Geologic Map of the Fredericksburg 30' x 60' Quadrangle, Virginia and Maryland

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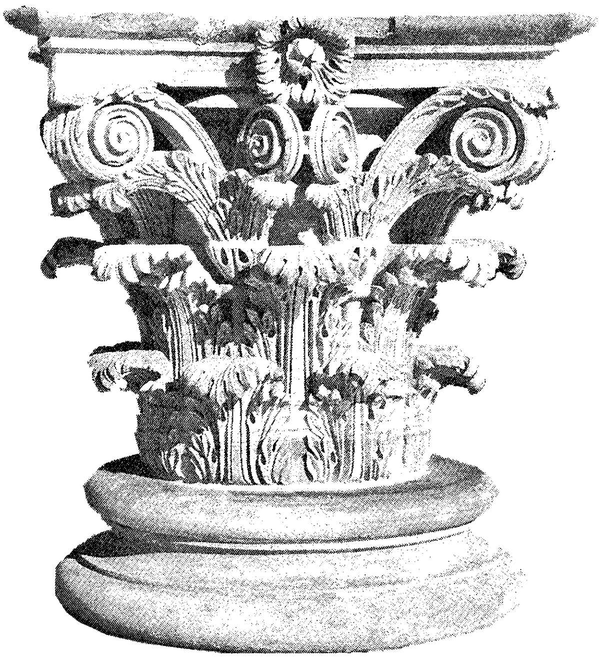

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COVER: This Corinthian capital, originally used atop columns of the east-central portico of the U.S. Capitol building in Washington, D.C., was carved in the early 19th century from sandstone of the Cretaceous Potomac Formation. The stone was obtained within the map area of the Fredericksburg quadrangle from a quarry on Government Island at the head of tidewater on Aquia Creek, Va. The well-defined sedimentary structures (visible as faint, subhorizontal crossbeds in the photograph), the moderately spaced jointing, and the clay matrix made quarrying of the sandstone and carving of the capitals easier. However, these qualities also caused rapid weathering and structural failure, which led to the removal of the columns and capitals from the U.S. Capitol in 1958. This capital was later moved to its present location on the grounds of the U.S. National Arboretum, Washington, D.C. Photograph by J. Stephen Schindler.

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CHAPTER I

OVERVIEW OF THE FREDERICKSBURG 30′ x 60′ QUADRANGLE, VIRGINIA AND MARYLAND

By Robert B. Mixon

INTRODUCTION

The Fredericksburg 30′ x 60′ minute quadrangle, comprising approximately 4,884 square kilometers (1,343 square miles) of northeastern Virginia and southern Maryland, is about halfway between the Washington, D.C., and Richmond, Va., metropolitan areas. The map area consists mostly of small- to medium-sized farms and suburban homes interspersed with large tracts of deciduous woodland; thus, it is dominantly rural in character. The cities of Fredericksburg at the head of tidewater on the Rappahannock River and Culpeper at the northwestern corner of the map area are the principal centers of commerce. The region's main north-south transportation corridor east of the Blue Ridge Mountains, consisting of Interstate Highway 95, U.S. Highway 1, and the Richmond, Fredericksburg, and Potomac Railroad, passes through the city of Fredericksburg—almost bisecting the map area. Extensive Federal land holdings include the Fort A.P. Hill Military Reservation in the southeastern part of the map area, the U.S. Naval Surface Weapons Center at Dahlgren, Va., the Blossom Point Proving Grounds in Charles County, Maryland, and parts of the Quantico Marine Corps Development and Education Command in the vicinity of Garrisonville, Va. The North Anna Nuclear Powerplant is on Lake Anna in the southwestern part of the map area.

The area's economy is based in large part on farming, lumbering, light manufacturing, and the construction and service industries. Important natural resources include crushed stone and extensive sand and gravel deposits that are mined for fill for roads, highways, building sites, earth-fill dams, and other construction purposes. Diatomaceous silty clay in the Calvert Formation is mined in nearby areas for expandable clay and is a potentially valuable commodity in the Fredericksburg map area. However, the most valuable resource is the large volume of fresh water obtainable from various Coastal Plain aquifers (Meng and Harsh, 1988). Of these aquifers, the most important for large industrial users and municipalities are the sands in the Lower and Upper(?) Cretaceous Potomac Formation. In some areas, sands in the Aquia Formation are also an important source of water. Potential mineral resources include lead and zinc ores that were formerly mined near and north of Mineral, Va., along the outcrop belt of the Chopawamsic Formation (Pavlides and others, 1982a). Gold was also mined and prospected in a belt extending from Mineral northeastward to Wilderness in easternmost Orange County, Virginia (Lonsdale, 1927; Sweet and Trimble, 1983).

GEOLOGIC AND GEOMORPHIC SETTING

The Fredericksburg map area encompasses parts of four very different terrains: the Atlantic Coastal Plain, the Appalachian Piedmont, the early Mesozoic Culpeper basin, and the Blue Ridge anticlinorium. In the eastern part of the map area, the early Mesozoic Taylorsville basin is present beneath a thick cover of Coastal Plain deposits.

Atlantic Coastal Plain.—In the map area, the Coastal Plain is a broad, dissected upland that slopes from the Piedmont "highlands" (Hack, 1982) to the east and southeast toward the Chesapeake Bay. The deeply incised estuaries of the Potomac and Rappahannock Rivers and the wide alluvial valley of the Mattaponi River separate the Coastal Plain upland into long, narrow, southeast-trending "necks" or "peninsulas"—terms dating back to Colonial times when ships were the most important means of transportation. The higher, relatively less dissected interfluve areas between the major river drainages range in altitude from 85 m (280 ft) at the western edge of the Coastal Plain to about 45 m (150 ft) east of the Broad Rock scarp in the southeastern corner of the map area. These remnants of a formerly much more extensive, gently seaward-dipping upland surface are believed to closely approximate the original depositional surfaces of fluvial-deltaic sand and gravel sheets of late Pliocene age that commonly cap inner Coastal Plain sedimentary sections.

In three dimensions, the Coastal Plain deposits form a wedge-shaped body of mostly unlithified sand, gravel, silt, and clay that thickens toward the deeper parts of the Salisbury basin in southern Maryland, Delaware, and the Virginia Eastern Shore (Spangler and Petersen, 1950; Murray, 1961; Brown and others, 1972). Sediment thickness ranges from about 600 m (2000 ft) near Oak Grove and Maple Grove in the easternmost part of the map area to a feather edge along the western margin of the Coastal Plain where Cretaceous and Tertiary strata lap onto crystalline rocks of the Piedmont (see cross sections *A–A′* and *D–D′*). The lower part of the Coastal Plain sequence consists of fine to coarse fluvial-deltaic deposits of the Potomac Formation of Early and Late(?) Cretaceous age. These dominantly continental strata are overlain with great unconformity by relatively fine-grained marine deposits, in part shelly, glauconitic, and diatomaceous, that compose the Tertiary formations of the upper part of the Coastal Plain sequence. The surficial sandy gravels and gravelly sands that cap the Coastal Plain upland mark a return in the late Pliocene to dominantly continental sedimentation. The latest Pliocene and Quaternary history of the area is recorded mainly by a step-like series of ancient terraces and intervening scarps that parallel the present-day Rappahannock, Potomac, and Mattaponi Rivers. Sand, silt, and gravel of fluvial and estuarine origin associated with each set of terraces reflect deposition during interglacial highstands of the sea and downcutting during glacially influenced periods of low sea level.

