg (Rothausen, 1961). M. sharkovi Dubrovo and Sharkov (1971), a small squalodontid that also has M. ambiguus-like teeth, was described from a partial skull and two mandibles from upper Oligocene deposits on the Magyshlak Peninsula in western Kazakhstan, U.S.S.R. (Dubrovo and Sharkov, 1971). Because the species represented by Genus X is clearly not referable to M. sharkovi, and because there is no cranial material for the holotype of Microcetus ambiguus, Genus X cannot be safely assigned to Microcetus.

As noted by Sanders (1980), the stage of telescoping exemplified by Eosqualodon and Genus X is represented also in the holotype of Agriocetus incertus (Brandt, 1874), a poorly preserved partial skull from the Linz Sands at Linz, Austria. These sands have been referred to nannoplankton zone NP25 (uppermost Oligocene) by Rabeder and Steininger (1975, p. 177). Sulakocetus dagestanicus Mchedlidze (1976), from upper Oligocene deposits in the northern Caucasus region of the U.S.S.R. (Mchedlidze, 1976, p. 42, fig. 12), and "Prosqualodon" hamiltoni Benham (1937, figs. 1-3), from upper Oligocene beds in New Zealand, represent grades of this stage of telescoping, as suggested by Rothausen (1970, p. 186). "P." hamiltoni is not a valid member of the genus Prosqualodon and needs to be reassigned.

Genus Y from the Chandler Bridge Formation demonstrates a more primitive stage of telescoping in which there is a pronounced intertemporal constriction (Whitmore and Sanders, 1976, fig. 4), a characteristic feature in early odontocetes in which the parietals form a part of the skull roof. During later stages of telescoping, these bones were eliminated gradually from the skull roof, apparently through atrophy, and are restricted to the sides of the skull in odontocetes of Miocene to Holocene time. Among the extinct odontocetes described to date, Genus Y can be compared only with the squalodont Patriocetus ehrlichi (van Beneden, 1865; Rothausen, 1968a, fig. 2a) from the Linz Sands, noted earlier as being of latest Oligocene age (Rabeder and Steininger, 1975). Although these two forms have many differences, they appear to represent the same general stage of telescoping, to which Rothausen (1968a, p. 87) has applied the term "protosqualodontid," in reference to the evolutionary level reflected in the cranial morphology of

Patriocetus. Rothausen (1970, p. 180) suggested that Patriocetus was a survivor "of an evolutionary level that was more usual in earlier times, maybe in the middle Oligocene."

Genus Z, from bed 2 of the Chandler Bridge Formation, documents a stage of telescoping that is morphologically intermediate between Genus X and Genus Y in some respects but is closer to Genus Y in others. The combination of features in the skull of this form is not duplicated in any squalodont described to date, but the effects of the intertemporal constriction upon the adjacent bones are the same as those in Genus Y (Whitmore and Sanders, 1976, p. 312; Sanders, 1980).

By far the most primitive of the odontocetes of the Chandler Bridge Formation, the new species of Xenorophus has the most archaic braincase structure yet recorded in the Odontoceti (Whitmore and Sanders, 1976, fig. 1a). Although telescoping of the rostral elements is well advanced, the relative positions of the bones of the braincase are essentially the same as those in typical land mammals, quite unlike the radically restructured braincase in the toothed whales of Miocene to Holocene time.

The significance of the degrees of telescoping manifested in the Chandler Bridge cetaceans is that this process is complete in all known odontocetes from the lower Miocene. As summarized by Rothausen (1970, p. 187-188):

> "the last distinct [evolutionary] advance is that between the Chattian and the Aquitanian. But the main steps of evolution in the squalodontoidea had already occurred within the Oligocene. In the Chattian there are very clear trends toward Neogene faunal assemblages of the Squalodontoidea. With the Aquitanian [early Miocene] had disappeared also the survivors of various earlier stages of evolution that had still been in existence in both hemispheres during the Chattian."

Although a few as-yet-undiscovered forms having incomplete telescoping may have straggled into the earliest Miocene, the early Miocene odontocetes Squalodon (S. catulli, S. bellunense), Prosqualodon (P. australis), Neosqualodon, Phoberodon, Diochoticus, Acrodelphis, Argyrocetus, Cyrtodelphis, Eurhinodelphis, and Diaphorocetus all possessed fully telescoped crania and together serve to demonstrate that this stage of telescoping was fully established in the odontocetes from the lower Miocene (Kellogg, 1928, p. 62).

The primitive aspects of the Chandler Bridge cetacean fauna seem, therefore, to provide an adequate indication of the age of this formation. Telescoping of the cranial elements is incomplete (that is, the parietals are still in place in the skull roof) in three of the four squalodont genera represented (Genus X, Genus Y, Genus Z) and has not taken place at all in the braincase of Xenorophus. Thus, telescoping is incomplete in four (80 percent) of the five odontocete genera recorded from the type locality, the lone example of completed telescoping being the partial skull tentatively referred to the genus Squalodon. In contrast to this record of a single individual having completed telescoping, Xenorophus is represented by the remains of at least 10 animals, Genus X, by at least 8, and Genus Y, by 4 specimens, one of which is a virtually complete skeleton. Genus Z is known from one individual. When the three archaeocete specimens from the

Chandler Bridge beds are added, the primitive composition of this assemblage becomes even more apparent. The evolutionary levels represented therein are uncharacteristic of Miocene faunas known from elsewhere and correspond much more closely to those of forms from the upper Oligocene of Europe. On these grounds, and in the absence of evidence to the contrary, it seems safe to conclude that the Chandler Bridge Formation is of Chattian (late Oligocene) Age.

Huddlestun (1973) alluded to the presence of an upper (?) Oligocene unit and a lower Miocene unit in the vicinity of the lower Savannah River, Ga. The older ("Alum Bluff") unit belongs in P22 or N4, whereas the younger ("lower Marks Head marl") unit belongs in N4 or N5 (J. E. Hazel written commun., 1979). The older of these units might be, but probably is not, time correlative with the Chandler Bridge Formation; biostratigraphic data are too few for us to make a determination. Other than Huddlestun's older unit, no possible correlatives to the Chandler Bridge Formation have been reported in the Atlantic Coastal Plain.

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# NEW MEMBERS OF THE MADISON LIMESTONE (DEVONIAN AND MISSISSIPPIAN), NORTH-CENTRAL WYOMING AND SOUTHERN MONTANA

By William J. Sando

## INTRODUCTION

The stratigraphy of the Madison Limestone in the mountain ranges of north-central Wyoming and southern Montana was studied by the author during the summers of 1964, 1965, 1966, 1968, and 1970. Three formal members and three informal members of the Madison were described in a series of reports on the stratigraphy of the area (Sandberg and Klapper, 1967; Sando, 1967, 1968, 1972, 1974, 1975, 1976, 1977, 1979). The purpose of this report is to propose formal nomenclature for the three previously informal members of the Madison.

Complete sequences of the members of the Madison are recognized in the Beartooth Mountains, Absaroka Range, Washakie Range, Gros Ventre Range, Wind River Range, Owl Creek Mountains, and Bighorn Mountains (fig. 28). Some of the members extend to outcrop areas to the north and southeast of the area of this study. Preliminary examination of well logs suggests that the members are recognizable also in the subsurface of the Bighorn, Wind River, and Powder River Basins.

Stratigraphic sections measured in the Bighorn Mountains provide complete sequences of the Madison that show variations in thickness and lithology characteristic of its members. The best reference section for the Madison in north-central Wyoming and southern Montana is the well-exposed and readily accessible Little Tongue River section described by Sando (1976, p. 48-52) along U.S. Highway 14 above the Little Tongue River west of Dayton, Wyo. (fig. 28). This section also serves as the type section for the three new members proposed here.

## STRATIGRAPHY

Previously recognized members of the Madison Limestone in north-central Wyoming and southern Montana are, in ascending order, the Cottonwood Canyon Member (Sandberg and Klapper, 1967), the lower dolomite member (Sando, 1972), the Woodhurst Member (Weed, 1899a, b), the cherty dolomite member (Sando, 1972), the cliffy limestone member (Sando, 1972), and the Bull Ridge Member (Sando, 1968). The revised nomenclature is compared with previous nomenclature on figure 29.

### Little Bighorn Member (new name)

The lower dolomite member of Sando (1972) is here formally named Little Bighorn Member for exposures in Little Bighorn Canyon in NW 1/4 sec. 30, T. 58 N., R. 89 W., Sheridan County, Wyo. (see Sando, 1976, pl. 1, for graphic section). The type section is the Little Tongue River section of Sando (1976, p. 48-52) in NE1/4 sec. 27, T. 57 N., R. 87 W.,

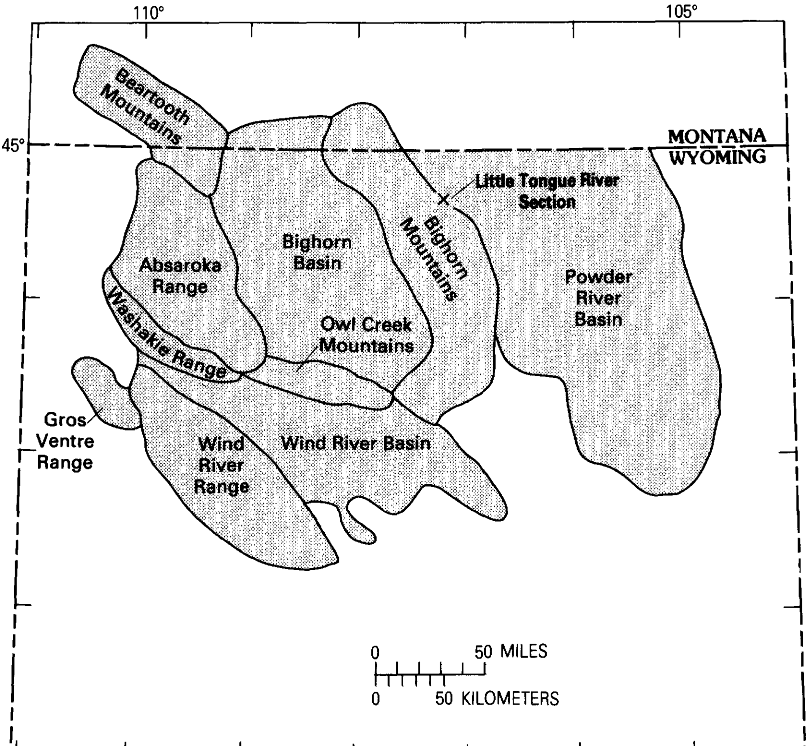


FIGURE 28.—Mountain ranges and basins of north-central Wyoming and southern Montana where new members of the Madison Limestone are recognized and location of type section.

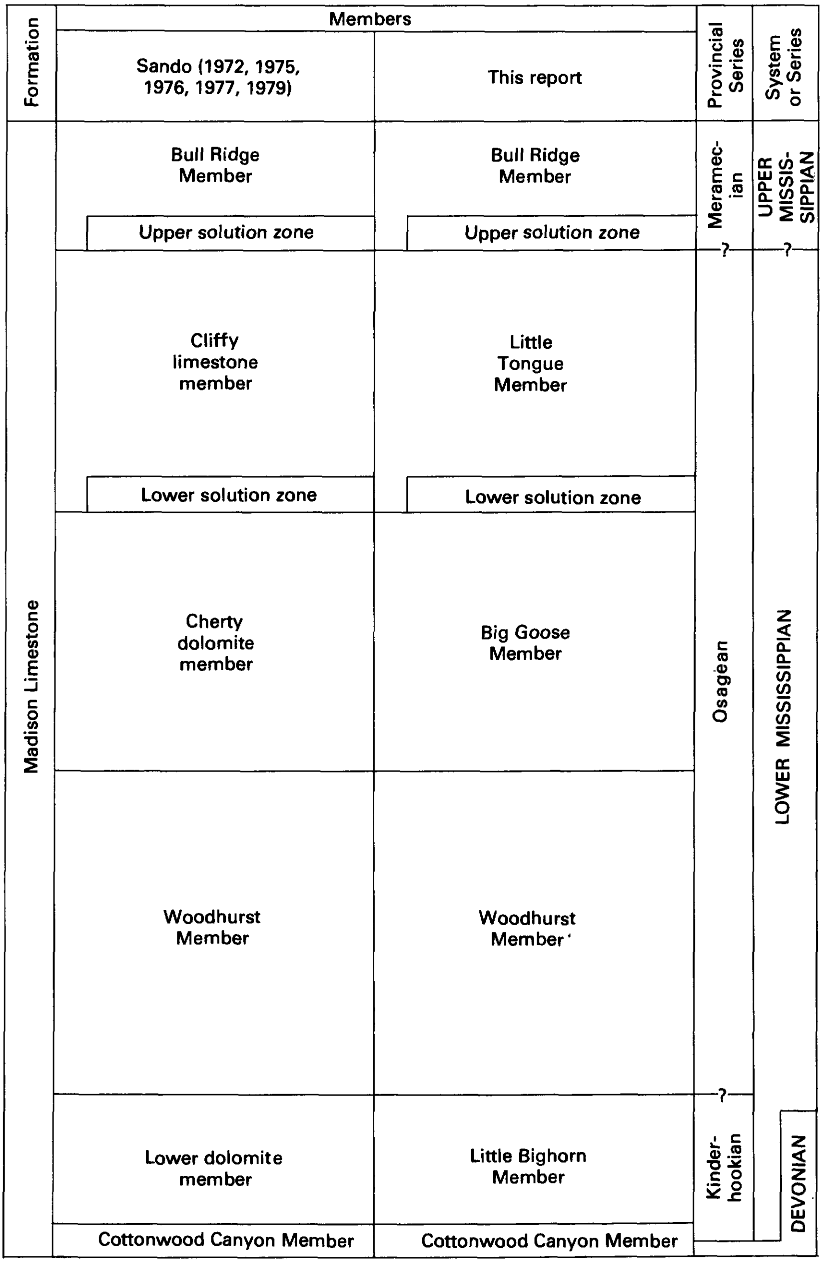


FIGURE 29. – Stratigraphic diagram comparing new nomenclature of Madison Limestone in north-central Wyoming and southern Montana with previous nomenclature and showing ages of the members.

Sheridan County, Wyo. The member includes units 5 and 6 of Sando's (1976, p. 51) section description and is 17.5 m thick in the type section.