Appalachian Piedmont.—In the central part of the map area, the Piedmont is a north- to northeast-trending highland of crystalline rocks lying between the more easily eroded sediments of the Coastal Plain to the east and the Culpeper basin to the west. This crystalline terrane consists mainly of metamorphosed sedimentary, volcanic, and plutonic rocks of Proterozoic and Paleozoic age. These rocks are highly deformed and exhibit a strong northeast-southwest structural grain. Fresh, unweathered rock, exposed in the deeper stream valleys and, locally, on the steeper slopes, is hard and resistant to erosion. In contrast, the gentler slopes and ridge tops are mantled by soft decomposed rock (saprolite), as much as 18 m (60 ft) thick, derived by prolonged weathering of the underlying crystalline rock. The surface of the crystalline rocks is irregular but dips generally eastward beneath the onlapping Coastal Plain strata and westward beneath the sedimentary rocks of the early Mesozoic Culpeper basin. In the map area, the altitude of the outcropping crystalline rock surface ranges from sea level at the Fall Line on the Rappahannock River at Fredericksburg to about 160 m (525 ft) near Rhoadesville and Paytes in the western part of the Piedmont.

Blue Ridge anticlinorium.—The Blue Ridge anticlinorium of northern Virginia is a broad, northeast-trending, structurally complex metamorphic terrane that has undergone some uplift relative to the Piedmont adjacent to the east and the Valley and Ridge rocks adjacent to the west (King, 1951, 1959). The Blue Ridge terrane consists of stratified metasedimentary rocks and metabasalts of early Paleozoic and Late Proterozoic age and an underlying gneissic and granitic basement-rock complex of Middle to Late Proterozoic age (Virginia Division of Mineral Resources, 1993). The Blue Ridge rocks are thought to represent the continental margin of Laurentia (ancestral North America).

Rocks of the east limb of the Blue Ridge anticlinorium (continental-margin terrane) crop out in several small areas in the northwestern part of the Fredericksburg map area. The larger body of these rocks underlies Clark Mountain (altitude 300 m (984 ft)) and associated ridges in the vicinity of Everona and True Blue, Va.; this rock terrane, consisting of the Catoctin and True Blue Formations, extends from the western edge of the map area northeastward for about 16 km (10 mi) between the Culpeper basin on the west and the mélange terrane (Mine Run Complex) of the Piedmont on the east. Other areas of outcrop of Blue Ridge rocks include a small, north-trending horst of metamorphic rocks (units Zc and Ꞓm) forming "The Ridge" near Stevensburg, Va., well within the Culpeper basin, and the relatively upthrown fault block (units Zch and Zc) adjacent on the west of the Culpeper basin border fault at Culpeper, Va.

In the map area, the continental-margin terrane of the Blue Ridge and the mélange terrane of the Piedmont are separated by the Mountain Run fault zone, a major zone of thrusting with a dextral strike-slip component of movement (Pavlides, 1989). Thrusting along this structure may have begun during the Ordovician Taconic orogeny when the Mine Run Complex, interpreted to be deposits of an early Paleozoic back-arc basin, was thrust westward onto the continental margin.

Culpeper basin.—In the northwestern part of the map area, the crystalline basement rock is overlain with great angularity by thick continental deposits of the Culpeper basin, a long, relatively narrow half-graben that trends northeastward through parts of the northern Virginia and southern Maryland Piedmont. The Culpeper basin is only one of several deep, fault-bounded basins that formed along the eastern part of the Appalachians as a result of the extensional stress field prevailing during continental rifting in the early Mesozoic. In the map area, the geomorphic expression of the basin is the 15-km (9.5-mi)-wide lowland between Culpeper and Lignum, Va. In this area, the basin fill is of Triassic age and consists mainly of red and gray shale, siltstone, mudstone, sandstone, and conglomerate of fluvial and lacustrine origin. Thin to thick sills and dikes of diabase have intruded these sediments, forming aureoles of hornfels and other thermally metamorphosed rocks.

Taylorsville basin.—From its outcrop area at the edge of the Piedmont in Hanover and Caroline Counties, Virginia (Weems, 1980), the early Mesozoic Taylorsville basin extends northeastward beneath the Coastal Plain deposits for about 152 km (95 mi) to the vicinity of Sandy Point, Md., on the northern Chesapeake Bay. In Virginia, the basin is as much as 48 km (30 mi) wide (Milici and others, 1991) and appears to be bordered along or near its western margin by one or more extensional faults.

In recent years, the buried part of the Taylorsville basin has been an area of considerable interest for oil and gas exploration. To that end, the petroleum industry has made geophysical surveys across the Taylorsville basin and has drilled six deep coreholes and three oil and gas tests into the basin in and near the map area. Texaco's W.B. Wilkens No. 1 oil and gas test, drilled into the thicker basinal deposits near Maple Grove, Va., penetrated 625 m (2,050 ft) of Cenozoic and upper Mesozoic Coastal Plain strata and about 2,500 m (8,200 ft) of Triassic and, possibly, Jurassic rocks (Milici and others, 1991). Although several oil and (or) gas shows have been reported in the area, commercial quantities of hydrocarbons do not appear to have been found as yet.

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CHAPTER II

GEOLOGY OF THE PIEDMONT AND BLUE RIDGE PROVINCES

By Louis Pavlides

INTRODUCTION

Metamorphic and plutonic rocks of Proterozoic and Paleozoic age compose the bedrock of the Piedmont and Blue Ridge provinces in the map area. The Piedmont province, as used herein, includes the crystalline rock terrane between the Coastal Plain sediments to the east and the Culpeper basin to the west (see geologic map and fig. 1). The Blue Ridge province principally lies west of the Mountain Run fault zone. In the map area, many of the lithologic and tectonic features of these two provinces, including the ages of their rocks, have been described elsewhere (Pavlides, 1976, 1980, 1981, 1986, 1987, 1989, 1990; Pavlides and others, 1974, 1982a,b, 1994; Bobyarchick and others, 1981; Sutter and others, 1985; Flohr and Pavlides, 1986; Wier, 1977; and Wier and Pavlides, 1985). Because the framework geology is contained in these references, this report will emphasize the tectonic aspects of the geology.