The Little Bighorn Member rests conformably on the Cottonwood Canyon Member at most localities but is regionally unconformable on the Jefferson Dolomite (Devonian), Bighorn Dolomite (Ordovician), or Gallatin Limestone (Cambrian and Ordovician) where the Cottonwood Canyon Member is absent. The Little Bighorn Member is overlain conformably by the Woodhurst Member of the Madison. The member consists of 3 to 48 m of thick-bedded, fine- to medium-crystalline, crinoidal dolomite and dolomitic limestone. Conodonts recovered at a few localities date the member as Kinderhookian in age. Brachiopods and corals also occur rarely in the member.

### Big Goose Member (new name)

The cherty dolomite member of Sando (1972) is here formally named Big Goose Member for exposures on Big Goose Creek in NW 1/4 sec. 2, T. 54 N., R. 86 W., Sheridan County, Wyo. (see Sando, 1976, pl. 1, for graphic section). The type section is the Little Tongue River section of Sando (1976, p. 48-52) in NE 1/4 sec. 27, T. 56 N., R. 87 W., Sheridan County, Wyo. The member includes units 19 through 24 of Sando's (1976, p. 50) section description and is 55.5 m thick in the type section.

The Big Goose Member rests conformably on the Woodhurst Member and is overlain conformably by the Little Tongue Member. The Big Goose consists of 38 to 100 m of predominantly fine-grained, thin- to medium-bedded, very cherty dolomite and dolomitic limestone. The member contains rare limestone beds at some localities. Autobrecciation and shattering of the rock are distinctive features. The few corals and brachiopods found in the member support an Osagean age assignment determined mainly on the occurrence of the member between members dated as Osagean on the basis of foraminifers.

### Little Tongue Member (new name)

The cliffy limestone member of Sando (1972) is here formally named Little Tongue Member for exposure at Little Tongue River in NW 1/4 sec. 26, T. 56 N., R. 87 W., Sheridan County, Wyo. (see Sando, 1976, pl. 1, for graphic section). The type section is the Little Tongue River section of Sando (1976, p. 48-52) at the same locality. The member includes units 25 through 29 of Sando's (1976, p. 50) section description and is 66.9 m thick in the type section.

The Little Tongue Member rests conformably on the Big Goose Member and is overlain either conformably by the Bull Ridge Member or disconformably by the Amsden Formation where the Bull Ridge Member was removed by post-Madison, pre-Amsden erosion. The Little Tongue Member consists mostly of cliff-forming medium- to thick-bedded, cherty, crinoidal limestone and dolomitic limestone, but at some localities the member is predominantly or entirely crinoidal dolomitic limestone and dolomite. At most localities, the base of the member is marked by a solution breccia (lower solution zone of Sando, 1972) 2 to 13 m thick that represents a leached interval of evaporite, carbonate, and terrigenous rocks; anhydrite occurs at this level at some subsurface localities. The limestone beds immediately above the solution zone are ordinarily brecciated because of foundering of the roof of the solution zone, forming a collapse zone 5 to 24 m thick. The Big Goose Member ranges in thickness from 24 to 87 m. Large spiriferoid brachiopods and corals are common to abundant in the limestone beds. Foraminifers of Mamet Zone 9 are common and indicate a late Osagean age for the member; Zone 10 foraminifers also are present at one locality.

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# HUNTERSVILLE CHERT (DEVONIAN) EXTENDING FROM SOUTHWESTERN VIRGINIA INTO SOUTHWESTERN NEW YORK, AND ITS BOBS RIDGE SANDSTONE MEMBER

By E. G. A. Weed

The Huntersville Chert is a discrete mappable lithologic unit, extensive in outcrop and in the subsurface (fig. 30), and mappable at 1:24,000. The name was introduced by Price (1929) of the West Virginia Geological Survey and has been used in their published reports during the past 50 years. This paper recommends that Huntersville Chert be accepted as a valid geologic name for use by the U.S. Geological Survey.

The name Huntersville was first applied by Price (1929, p. 236-239) to a yellow to dark-gray sandy chert, which contains a sparse marine fauna of Early Devonian age. Price (1929, p. 397-398) also described an almost white, fine-grained sandstone unit commonly iron stained in outcrop at the top of the Huntersville.

The Huntersville Chert crops out along the Allegheny Front in parts of Virginia and West Virginia. The type locality is in the vicinity of Huntersville in southeastern Pocahontas County, W. Va. (Price, 1929). Its geographic extent in the subsurface is from southwestern Virginia, northward through West Virginia, Kentucky, Ohio, Maryland, and Pennsylvania, to extreme southwestern New York State (Oliver and others, 1971). It ranges in thickness from 5 to more than 60 m in northern West Virginia. An arbitrary limit at less than 50 percent chert is used to define the separation of Huntersville Chert from its neighboring units. The Huntersville Chert is best developed in northeastern West Virginia and southern Pennsylvania. It overlies the Oriskany Sandstone and underlies the Marcellus Shale. Where present, the Tioga Ash Bed lies at or near the base of the Marcellus Shale and above the Huntersville Chert.

Woodward (1943) described the gradational relation between the Huntersville Chert and the Needmore Shale to the east.

Dennison (1961) placed the Huntersville in the Onesquethawan Stage (Lower and Middle Devonian) in Virginia and West Virginia. It intertongues with the Onondaga Limestone and the Needmore Shale to the northeast and east. It intergrades with the Columbus Limestone to the west and northwest in Ohio, and along the south shore of Lake Erie the Huntersville merges into the cherty limestone of the Bois Blanc Formation (Dennison, 1960).

Dennison (1961, p. 25-33) presented a detailed description of the subsurface Huntersville Chert. All gradations between chert and shale exist in the eastern part of the Huntersville. In the central part of the Huntersville, the chert is nearly free of carbonate. To the west, the chert becomes more calcareous and dolomitic. The chert has a higher clastic content in the south.

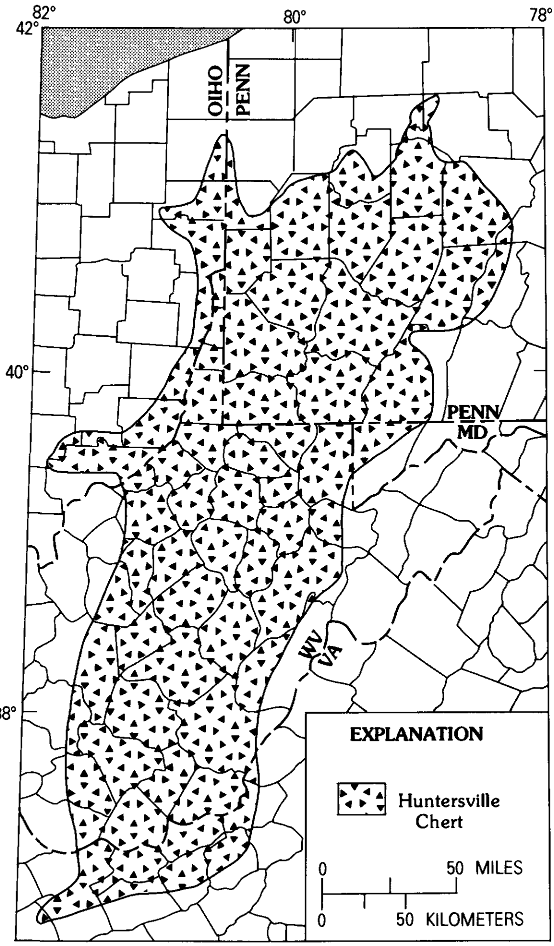


FIGURE 30.—Main subsurface extent of the Huntersville Chert, southwestern Virginia to southwestern Pennsylvania (modified from Oliver and others, 1971).

Dennison also named Price's sandstone member in the upper part of the Huntersville Chert, calling it the Bobs Ridge Sandstone Member of the Huntersville Chert (1961, p. 33-35). Dennison's Bobs Ridge is also herein accepted as a valid geologic name for U.S. Geological Survey usage in West Virginia and Virginia. Its type locality is Bobs Ridge in Greenbrier County, W. Va., and it crops out along Browns Mountain anticline. The member is fine-grained quartz sandstone with much "glauconite." It is 2 to 3 m in thickness at the type locality and underlies the Tioga Ash Bed of the Marcellus except where a tongue of the Needmore Shale is interposed.

Oliver and others (1967, p. 1018-1019) discussed the relative position of the Huntersville in the Onesquethawan Stage and showed its extent in the subsurface and its relation to the Bois Blanc Formation, the Columbus Limestone, the Needmore Shale, and the Onondaga Limestone (Oliver and others, 1969).

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# NEALMONT LIMESTONE (MIDDLE ORDOVICIAN) EXTENDING FROM SOUTHWESTERN VIRGINIA INTO SOUTHWESTERN NEW YORK IN THE CENTRAL APPALACHIANS

By E. G. A. Weed

The Nealmont Limestone is a discrete lithologic unit, mappable at 1:24,000, and extensive in outcrop and in the subsurface in Pennsylvania, West Virginia, and Virginia. Kay introduced the Nealmont in an abstract in 1941 and formally defined it in 1944. The name has been used in many published reports during the past 40 years. This paper recommends that Nealmont Limestone be accepted as a valid geologic name for use by the U.S. Geological Survey.

The name Nealmont was first applied by Kay (1941) to a limestone in the Trenton Group of Middle Ordovician age, including as members the type Rodman, Lemont, and Centre Hall. It ranges in thickness from 20 to 40 m. Kay (1943) included the Nealmont in his discussion of limestone resources in central Pennsylvania.

The Nealmont type section at Union Furnace, Huntingdon County, Pa., is 40 m in thickness (Kay, 1944). It is underlain by rocks of the Black River Group or Limestone and overlain by Kay's Salona Limestone. In descending order, Kay's (1944) Nealmont includes the Rodman, Centre Hall, and Oak Hall Members. Kay described the Rodman Member as dark impure fossiliferous limestone, the Centre Hall Member as gray shaly limestone, and the Oak Hall Member as gray fossiliferous limestone of medium texture with interbedded coquinal ledges. Perry (1964) identified the Nealmont in the Ray Sponaugle well, Pendleton County, W. Va., and described it (1972) as about 85 m of predominantly thin wavy-bedded fine-grained limestone with clay partings overlain and underlain by beds similar to those in the type area. Kay's Oak Hall Member, the basal member, included two beds of altered volcanic ash (Kay, 1956). Equivalent ash beds are present in the Nealmont in West Virginia (Kay, 1956; Perry, 1964, 1972). Knowles (1966) described the Nealmont at Ashcom, Pa., as a dark-gray platy thick-bedded (up to 15 cm) limestone containing many bioclastic beds.

Some of the volcanic ash beds, also termed K-bentonite or metabentonite, present in the Middle Ordovician (Kay, 1944; Craig, 1949; Perry, 1964; Wagner, 1966) have been shown to be correlative throughout the Appalachian basin, especially the two lower beds of Kay's Oak Hall Member.

Cooper and Cooper (1946) and Pierce (1966) discuss the occurrence of the partly equivalent Edinburg Formation and Chambersburg Limestone in Virginia and Pennsylvania.

The Nealmont, striking northeast-southwest, crops out in the western anticlines of the Appalachians in Pennsylvania, West Virginia, and Virginia (Kay, 1944; Cardwell and others, 1968; Perry, 1972). It merges eastward into the Mercersburg Formation of Kay (1944) and Craig (1949) and extends southwestward into West Virginia and Virginia (Kay, 1956), where it becomes increasingly argillaceous and grades into the reddish Moccasin Formation in

Bath County (Bick, 1962). The Nealmont is recognized in outcrops in central Pennsylvania east of the Allegheny Front. Wagner (1966, pl. 7) shows it extending westward in the subsurface into the Pittsburgh-Huntington basin. Northwestward, the Nealmont grades into the Verulam Formation in Ontario (Wagner, 1966). To the north, in New York, the upper part of the Nealmont appears to be equivalent to the Kirkfield of Ontario (Wagner, 1966).

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# RECOGNITION AND FORMALIZATION OF THE PLIOCENE "GOOSE CREEK PHASE" IN THE CHARLESTON, SOUTH CAROLINA, AREA

By Robert E. Weems, Earl M. Lemon, Jr., Lucy McCartan, Laurel M. Bybell, and Albert E. Sanders[1]

## ABSTRACT

Although never formally introduced into U.S. Geological Survey usage, Earle Sloan's (1908) "Goose Creek phase" of the Charleston, S.C., area has proven to be a valid and properly defined stratigraphic unit. As such, it is revived herein and formally recognized as the Goose Creek Limestone. The age of this unit is definitely Pliocene, but its position within that epoch has been controversial. On paleontologic grounds, it is probably slightly older than the Bear Bluff Formation at Myrtle Beach, S.C. Because both the Goose Creek Limestone and the Bear Bluff Formation are soft, medium-grained calcarenites from which aragonite has been removed and (or) recrystallized, lithic stratigraphers have suggested direct correlation of the two units. However, because these units are not known to occur in superposition at any locality, resolution of their physical stratigraphic relations is impossible at this time.

## INTRODUCTION

Sloan (1908) named the "Goose Creek phase" for a section along Goose Creek north of Charleston, S.C. Although "phase" is not a modern stratigraphic term, it is clear that Sloan's "phases" were intended to represent transgressive-regressive episodes of deposition that left behind lithologically distinctive deposits across parts of the South Carolina coastal plain. As such, this concept conforms closely to the modern concept of stratigraphic classification and nomenclature used in the coastal plain. In some cases, Sloan's "phases" are ambiguous or actually composite entities, but, in the case of the "Goose Creek phase," a type section was designated that contains a single, fairly homogeneous, lithologically distinctive deposit that appears to represent a single depositional episode. Much of the aragonitic material that was once present is leached out, but calcareous and phosphatic fossils have persisted and indicate this unit is of Pliocene age.