TECTONOSTRATIGRAPHIC TERRANES

The lithostratigraphic units of the Blue Ridge and Piedmont provinces are grouped into six tectonostratigraphic terranes as follows: continental margin, back-arc basin, island arc, successor basins, Salem Church allochthon, and Matta nappe. These terranes contain metasedimentary and meta-igneous rocks of Proterozoic through Carboniferous age and have undergone episodic deformation.

Continental-Margin Terrane

In the northwesternmost part of the map area, the metabasalt and phyllitic tuff of the Late Proterozoic Catoctin Formation (Zc) and the overlying slate, siltstone, and argillite of the lower Paleozoic True Blue Formation (O€t) are thought to compose part of the continental margin of ancestral North America. Structurally, these rocks are considered to be part of the eastern limb of the Blue Ridge anticlinorium. The continental-margin terrane is bounded to the east by the Mountain Run fault zone, a major zone of shearing along which terranes to the east accreted by thrusting prior to about 450 Ma. Subsequently, the fault zone underwent strike-slip faulting in pre-Jurassic time and high-angle normal faulting in the late Cenozoic.

Back-Arc-Basin Terrane

The back-arc-basin terrane of the map area, which is juxtaposed against the continental-margin terrane along the Mountain Run fault zone, consists of two types of mélange deposits (Pavlides, 1989). Both types of mélange have a metasedimentary matrix in which sparse to abundant metamorphosed rock fragments and exotic blocks of varied composition are embedded. The most widespread type of mélange is the block-in-phyllite (small to large rock fragments enclosed in a matrix of phyllite or schist) that composes the Mine Run Complex (O€ml–IV). A less common type of mélange is the block-in-quartzofeldspathic matrix which consists of rock fragments in a muddy to sandy matrix of graywacke composition. This mélange type constitutes the metadiamictite of the Lunga Reservoir Formation (O€l).The mélanges are interpreted as having formed in a back-arc basin between the continental margin to the west and an island arc to the east.

The block-in-phyllite mélange of the Mine Run Complex is considered to be an olistostromal-tectonic mélange that occurs in four imbricated thrust slices (numbered I through IV from east to west). Each thrust slice has its own characteristic exotic-block assemblage. The two easternmost thrust slices (mélange zones I (O€mI) and II (O€mII)) contain blocks of metavolcanic and granitoid rocks that are petrographically similar to rocks within the Chopawamsic Formation (€c) to the east. The exotic blocks of these mélanges are interpreted as fragments shed from the metamorphosed and deformed Chopawamsic Formation as it was thrust onto the accumulating sediments of the Mine Run Complex. Mélange zone III (O€mIII) contains metamorphosed mafic and ultramafic blocks, interpreted as fragments of the back-arc basin ocean floor, that were derived from various sources such as ultramafic protrusions or talus rubble along steep submarine scarps (Pavlides, 1989). Mélange zone IV (O€mIV) lacks exotic material in the map area but contains mafic and ultramafic blocks along strike to the southwest.

Metadiamictite mélange occurs at the north edge of the map area in the Lunga Reservoir Formation (O€l). In this report, diamictite is used as a nongenetic term as defined by Flint and others (1960a,b), and the modifier mélange is added to designate the type of diamictites under discussion. Mélange of this type commonly is characterized by a nonstratified, micaceous, quartzofeldspathic matrix (metagraywacke) in which clasts of metasedimentary and meta-igneous rocks, ranging in size from granule to boulder and larger sizes, are embedded. Foliation in the metadiamictite generally is not apparent in fresh rock but is visible in weathered outcrop. Some of the schist and gneissic clasts in the metadiamictite possess an internal foliation that is athwart that of the northeast-trending regional foliation of the matrix (Pavlides, 1976, p. 31). The source of these clasts was a metamorphic terrane (Pavlides, 1989). One of the characterizing features of metadiamictite mélange is the abundance of granules, pebbles, and, less commonly, boulders of quartz. Many of these quartz clasts are aggregates of granoblastic quartz that are referred to locally as quartz lumps. The quartzofeldspathic groundmass, quartz lumps, and schist fragments are considered to be the sedimentary milieu of the diamictite-type mélanges, which were derived from a common source area during discrete sedimentation episodes. This conglomeratic sediment is the matrix for small to large, exotically derived blocks (see Pavlides, 1989, fig. 9B).

Within the map area, mafic and ultramafic clasts in the Lunga Reservoir Formation are sparse. Only a few pebble- to cobble-sized clasts of amphibolite and a boulder of greenstone have been found.

Island-Arc Terrane

The Central Virginia Volcanic-Plutonic Belt, which is composed of the Chopawamsic Formation (Ꞓc) and Ta River Metamorphic Suite (Ꞓt) (Pavlides, 1981), and the metavolcanic James Run Formation of Maryland compose an allochthonous island-arc terrane (Pavlides, 1989). The Virginia and the Maryland metavolcanic belts are considered to be coeval, at least in part, and are interpreted to have formed an island-arc chain offshore of North America in Cambrian[1] time. The island-arc rocks extend northeastward into Delaware, where the metavolcanic rocks of the Wilmington Complex of Ward (1959) are considered to be similar to rocks in Maryland (Southwick, 1969) that are assigned to the James Run Formation (Higgins, 1972). The metavolcanic rocks of the island-arc terrane are characterized chemically as either island-arc tholeiites or calc-alkaline or both and were intruded by Ordovician and (or) Cambrian plagiogranite and trondhjemitic plutons (Pavlides, 1981; Pavlides and others, 1982b; Southwick, 1979; Higgins, 1990).

The Chopawamsic Formation (Ꞓc) is interpreted as the continentward side of the island arc and the Ta River Metamorphic Suite (Ꞓt) as the oceanward equivalent. This interpretation is based on the lithologic, petrographic, and geochemical features of the metavolcanic rocks as well as the intrusive rocks within them, such as the oceanic trondhjemite (OꞒtj) of the Horsepen Run pluton (fig. 1), the plagiogranites (OꞒpg) of the Richland Run pluton (fig. 1), and the volcanogenic stratabound sulfide deposits near Mineral, Va. (Pavlides, 1981, 1989; Pavlides and others, 1982a).

The Garrisonville Mafic Complex (ꞒZg) occurs within the island-arc terrane and is thrust upon the metadiamictite mélange of the Lunga Reservoir Formation (OꞒl). Some of the mafic fragments within the Lunga Reservoir Formation may have been derived from the Garrisonville Mafic Complex.

This complex consists chiefly of massive to foliated amphibolite and hornblendite (fig. 7). The western part of the complex contains metapyroxenite (composed mainly of clinopyroxene and enstatite) and other altered rocks whose protoliths may have been websterite and norite. Talc-amphibole schist is present locally along the western margin of the complex. Amphibolite normally consists of common hornblende that may or may not have blue-green "uralitic" hornblende. In places, large hornblende grains are poikilitic. Actinolitic and cummingtonitic amphibole is present locally. Plagioclase ranging in composition from oligoclase to calcic andesine (flat-stage measurements) is generally slightly to heavily altered to clinozoisite or epidote. Quartz and chlorite are present in variable amounts in some of the amphibolites; magnetite and ilmenite are common accessory minerals. Hornblendite consists essentially of coarse-grained common hornblende in a groundmass of finer grained common hornblende and sparse amounts of quartz or plagioclase or both.