## STRATIGRAPHY

Sloan's (1908) "Goose Creek phase" is herein revived, formalized, and renamed the Goose Creek Limestone. The unit's type section in a bluff along Goose Creek 0.3 km east of the Seaboard Coastline railroad bridge over Goose Creek in the North Charleston 7 1/2-min quadrangle (fig. 31) remains unchanged. This section, described in Holmes (1860, p. iv) and in Sloan (1908, p. 296, locality 441), is summarized in figure 32. The part of the section below sea level was determined from a core (field number CNC-20-D), which is stored at the U.S. Geological Survey (Reston). Although the type section is only 3 m thick, the unit is known to range up to 8.5 m thick in

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U.S. Geol. Survey Bull. 1529-H, 1982, p. H137-H148

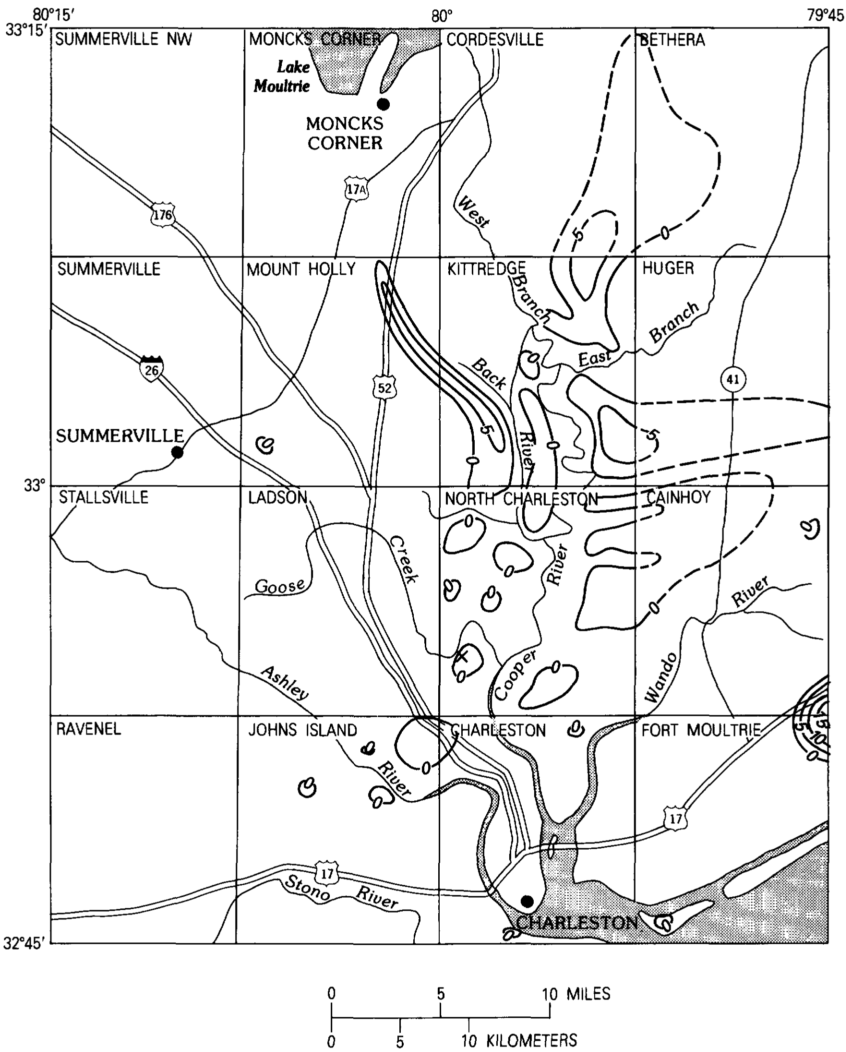


FIGURE 31.–Location of type section (X) and isopach map of the Goose Creek Limestone in the Charleston, S.C., area. Isopach contour interval 5 m. Distribution of subcrop based on nearly 800 auger holes and a few scattered outcrops. 7½-min quadrangles shown by grid and names.

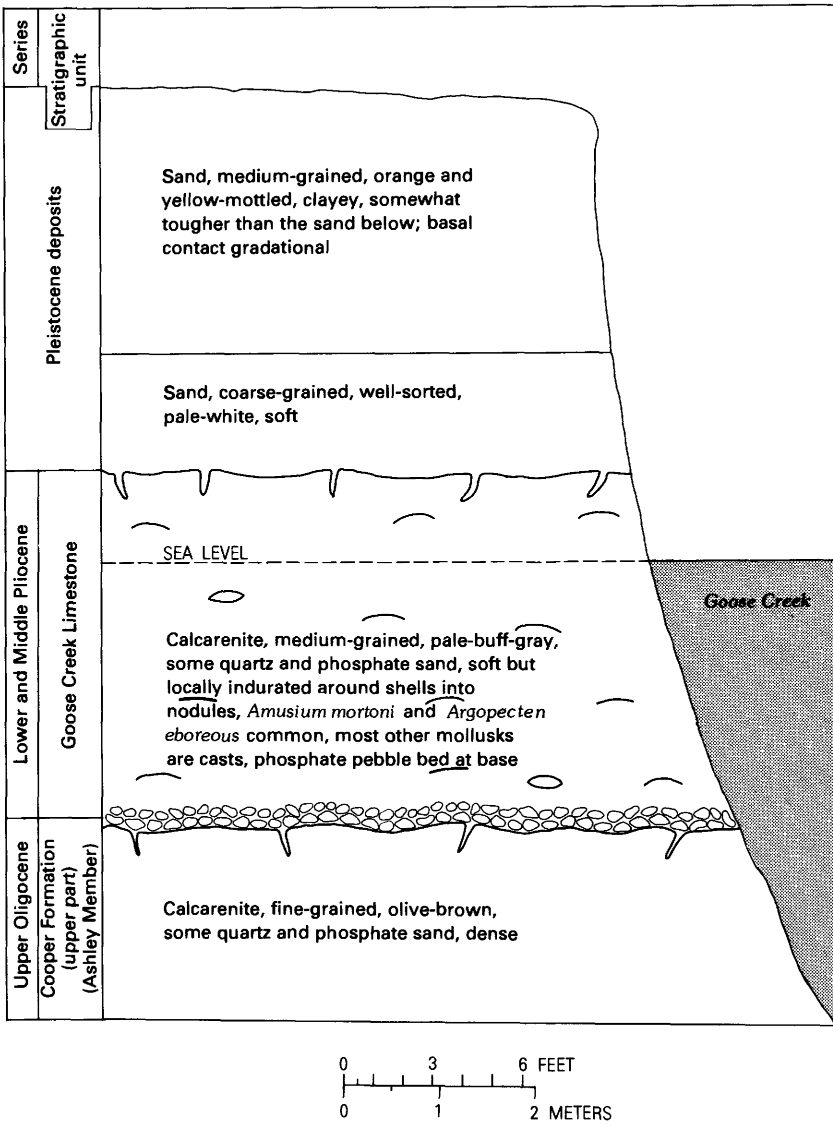


FIGURE 32.–Type section of the Goose Creek Limestone, Charleston, S.C., area. Location shown on figure 31. No vertical exaggeration.

the Kittredge quadrangle, up to 11.8 m thick in a channel in the Mount Holly quadrangle, and up to 18 m thick in the Fort Moultrie quadrangle (fig. 31).

In the Charleston area, the Goose Creek Limestone (table 4) is a medium- to coarse-grained, quartzose and phosphatic, sparsely shelly, pale-buff-gray (wet) to chalk white (dry) calcarenite. Shell material was once abundant, but aragonitic forms have been extensively leached, leaving only molds and casts. It contains kaolinite in the greater than 2 micron-size fraction and both kaolinite and illite-smectite in the less than 2 micron-size fraction. Locally, its composition is influenced by the nature of the substrate. For example, near Mount Holly, a channel containing Goose Creek Limestone is incised deeply into the Eocene Parkers Ferry Member of the Cooper Formation (Ward and others, 1979); there, the Goose Creek is nearly a calcilutite, like the Parkers Ferry, and contains calcilutite clasts obviously derived from the subjacent and adjacent beds of the Parkers Ferry. Where the Goose Creek Limestone lies on the Oligocene Ashley Member of the Cooper Formation, which is a compact, quartzose and phosphatic, fine-grained calcarenite, it is similar to that lithology, though coarser grained, much less dense, and light-buff-gray rather than olive-brown. Locally, in the Charleston area, the distribution of the Goose Creek is well known (fig. 31). Except for five localities, however, it is known only from auger holes.

The fauna of the Goose Creek Limestone is not well known. Aragonitic mollusk shells were once abundant but generally have been dissolved, leaving only molds and casts. Typically, only calcitic mollusk shells have survived in their original form. These include *Amusium mortoni*, *Argopecten eboreus*, *Pecten hemicyclica*, *Ostrea raveneliana*, *O. sculpturata*, and *Crassostrea*? sp. (L. W. Ward and B. W. Blackwelder, written commun., 1979). *A. mortoni* is consistently present among recently collected Charleston Museum specimens from the Goose Creek Limestone at several localities near Charleston. Mollusks from Tuomey and Holmes' (1857) localities on the Cooper River, Edisto River, and Goose Creek were all representative of the Goose Creek Limestone (see Sloan, 1908, p. 285, locality 436) and are the only specimens from the fauna of this formation that have been figured in the literature. Tuomey (1848, p. 179) and Malde (1959, p. 31-32) are the only other publications known certainly to have described a part of this fauna.

There are no published reports of Pliocene vertebrate fossils that can be assigned readily to the Goose Creek Limestone, but two recent discoveries demonstrate that vertebrate remains do occur in place in this formation. In 1974, Sanders recovered a partial skeleton of a baleen whale from Goose Creek deposits in a pipeline excavation at the South Carolina Electric and Gas Company plant on the west bank of the Cooper River in Berkeley County, S.C. (USGS North Charleston 7 1/2-min quadrangle) (fig. 31), approximately 20 km north of Charleston. Preliminary studies by Dr. Frank C. Whitmore, Jr. (U.S. Geological Survey), and by Sanders suggest that the remains are those of a Minke Whale, *Balaenoptera acutorostrata* (Balaenopteridae, Cetacea), a form occurring today in most of the oceans and seas of the world but mainly in temperate and colder waters (Lowery, 1974). A modern record of this species on the coast of South Carolina has been reported by Sanders (1978). During the summer of 1980, James Malcom, a local high school student, collected well-preserved elements of a

Table 4. Sediment characteristics of Pliocene units in the Charleston, S.C., area.
[Numbers refer to localities in figure 34]

| | Sand, silt, and clay (bulk) (percent) | | | | | Sand (HCl insoluble) (percent) | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Total sand | Very coarse- to coarse | Medium | Fine- to very fine | Silt and clay | Total sand | Very coarse- to coarse | Medium | Fine- to very fine | Ratio of insoluble sand/total sand |
| Raysor Formation | | | | | | | | | | |
| 1) Raysor Bridge | 86 | 14 | 62 | 10 | 14 | 63 | 5 | 79 | 16 | 0.73 |
| 2) Cross quarry | 83 | 11 | 64 | 8 | 17 | 60 | 2 | 85 | 13 | 0.72 |
| 3) Canadys Bridge | 55 | 4 | 30 | 21 | 45 | 41 | 5 | 55 | 40 | 0.75 |
| Goose Creek Limestone | | | | | | | | | | |
| 4a) Givhan's Ferry 1 | 76 | 11 | 50 | 15 | 24 | 23 | 1 | 66 | 33 | 0.30 |
| 4b) Givhan's Ferry 2 | 76 | 14 | 57 | 5 | 24 | 25 | 1 | 67 | 32 | 0.33 |
| 5) K-53 | 78 | 41 | 36 | 1 | 22 | 12 | 19 | 74 | 6 | 0.15 |
| 6) Goose Creek | 72 | 16 | 47 | 9 | 28 | 10 | 1 | 93 | 7 | 0.14 |
| 7) Clouter Creek | 79 | [1]43 | 35 | 1 | 21 | 9 | 5 | 87 | 8 | 0.11 |
| 8) Mark Clark pit | 78 | 11 | 63 | 4 | 22 | 11 | 1 | 94 | 6 | 0.14 |
| 9) Tall Pines | 80 | [2]46 | 31 | 3 | 20 | 27 | 47 | 46 | 7 | 0.34 |

[1] Extensive secondary cement, original modal size, probably medium grained.

[2] Deposit about 30 cm thick, coarseness and high quartz content probably due to this being a basal lag deposit with abundant phosphate pebbles and (or) bioturbation churning in quartz sand from Pleistocene bed above.

bird skeleton from Goose Creek Limestone sediments immediately above a basal lag deposit of massive phosphate nodules exposed in a ditch paralleling the Seaboard Coast Line Railway tracks at a position approximately 1.6 km west of S.C. Route 165 (Dorchester Road) in Charleston County (USGS Johns Island 7 1/2-min quadrangle) (fig. 31). Now in the collection of the Charleston Museum, the specimen is under study by Dr. Storrs Olson of the National Museum of Natural History, who has determined it as a gannet, Morus sp. (Sulidae, Pelecaniformes). "True gannets (Morus) are of temperate range in the North Atlantic or belong to somewhat similar, relatively cool-current regions in southern Africa, southern Australia, and New Zealand" (Murphy, 1936, p. 827). The gannet M. bassanus is a fairly common visitor to the South Carolina coast from early October to mid-June (Forsythe, 1978).

Teeth of the shark Carcharodon carcharias occur in the Goose Creek Limestone, but those of the late Miocene form Carcharodon megalodon are not known to be present except as reworked material in basal lag deposits. These circumstances indicate that the Goose Creek is no older than early Pliocene in age (Leriche, 1942). However, calcareous nannofossils suggest that the Goose Creek is probably no younger than the Reticulofenestra pseudoumbilica zone (NN15), which is considered to be middle Pliocene because of the presence of Reticulofenestra pseudoumbilica (Gartner, 1967) and Sphenolithus abies Deflandre, 1954. Together these fossil data imply that the Goose Creek Limestone is early to middle Pliocene (between 5 m.y. and 3 m.y.) in age.

The Goose Creek Limestone occurs extensively along the axis of the modern Cooper River and may fill a tectonic trough and (or) an old river valley (fig. 31). In the Mount Holly quadrangle, however, the channel-like nature of the base of the Goose Creek is more obvious (fig. 33).