The Garrisonville is intruded by plagiogranite of the Richland Run pluton as well as by plagiogranite dikes. Such dikes are abundant along and near the southern margin of the complex where they form complex networks of intrusion breccia (this report, fig. 7; Pavlides, 1976, fig. 6). The Garrisonville Mafic Complex (ꞒZg) is interpreted as a possible fragment of oceanic floor of the back-arc basin terrane and thus part of the substrate upon which metavolcanic rocks of the Chopawamsic Formation formed in Early Cambrian time (Pavlides, 1989). However, the contact between the Garrisonville and Chopawamsic is not exposed. The age of the Garrisonville Mafic Complex is not well known. It is assigned a Late Proterozoic and (or) Early Cambrian age on the basis that it is pre-Chopawamsic.

Successor Basins Terrane

The metasedimentary Quantico Formation (Oq) and interbedded lenses of quartzite (Oqq) occur within the island-arc terrane. The Quantico's clay- to sand-sized sediments were deposited unconformably on the deformed and eroded island-arc terrane after that terrane collapsed into a successor basin (Pavlides, 1989). The unconformity between the Quantico and its subjacent rocks was excavated within Dale City, Va., northeast of the map area. There, the fossiliferous Quantico Formation of Upper Ordovician age rests unconformably on the Dale City pluton, an intrusion within the Chopawamsic Formation (Pavlides and others, 1980). The Quantico and Chopawamsic Formations are considered to be unconformable northeast of the Accokeek fault (see geologic map and fig. 1). Southwestward from the Accokeek fault, the Long Branch thrust fault is the boundary between the western side of the Quantico Formation and the subjacent Chopawamsic Formation.

Metasedimentary units Ou and Op are preserved in two synclinal areas between the communities of Cropp and Heflin in westernmost Stafford County. These metasedimentary rocks, which unconformably overlie the island-arc terrane (Chopawamsic Formation (Ꞓc)) and the back-arc-basin rocks of mélange zone I (OꞒml), are thought to have been deposited in one or more successor basins after the subjacent terranes collapsed and formed depositional troughs. The mylonitic schist and phyllite of units Op and Opm are difficult to relate to adjacent rock units and are an unassigned terrane on the tectonostratigraphic map.

Salem Church Allochthon

The Salem Church allochthon (fig. 1) lies mostly on the Quantico Formation (Oq) and part of the Ta River Metamorphic Suite (Ꞓt) and is geographically within the island-arc terrane. However, because it is allochthonous, it is not lithotectonically related to the island-arc terrane. The allochthon consists of the Late Proterozoic and (or) Cambrian Holly Corner Gneiss (ꞒZh), the Falls Run Granite Gneiss of the Berea pluton (Sf), and associated thin, discontinuous quartzite lenses similar to those of the Quantico Formation (Oqq), which are found along parts of the allochthon's southern limb. The quartzite lenses may be a distal part of the basal quartzite of the Upper Ordovician Quantico Formation, which was deposited upon the Holly Corner Gneiss prior to its westward transport. The Holly Corner is intruded by the Silurian Berea pluton (Falls Run Granite Gneiss) (Pavlides and others, 1982b). The Holly Corner Gneiss is primarily a well-foliated hornblende- and biotite-rich gneiss with sparse, thin, calc-silicate layers of quartz, andesine, epidote, and diopside. Locally, potassic feldspar porphyroblasts occur in thin zones within the Holly Corner at the contact with the Falls Run. These porphyroblasts are attributed to contact metamorphism accompanied by potassic metasomatism during emplacement of the protolith of the Falls Run Granite Gneiss (Pavlides, 1980).

The Holly Corner Gneiss may be interpreted as part of an island-arc assemblage which, from west to east, consists of the Cambrian Chopawamsic Formation, the Ta River Metamorphic Suite, and the Holly Corner as a possible distal oceanward facies (Pavlides, 1980, p. 7; Pavlides, 1981). Alternatively, the Holly Corner may be an older formation, possibly of Proterozoic age,

[1]The U.S. Geological Survey has accepted the Cambrian-Proterozoic boundary as being at about 570 million years ago in the standard geologic time scale. However, uranium-lead zircon chronology of volcanic rocks interlayered within a sequence of paleontologically well-established Cambrian rocks in Siberia, suggest that the Cambrian-Proterozoic boundary is better placed at about 544 million years ago (Bowring and others, 1993). Thus, tectonic events related to rocks older than 544 million years described in this report may have different temporal settings than those described herein.

which happens to be overlain by the basal quartzite unit of the Quantico Formation. Because of these temporal uncertainties the Holly Corner is now assigned a Late Proterozoic and (or) Cambrian age.

The rocks within the Salem Church allochthon (the Holly Corner Gneiss and the Falls Run Granite Gneiss) are not considered to be autochthonous for several reasons. Firstly, their distribution along the Rappahannock anticlinorium (fig. 1) is restricted to a relatively small area (the Salem Church 7.5-min quadrangle and the southeastern corner of the Storck 7.5-min quadrangle). This contrasts greatly with the widespread distribution of subjacent rocks (the Quantico Formation and the Ta River Metamorphic Suite) along or parallel with the anticlinorium. Secondly, if the Salem Church allochthon rocks actually were autochthonous, then the Silurian Falls Run must have intruded through the older Quantico Formation (Upper Ordovician) and the Chopawamsic Formation (Cambrian) and should be found as plutons, dikes, or plugs in these older rocks. The apparent restriction of the Falls Run Granite Gneiss (Berea pluton) within the Holly Corner Gneiss and its absence in the subjacent formations supports an allochthonous origin for the Holly Corner host rocks and the Falls Run. An additional support for the allochthonous nature of the Falls Run Granite Gneiss is the absence of contact metamorphism within the Quantico Formation where the Quantico and Falls Run are in direct contact at the southeastern end of the Storck quadrangle. The thin basal quartzite (Oqq) of the Quantico Formation just north and south of the Rappahannock River in the Salem Church quadrangle, which is tentatively included in the Salem Church allochthon, is locally closely plicated into small intrastratal folds, generally less than one foot in amplitude. These folds can be seen in the quartzite exposed on the eastern side of Motts Run below the outlets of Motts Run Reservoir and south of State Route 618 (see Salem Church 7.5-min quadrangle). The intrastratal folds in the basal Quantico Formation quartzites at this locality are thought to reflect deformation during emplacement of the allochthon. Elsewhere, the basal quartzites (Oqq) of the Quantico Formation are free of small intrastratal folds.