## STATUS OF THE RAYSOR FORMATION

Cooke (1936 p. 116, locality 216) named the "Raysor marl" on the basis of a locality first listed by Sloan (1908, p. 280, locality 366) as being present at Raysor Bridge on the Edisto River. The only extant locality that seems to closely match Sloan's description, however, is a bluff located at what is presently called Canadys Bridge. This bluff, badly overgrown and accessible only by drilling, was considered by Blackwelder and Ward (1979) to be the original type locality of Cooke's Raysor. The material cored at this locality (table 4) is siltier than the Raysor Bridge material (collected by Sloan) available at the Charleston Museum but is otherwise comparable. It does not appear that Cooke actually saw the type locality; instead, he relied entirely on the collections of shells made by Sloan that are now stored in the Charleston Museum and at the U.S. National Museum. On the basis of mollusk identifications probably made by either Julia Gardner or W. C. Mansfield, Cooke at first thought this locality represented a unit distinctly older than the "Duplin marl" of North Carolina. At that time, Cooke equated the "Duplin" on faunal grounds to most of the Sloan's (1908) "Goose Creek marl or phase," though two localities were considered to possibly correlate with either the "Duplin" or the Waccamaw. In later years, however, Cooke (1945) decided there was no significant age difference between his "Raysor marl" and the "Goose Creek phase" and, therefore, equated them both with the "Duplin." This status was later accepted by Malde (1959).

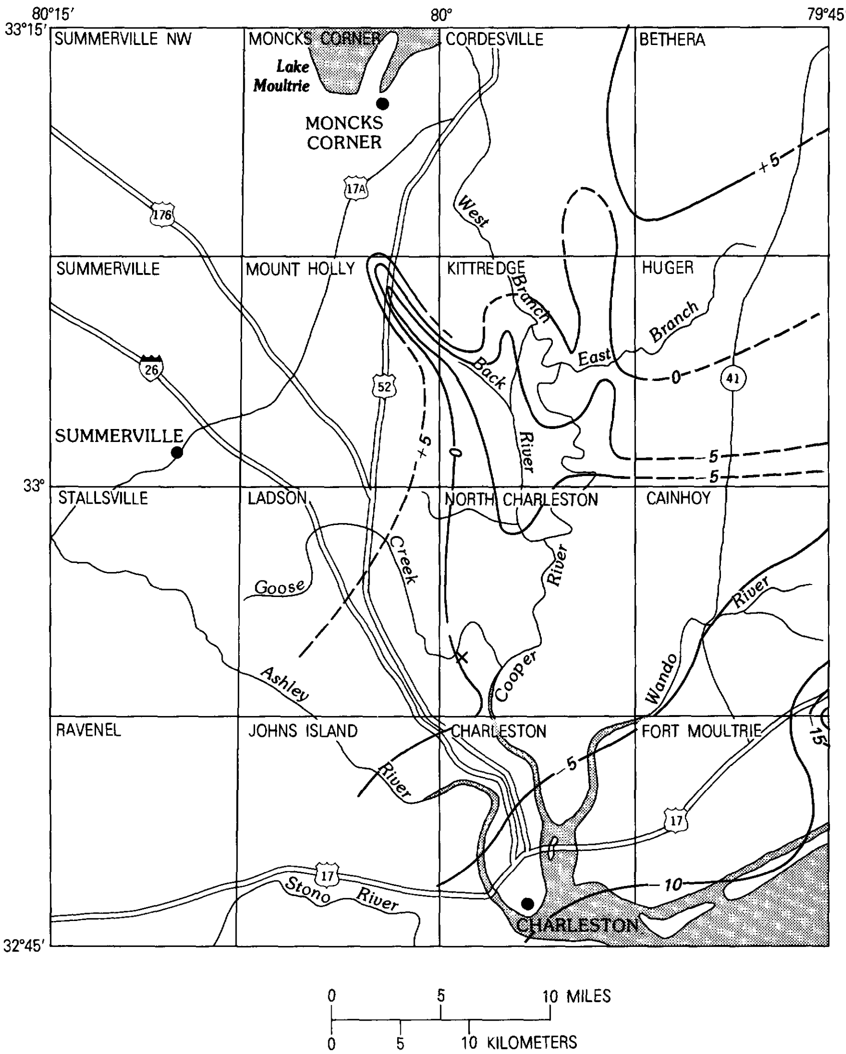


FIGURE 33.–Structure countour map of the base of the Goose Creek Limestone in the Charleston, S.C., area. Contour interval is 5 m; 0 is sea level.

Recently Blackwelder and Ward (1979) have reinstated the "Raysor marl" as the Raysor Formation. They did so in the belief that the Raysor was a shell bed with a largely calcareous matrix as opposed to the "Duplin," which they considered synonymous with the Yorktown Formation, that has a largely clastic matrix. They seem to have derived this apparent difference from an analysis of the "Raysor marl" made by Sloan (1908), who stated that the marl was 80.82 percent calcium carbonate. But collection of matrix from several shells in the Sloan collection from Raysor Bridge, now housed at the Charleston Museum, has shown that this matrix material actually is 63 percent medium-grained quartz sand (table 4) excluding clastic silt and clay (fig. 34). A sample from Canadys Bridge, though more silty, still contains a comparable relative amount of sand (right column, table 4). Since Sloan stated the marl contained 80.82 percent $CaCO_3$ and the marl consisted of "shells in a dark blue matrix," we now conclude in retrospect that Sloan analyzed a very shell-rich bulk sample and not the matrix alone. This would seem to be confirmed by Sloan's referral of the Raysor Bridge locality to his "upper Pee Dee phase" (Sloan, 1908, p. 474), which he characterized as consisting of "a profusion of shells in a dark blue soft clay matrix which encloses variable amounts of sand [italics ours]." Therefore, in bulk sample the Raysor is a calcarenite as reported by Blackwelder and Ward, but the matrix alone is not.

Blackwelder and Ward (1979) suggested that the stratotype of the Raysor Formation be moved to Givhan's Ferry, a locality that neither Cooke, originally, nor Sloan believed to be equivalent to the "marl" at Raysor Bridge. The dissimilarity of the lithologies at these two sites is confirmed by our own studies, which show that the matrix of the pale-buff-gray Givhan's Ferry calcarenite contains about 25 percent quartz sand and only minor kaolinite and aragonite. This analysis matches well with the Goose Creek Limestone (to which Sloan, 1908, referred this locality) and is in marked contrast to the Raysor Bridge locality that yielded dark-blue (fresh) to gray-green (dry) sediment containing 63 percent quartz sand with about twice as much kaolinite, more aragonite, and more hematite than the Goose Creek. The material from the Raysor Bridge locality, however, does match the material correlated with it at the Martin Marietta Company Berkeley quarry (Ward and others, 1979, p. 6), although the matrix of this latter material is not so calcareous as they had indicated it to be. Articulated shells are filled with very fine grained calcarenite, perhaps because their cavities were not widely open to the bottom currents. Discussion with L. W. Ward indicates that such shells were the source of the matrix that Ward and Blackwelder described, but this material is atypical of the matrix collected in bulk quantity. Generally, the yellowish-gray-green shell bed at the Berkeley quarry locality also has abundant kaolinite, hematite, and aragonite and has a shell-rich matrix that contains 60 percent medium-grained quartz sand. Thus, on the basis of bulk matrix, the Berkeley quarry, Canadys Bridge, and original Raysor Bridge localities appear to be lithologically equivalent, but the proposed neostratotype for the Raysor Formation at Givhan's Ferry has a quite different lithology that is readily referred to the Goose Creek Limestone as Sloan originally suggested. Therefore, since the neostratotype proposed by Blackwelder and Ward (1979) would both totally alter the concept of the Raysor as originally established by Cooke (1936) and directly conflict with the concept of the "Goose Creek phase" (herein proposed as the Goose Creek Limestone) as established originally by Sloan (1908), the neostratotype of the Raysor Formation should be abandoned. The

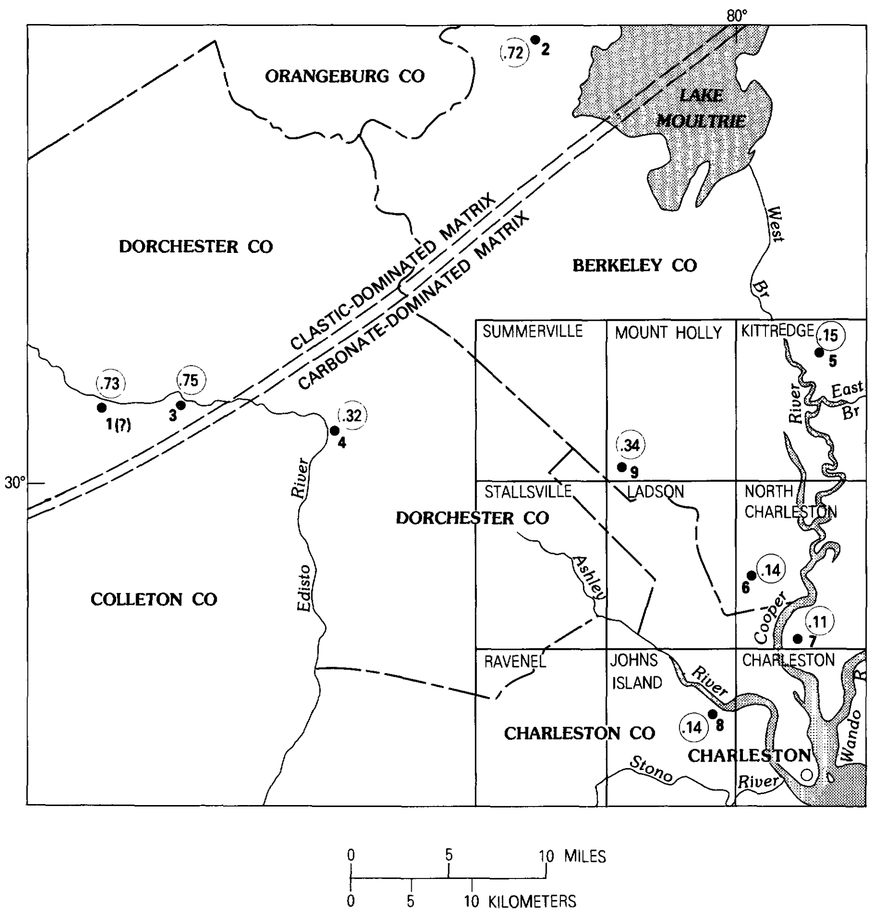


FIGURE 34.–Known distribution of Pliocene localities in the Charleston, S.C., and inner coastal plain area and representative values for quartz sand fraction to total sand fraction (see table 4). Uncircled number indicates location name from table 4; circled number indicates ratio of insoluble sand to total sand.

term Raysor Formation should be restricted to lower Pliocene biocalcarenites with a quartzose matrix, whereas the term Goose Creek Limestone should be restricted to lower to middle Pliocene biocalcarenites with a calcareous matrix.

## STRATIGRAPHIC RELATIONSHIPS OF THE GOOSE CREEK LIMESTONE TO THE RAYSOR FORMATION

The physical stratigraphic relations between the Raysor Formation and the Goose Creek Limestone are presently unclear. The relative sparsity of quartz, kaolinite, hematite, and aragonite in the Goose Creek Limestone at Givhan's Ferry, as compared to the Raysor Formation at Raysor Bridge, Canadys Bridge, and Cross quarry, is striking. This difference is much greater than any seen within the known outcrop range of either unit (fig. 34). It is plausible that these two units represent onshore and offshore facies, respectively, of a single depositional cycle that graded very rapidly from a nearshore clastic phase to a shelf carbonate phase. For example, rapid gradation from a nearshore terrigenous phase to a shelf carbonate phase over a distance of less than 20 km has been documented for the Holocene sediments off the coast of Belize (Purdy and others, 1975). Conversely, the two units could represent two separate transgressions within early to middle Pliocene time and may have no gradational interrelation at all. Although a large fauna was collected from Raysor Bridge by Sloan (Blackwelder, 1967), the poor preservation of the Goose Creek fauna hampers meaningful detailed mollusk comparisons between the two. So far, the Raysor has yielded no diagnostic nannofossils. If future drilling furnishes new sections between the known Raysor and Goose Creek outcrop belts, the physical stratigraphic relations may become clear. In the meantime, all we can demonstrate is that the units are lithologically different where they are known. Should future work show the boundary to be gradational, 40 percent quartz in the sand-size matrix is tentatively suggested as the arbitrary lithologic boundary unless future field mapping shows some other parameter that better characterizes the change from the typical quartzose calcarenite (10-15 percent quartz-sand content) of the Goose Creek Limestone to the shelly quartz sand (60-65 percent quartz-sand content) of the Raysor Formation.

## STRATIGRAPHIC RELATIONS OF THE GOOSE CREEK LIMESTONE TO THE BEAR BLUFF FORMATION

DuBar and others (1974) proposed the name Bear Bluff Formation for Pliocene-age calcarenites exposed in the Myrtle Beach, S.C., area, 130 km northeast of the Charleston area. Analyses of samples from this area indicate an age for this unit within the range of the *Discoaster surculus* zone to the *D. brouweri* zone (NN16-18). This would make this unit late Pliocene in age (between 2.9 m.y. and 1.7 m.y. old) and suggests that the Bear Bluff in its type area is younger than the Goose Creek. It is not known if these two units are related as transgressive portions of a single depositional cycle or if they represent discrete transgressive events separated by an unconformity. In either case, they are lithologically very similar. Because physical stratigraphic relations between the Goose Creek Limestone and the Bear Bluff Formation are unclear, because the two areas are widely separate, and because the name Goose Creek is the older name, we favor the use of the name Goose Creek in at least the Charleston area until such time as the relation between the two units is firmly established. Although there is provision in the code of stratigraphic nomenclature (American Commission on Stratigraphic Nomenclature, 1970) for retention of younger, well-established names of stratigraphic units in preference to older, little used names, we suggest in this case that, if the two units do prove to be lithologically continuous, the older name, Goose Creek, should be retained. The Goose Creek was lithologically well defined by Sloan, so its subsequent abandonment based on purely paleontological correlations should not be grounds for considering it an invalid lithologic unit.

If the possible age differences among the Raysor, the Goose Creek, and the Bear Bluff are correct, a threefold chronostratigraphic division of the Pliocene in South Carolina is indicated, as suggested by Campbell and others (1975). In South Carolina, these authors recognized the Raysor and the Bear Bluff but referred to the middle unit as the "Natural Well member of the Duplin," apparently in reference to Natural Well, N.C. Their middle unit may be equivalent to the Goose Creek Limestone, judging from its position relative to the Raysor and the Bear Bluff. Again, the name "Natural Well member" has been applied only recently, and the Goose Creek is by far the older term.

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