Matta Nappe

The schist and gneiss (fig. 8) of the Po River Metamorphic Suite (PzZp) and the plutons that intrude it constitute the northwest-verging, recumbent Matta nappe (fig. 1 and cross section *C–C'*) that is thrust onto the island-arc terrane. Along strike to the southwest the Po River Metamorphic Suite merges into the Maidens Gneiss of the granulite-facies Goochland terrane as defined by Farrar (1984). There the Maidens Gneiss, also interpreted as contained in a westward-verging nappe (Glover and others, 1989, fig. 3), is juxtaposed along the Spotsylvania fault against the island-arc terrane or Chopawamsic Formation (as used by Farrar, 1984, fig. 2). The Po River Metamorphic Suite is interpreted to be part of an exotic or suspect microcontinental block, initially transported by transcurrent faulting, that accreted in stages onto the terranes to the west (Pavlides, 1989).

INTRUSIVE ROCKS

Felsic Plutons

Felsic plutons are the most abundant intrusions within the Piedmont. Isotopic ages are available for only a few of them, so the age assignments of many plutons are based on regional relations or petrographic comparisons with isotopically dated rocks. The plutons are discussed in order of decreasing age.

Many foliated gneissic granitoid rocks, including pegmatoids, are found as tabular and nontabular bodies in the Po River Metamorphic Suite (PzZp) of the Matta nappe. The largest intrusion in this terrane is the Potomac Creek pluton (PzZd), which lies mostly in the southwestern corner of the Stafford quadrangle. The pluton consists mainly of gray, fine- to medium-grained, well-foliated, quartz-plagioclase-biotite granodioritic gneiss that contains abundant well-formed, highly lineated accessory epidote, locally sparse garnet, and minor amounts of muscovite. Tabular, smaller granitoid and pegmatoid bodies ranging in width from less than 2.5 cm to ~7.6 m (~1 in to 25 ft) form concordant, sill-like layers within the gneiss. The nontabular, irregularly shaped granitoid bodies generally form masses that may be parts of plugs of various sizes. Locally, thinner granitoid layers about 0.5 to 1.0 cm wide are conformable with the foliation in the gneiss (Pavlides, 1980, fig. 4). The lack of selvages of biotite or hornblende suggests that the granitoid layers are veins.

In general, the granitoid rocks of the Po River are two-mica gneissic monzogranite and granodiorite (PzZpg) (Pavlides, 1980). Monzogranite contains quartz (22–30 percent), plagioclase (31–39 percent), potassic feldspar (20–47 percent), biotite (0.5–11 percent), and muscovite (1–14 percent). Green and brown biotites are commonly intergrown with muscovite; locally, biotite is altered to chlorite. Ubiquitous myrmekite occurs at the margins of plagioclase in contact with potassic feldspar both at rims and as bulbous protrusions into the plagioclase. Some feldspars are poikilitic and some occur as phenocrysts enclosed by a finer grained quartz-feldspar matrix locally characterized by recrystallized, granoblastic-textured quartz or quartz that locally suggests a recrystallized mortar texture. Quartz in some of these rocks occurs in elongate, discontinuous ribbons. Garnet is a rare accessory in subhedral, partial grains that locally are poikilitic.

The metagranodiorites of the Po River Metamorphic Suite generally contain quartz (23–30 percent), poikilitic and nonpoikilitic plagioclase (22–30 percent), potassic feldspar (7–42 percent), biotite (2–9 percent), and muscovite (0–9 percent). Biotite ranges in color from brown to red to reddish brown to green and locally is altered to chlorite. Epidote is generally present with the chlorite. Muscovite is locally intergrown with biotite, and both minerals are dimensionally aligned and impart a strong foliation to the rock. Myrmekite is present, and garnet is a rare constituent. Evidence of late strain and recrystallization is similarly present in many granodiorites.

Within the Po River Metamorphic Suite, granitoid dikes and sills of the Falmouth Intrusive Suite (Cf) have the same general composition as the Po River granitoids. The Po River granitoids, however, contain somewhat less quartz than the Falmouth granitoids. Where suitable exposures are present, the well-foliated Po River granitoids are crosscut by the less foliated to nonfoliated Falmouth granitoids. The age of the granitoid rocks of the Po River Metamorphic Suite is uncertain; they are considered to be Late Proterozoic and (or) early Paleozoic because of their position within the Po River.

The large plagiogranitic Richland Run pluton (O€pg) (fig. 1) extensively intrudes both the Garrisonville Mafic Complex and the Chopawamsic Formation. A sample near the northern end of the Richland Run pluton has yielded a discordant U/Pb age and a $^{207}Pb/^{206}Pb$ age of 459 m.y. (Pavlides, 1976, table 1). This age is based on an analysis of the total population of zircons obtained from the sample. Because of this, as well as its discordant nature, it may not reflect a reasonable age for the plagiogranite. Its age, therefore, is interpreted from the field relations. The Chopawamsic, which is intruded by the plagiogranite, is overlain by the Upper Ordovician Quantico Formation that is not intruded by the plagiogranite. Therefore, a provisional age of Cambrian or Ordovician or both is assigned to the plagiogranitic rocks. For similar reasons the oceanic trondhjemite (Pavlides, 1981) of the Horsepen Run pluton (O€tj), which only intrudes the Chopawamsic, is also considered to be Cambrian or Ordovician, or both, in age. It is considered to be somewhat younger than the plagiogranite because it is generally less deformed and contains relict granophyric texture. Trondhjemite and plagiogranite are present in modern and ancient island arcs and the presence of these rocks within the island-arc terrane of the Piedmont is a consistent association. Thus, the rocks that intrude the Chopawamsic are interpreted to have formed above a subduction zone that, during Cambrian time, dipped westward (present-day coordinates) beneath the ancient island arc. Earlier subduction is interpreted to have formed the volcanic rocks of the continentward Chopawamsic Formation and possibly the oceanward facing, more mafic rocks of the Ta River Metamorphic Suite (Pavlides, 1981).

The metamonzogranite of the Goldvein pluton (Og) originally was dated somewhat equivocally as Ordovician and (or) Cambrian. Four discordant U/Pb ages of zircon from the total population of each sample have been obtained from the pluton (T.W. Stern, written commun., 1983 and 1984). One sample is from within the map area, and three are from the northeastward extension of the pluton (Pavlides, 1990). The $^{207}Pb/^{206}Pb$ age of these four samples is between 525 Ma (Middle Cambrian) and 487 Ma (Early Ordovician). Using zircon separates supplied by Tom Stern, J.N. Aleinikoff (written commun., 1995) obtained a $^{206}Pb/^{238}U$ age of 458 ± 8 Ma, the age now accepted for the Goldvein pluton. This age was obtained by using data from the SHRIMP (sensitive high-resolution ion microprobe) at the Research School of Earth Sciences, Australian National University, Canberra, Australia.

Isotopically dated felsic plutons of Ordovician age and Early Silurian or Late Ordovician age also intrude rocks of the back-arc basin (mélange terrane). The intrusions are the pyroxene- and amphibole-bearing shoshonitic, K-alkalic monzonites (~450 Ma) of the Lahore pluton (Olp and Ola, respectively) and the granodiorite (~440 Ma) of the Ellisville pluton (SOe) (Pavlides and others, 1994). The ages were obtained from a combination of geochronologic techniques including U/Pb (zircon), Rb/Sr (whole-rock isochron), and $^{40}Ar/^{39}Ar$ (age spectra and isochrons of amphibole). Both the shoshonitic monzonites of the Lahore pluton and the granodiorite of the Ellisville pluton could have formed along the continental margin of ancestral North America after accretion of the metamorphosed and deformed back-arc-basin and island-arc terranes along the Mountain Run fault zone. An alternative scenario for the Lahore pluton is that it formed as a subduction-related, distal, continentward intrusion from a shoshonitic magma after accretion. The shoshonitic magma could have been supplied from a mantle wedge above a still active subduction zone which to the east had been involved in magma-generation that formed the island-arc volcanic-plutonic rocks. The Lahore monzonites have an initial $^{87}Sr/^{86}Sr$ ratio of 0.7046 and could have formed from a mantle source whereas the initial $^{87}Sr/^{86}Sr$ ratio of the Ellisville granodiorite is 0.7062 and may have formed at deep crustal levels (Pavlides and others, 1982b, 1994). The small granite (SOg) northeast of Garrisonville apparently intrudes the plagiogranite (O€pg) of the Richland Run pluton. It is not strongly deformed and because it intrudes the plagiogranite it is provisionally considered to be of Silurian or Ordovician age.

Within the Salem Church allochthon the Falls Run Granite Gneiss (Sf) of the Berea pluton forms a sill-like intrusion in the Holly Corner Gneiss. The Falls Run Granite Gneiss has been dated by U/Pb (zircon) and Rb/Sr (whole-rock isochron) methods as being of Silurian age, about 408 Ma. The Berea pluton's strongly deformed, coarse-grained metagranites and metamonzonites have an initial $^{87}Sr/^{86}Sr$ ratio of 0.7070 that suggests an origin in the Earth's crust rather than in the mantle (Pavlides and others, 1982b). The small pluton of Falls Run Granite Gneiss within the Po River Metamorphic Suite in the north-central part of the Spotsylvania 7.5-min quadrangle is included in the Falls Run mainly because of its general petrographic similarity to the rocks of the Berea pluton.

The youngest felsic rocks in the region belong to the Falmouth Intrusive Suite (Cf) (Pavlides and others, 1982b). The Falmouth Suite contains both deformed and relatively undeformed rocks. These rocks also have been dated by the U/Pb (zircon) and Rb/Sr (whole-rock isochron) methods. The monzogranites, granodiorites, and tonalites that constitute the Falmouth Intrusive Suite are divisible on the basis of chemical and petrographic features into three groups (A, B, and C) of which groups A and B have been dated isotopically. Group A has an age of about 322 Ma whereas group B, whose age is not closely confined, is about 309 Ma. Group A has an initial $^{87}Sr/^{86}Sr$ ratio of 0.703, which suggests a lower crust or mantle source, whereas the initial $^{87}Sr/^{86}Sr$ ratio of Group B is 0.7088 which suggests crustal involvement in its magma generation. The isotopic ages of the Falmouth Intrusive Suite were obtained from small plugs and dikes and not directly from the plutons that contain Falmouth Intrusive Suite granitoids (Pavlides and others, 1982b, fig. 1). The interpretation that Falmouth Intrusive Suite granitoids make up these plutons is based primarily on petrographic comparisons (Pavlides, 1980). The Falmouth granitoids intrude the Po River and Ta River Metamorphic Suites, the Holly Corner Gneiss and Falls Run Granite Gneiss of the Salem Church allochthon, and the Quantico Formation. One small Falmouth granitoid plug (not shown on the maps) intrudes the Chopawamsic Formation along the Rappahannock River in the area between the Quantico Formation and the Richland Run pluton. This is the westernmost extent of Falmouth Suite intrusion recognized in the Piedmont. The Falmouth granitoids locally are cut by pegmatoids also assigned to the Falmouth Intrusive Suite.

Mafic Plutons

Two small Cambrian(?) plutons of amphibole metagabbro (€g) intrude the island-arc terrane. One is found in the Chopawamsic Formation in Stafford County just west of Mountain View; this body contains inclusions of Chopawamsic (Pavlides, 1976, fig. 5). The other gabbroic pluton occurs in the Ta River Metamorphic Suite in Spotsylvania County south of Todds Tavern. The mafic pluton (Olm) in southern Orange County is considered to be an altered metapyroxenite intruded by the Late Ordovician Lahore pluton (Pavlides and others, 1994).

Other Intrusions

As discussed earlier, the Garrisonville Mafic Complex is provisionally considered to be a fragment of back-arc ocean floor and, if so, may have been emplaced diapirically when the back-arc basin formed in the Proterozoic or earliest Cambrian. Small basaltic dikes (not shown on the geologic map) intrude the gabbroic pluton emplaced in the Chopawamsic Formation and also the plagiogranite (Pavlides, 1976) of the Richland Run pluton. These basaltic dikes may be diachronous or of the same early Paleozoic age.

The youngest intrusive rocks within the Piedmont terrane are diabase dikes of Early Jurassic age (Jd) like those within the Ellisville and Lahore plutons or those that occur to the north of the Lahore pluton (geologic map).

STRUCTURAL FEATURES

Folds, Foliation, and Lineation

The least deformed metasedimentary rocks are those of the True Blue Formation (O€t). These rocks are cut by slaty cleavage and, generally, form open folds. Locally, the unconformity between these cover rocks and the subjacent Catoctin Formation (Zc) is interpreted to be a décollement (thrust fault) surface. Folding has not been recognized in the Catoctin because of poor exposure and absence of marker beds. In places, the Catoctin is transected by steeply dipping to vertical foliation that is parallel to the trend of the cleavage in the overlying True Blue Formation.

Large-scale folds in the mélange deposits (O€ml–IV) are difficult to detect due to lack of marker beds and inadequate exposure. The matrix rocks of the mélange deposits, however, have a well-developed phyllitic foliation that locally is axial planar to small folds. However, transposition both on a macro- and microscale is also locally present (Pavlides, 1989, fig. 9A), and some of the foliation becomes transposed into an anastomosing phacoidal foliation. Near the Mountain Run fault zone, transposition foliation grades into the more highly sheared rocks of the fault zone. Locally, crenulation cleavage (not shown on geologic map) transects the phyllitic foliation (Pavlides, 1989, fig. 4E).

The phyllitic foliation to the west and north of the Long Branch and Accokeek faults has a general northeast trend. Locally, the foliation is warped around the Ellisville pluton (geologic map and fig. 1). The Ellisville pluton intruded and thermally metamorphosed the surrounding country rock at about 450 Ma (Pavlides and others, 1994). However, the phyllitic foliation of this area and, particularly, that within the Mine Run Complex is of pre-Lahore pluton age (440 Ma).

Judging by the transposition foliation recognized in parts of the Mine Run Complex belt, much of the phyllitic foliation formed tectonically. However, in some areas, foliation that has been folded into map-scale folds originated as bedding-parallel foliation. This type of foliation formed when phyllosilicates, such as mica, which had been deposited as flat flakes along bedding surfaces in fine-grained rocks, later underwent mimetic recrystallization and deformation and, commonly, developed layer-parallel foliation.

The rocks of the island-arc terrane have two distinct structural styles, one exemplified by the stratigraphic folds of the Chopawamsic Formation outcrop belt and the other by the foliation folds of the Ta River outcrop belt. Locally, in the northern part of the map area, the Chopawamsic Formation has clearly defined stratigraphic or bedding folds below the unconformably overlying Quantico Formation. However, to the south where the Long Branch thrust separates the two formations, the structure within the Chopawamsic is more obscure. Here, as a result of close folding and probably transposition in the generally homoclinal sequence, bedding and compositional layering are commonly parallel with foliation.

Structurally, the complexly folded terrane of the Ta River Metamorphic Suite is part of the Rappahannock anticlinorium, a large feature enclosed by the Long Branch and Spotsylvania faults (see geologic map and fig. 1). The anticlinorium extends from the vicinity of Stafford, Va., in the north-central part of the map southward to the James River and beyond (Pavlides, 1989, fig. 1A, B) and includes the Whispering Creek anticline terrane of Espenshade and Potter (1960, pl. 2) near Dillwyn, Va. In this report, the Rappahannock anticlinorium includes parts of the Quantico Formation, the Ta River Metamorphic Suite, and the Salem Church allochthon.

Most folds along the Rappahannock anticlinorium are foliation folds formed by the folding of an earlier schistosity. In foliation folds, bedding and compositional layering may trend essentially parallel to the schistosity (geologic map) and, locally, represent transposed layering of closely folded to nearly isoclinally folded rocks. Although intrafolial folds are rare, some small-scale folds are present in the Holly Corner Gneiss (Pavlides, 1980, fig. 6) of the Salem Church allochthon. In the Quantico Formation, east of the Long Branch and south of the Accokeek faults, however, transposed folds have not been recognized. Probably, the layer-parallel foliation in this part of the Quantico, like some of the phyllitic foliation to the east, may have formed through mimetic crystallization of sedimentary, bedding-aligned phyllosilicates. In this part of the Quantico, which underwent greater deformation and metamorphism than its counterpart to the north (described later in this section), such mimetically formed layer-parallel foliation developed into schistosity.

In the southern part of the Rappahannock anticlinorium, foliation folds are upright, northeast plunging or doubly plunging, such as those in the Ta River Metamorphic Suite (€t) northeast of the Elk Creek pluton and in the Quantico Formation (Oq) northeast of the Northeast Creek pluton (fig. 1 and geologic map). In the northern part of the Rappahannock anticlinorium, foliation folds have been refolded by a late set of folds (discussed later in this section). Along the core of the Rappahannock anticlinorium, the Quantico Formation is contained in a synform that, in the region south of the Salem Church allochthon and north of the Ta River Metamorphic Suite, is overturned to the southwest. The recumbent synform is succeeded to the northwest by the slightly overturned (a few tens of degrees to the northwest) La Roque synform (LRS, see fig. 1), a faulted segment of the Quantico synclinorium. To the southwest, the Quantico is contained mostly within the Quantico synclinorium, which in this area is generally an upright foliation fold.

The Ta River Metamorphic Suite along the core of the Rappahannock anticlinorium also contains refolded folds whose axial surfaces, as in the case of the Quantico Formation to the north, are warped around late folds such as the Mine Run antiform (MRA, see fig. 1). This antiform also refolds the Massaponax synform (MS, see fig. 1) that is cored by a thin inlier of Quantico Formation originally deposited on the Ta River Metamorphic Suite.

The Salem Church allochthon (fig. 1), composed of the Holly Corner Gneiss (€Zh), quartzite lenses of the Quantico Formation (Oqq), and the Falls Run Granite Gneiss (Sf), is interpreted as the lower overturned limb of a recumbent nappe. This interpretation stems from the present structural position of the basal quartzite lenses (Oqq) of the Quantico Formation. The lenses occur structurally beneath the Holly Corner Gneiss in the Salem Church allochthon whereas they probably were deposited originally on top of the Holly Corner. The allochthon is a refolded, doubly plunging and downward-facing synform that is outlined by the Pipe Dam Run synform (PDRS, see fig. 1). This synform is a foliation fold that has been refolded by the Rocky Pen antiform (RPA, see fig. 1). Like the structurally conformable synforms and antiforms within the subjacent Quantico Formation and Ta River Metamorphic Suite, the Pipe Dam Run synform also is considered to have been originally a northeast-trending fold.

The Po River Metamorphic Suite is contained within the northwest-verging, recumbent Matta nappe (see cross section *C–C′*). This nappe was thrust westward onto the island-arc terrane along the Spotsylvania fault. Westward-verging, small-scale recumbent folds at several places within the Matta nappe suggest that the Po River Metamorphic Suite is contained in the upper limb of the nappe. The nappe has been plicated by northeast-trending, generally upright folds, and these folds have a foliation that is generally parallel to compositional layering. Therefore, the Matta nappe is also a large-scale recumbent foliation fold.

The youngest folds in the region refold the earlier foliation folds or schistosity within the Rappahannock anticlinorium and the Matta nappe. Generally, the late folds have rectilinear to slightly curvilinear, northeast-trending axial surfaces. However, in the structurally complex area at the northern end of the Rappahannock anticlinorium, late folds such as the Mine Run and Rocky Pen antiforms have axial surfaces that are curvilinear to sinuous. Very locally, as along the eastern part of the Pipe Dam Run synform, an axial planar foliation is present. However, the latest foliation is not axial planar, generally trending athwart the axial surfaces of foliation folds.

Early mineral lineation within both the Quantico and Ta River is warped around the Mine Run antiform, and the plunge direction of this lineation indicates a clockwise rotation of axial surfaces of folds from an original northeast trend to a southeast trend (see geologic map). A gentle (~15°–20°), northeast-plunging late mineral lineation commonly is associated with the late foliation in this area and may reflect the plunge of the Rappahannock anticlinorium. A similar mineral-lineation trend and plunge occurs within the Potomac Creek pluton in the Matta nappe terrane immediately to the northeast of the Salem Church allochthon. This suggests that the same kinematic process of penetrative deformation has affected both the Rappahannock anticlinorium and the Matta nappe.

Faults

Thrust Faults

The Mountain Run fault zone is one of the most clearly recognizable faults in the region. It occupies a northeast-trending zone of variable width which at different places contains sheared rocks, phyllonites, mylonites, breccias, and phyllites having fish-scale structure. The broad fault zone, which is near the western edge of the map area, is not closely constrained. The eastern boundary is defined by the easternmost occurrence of phyllite having weakly developed fish-scale or button-schist structure. Within the broad zone, phyllite with fish-scale structure is irregularly interspersed with non-deformed phyllite. Two pronounced topographic features within the fault zone close to its northwestern margin are the well-developed, northwest-facing and northeast-trending topographic scarps that are as much as 24 m (79 ft) in height. The Kellys Ford scarp (see fig. 1) occurs on the northeastern side of the Rappahannock River where the river crosses the Mountain Run fault zone (see geologic map). To the southwest, the northeast-trending Mountain Run scarp parallels the northeast-flowing Mountain Run and ends where this creek changes from a northeast course to a more irregular east-northeast course (see geologic map). Locally, along the scarps, gnarled and folded thin quartz veinlets are enclosed within sheared or brecciated rocks of the fault zone. A similar lithology occurs along strike in the westward-facing upland area intervening between the two scarps. At several places near the foot of the Mountain Run scarp, core drilling (U.S. Geological Survey, unpub. data) along the alluvium-floored valley of Mountain Run intersected carbonaceous phyllite with subhorizontal slickenside lineations.

The Mountain Run fault zone is interpreted to have formed initially as a thrust along which the back-arc terrane, and its contiguous island-arc terrane to the east, were accreted onto ancestral North America at the end of the Ordovician (Pavlides, 1989). Subsequently, various other fault movements have occurred along the fault zone (Pavlides, 1994). The horizontal slickensides in the carbonaceous phyllite beneath the valley of Mountain Run suggest strike-slip movement. Small-scale folds in the uplands along and near the scarps suggest dextral strike-slip motion. Undeformed diabase dikes of Early Jurassic age, which cut rocks of the Mountain Run fault zone at several places, indicate that thrusting, strike-slip movement, and associated mylonitization and brecciation within the fault zone occurred in pre-Early Jurassic time. The Kellys Ford and Mountain Run scarps are interpreted as fault-line scarps that formed along normal faults, with the scarp-forming blocks being on the upthrown side. The deep weathering that is characteristic of the Piedmont rocks has resulted in generally subdued topography throughout the province. Therefore, conspicuous scarps such as the Kellys Ford and Mountain Run suggest a relatively young age (late Cenozoic?).

The Chopawamsic thrust fault is interpreted to be a thrust that moved the island-arc terrane westward and northward onto the mélanges of the Mine Run Complex. The fault is marked at a number of places by steeply dipping mylonite. The fault attitude may be related to an original listric nature, or deformation after thrusting, or both. Because of poor exposure, direct evidence for the continuity of the fault is sparse. However, a seismic profile south of the map area along Interstate I–64 shows a thrust fault separating the Chopawamsic Formation from the rocks on its western side (Harris and others, 1982). Similar interpretations of the Chopawamsic thrust along the I–64 seismic traverse have been made by Glover and others (1989, fig. 3) and Pratt and others (1988).

The Mine Run Complex, which lies between the Mountain Run fault zone and the Chopawamsic thrust fault, is interpreted as a collage of four thrust slices containing mélange deposits. The thrusts within the Mine Run Complex are based on sparse, widely separated physical evidence, such as isolated outcrops of mylonite. Some of the postulated thrust slices of mélange, however, are characterized locally by aeromagnetic map patterns and each has a unique assemblage of exotic blocks that distinguishes one thrust slice from another (Pavlides, 1989). Scattered outcrops of mylonite within some of the thrust slices suggest internal faults but poor exposure and lack of marker beds preclude tracing them.

The Long Branch thrust fault, which bounds the Quantico Formation along its northwestern side, merges northeastward into the Accokeek strike-slip (tear?) fault. Fluxion-textured mylonite, developed in the basal quartzite of the Quantico Formation (Pavlides, 1976, fig. 3), has been recognized at several places along the strike of the fault. The linear trace of the fault along the northwestern side of the Quantico synclinorium may have formed as a consequence of late strike-slip movement. Also, the presence of subhorizontal mineral lineation within the basal quartzite of the Quantico along the fault northeast of Mineral is compatible with penetrative deformation associated with late strike-slip movement. The kyanite that defines this mineral lineation also exhibits pull-apart texture compatible with elongation parallel to the Long Branch thrust. Two small thrust slices, thought to be the same general age as the Long Branch thrust (see discussion in the section on Age of Deformation and Metamorphism), occur to the northeast of Mineral, Va. They contain rocks of the Ta River Metamorphic Suite and are in thrust contact with the Chopawamsic Formation but are overlain by the Quantico Formation along the Long Branch thrust.

The Spotsylvania thrust fault, although not exposed, is well documented by geophysical data. The rocks on either side of the fault have markedly different lithologies as well as strongly contrasting aeromagnetic and aeroradiometric signatures (Neuschel, 1970; Pavlides, 1980; Bobyarchick and others, 1981; Wier and Pavlides, 1985). However, rather than a single fault, the Spotsylvania thrust may represent a zone of faulting (Pavlides, 1980). The Spotsylvania thrust extends southwestward for a considerable distance beyond the map area (Pavlides, 1989). Its subsurface extent to the southwest had been recognized originally in the I–64 seismic traverse by Harris and others (1982) and geologically mapped near the James River by Farrar (1984, fig. 1). After the Spotsylvania fault first developed as a thrust, the rocks on its southeastern side are thought to have undergone southwestward strike-slip movement (see discussion in the section on Age of Deformation and Metamorphism). The fault's linear trend to about the latitude of the Salem Church allochthon may be the result, in part, of such strike-slip movement (fig. 1). Additional thrusting along the Spotsylvania fault may have occurred to the northeast of the allochthon. The Fall Hill high-angle reverse fault (Mixon and Newell, 1977, 1982; Wier and Pavlides, 1985) may have formed in the Mesozoic along a northeast-trending splay of the Spotsylvania fault where the fault changes from a northeastern trend to an irregular northern trend (see geologic map and fig. 1).

Other Faults

A number of strike-slip and reverse faults, as well as normal faults, also are present. Northeast-trending strike-slip faults occur along the borders or near the edge of the Rappahannock anticlinorium at several places. Two such faults, having dextral displacement, are on either flank of the anticlinorium a short distance southwest of the Salem Church allochthon. Two additional northeast-trending strike-slip faults, which have sinistral displacements, occur farther to the southwest along the anticlinorium near Margo and McHenry. One fault, to the southwest of Snell, Va., offsets the Spotsylvania thrust fault whereas the other, the Sturgeon Creek fault, displaces the contact between the Ta River Metamorphic Suite and the Quantico Formation and curves to the southwest into the axial region of the Quantico synclinorium. To the north of the Northeast Creek pluton, two faults that offset the Quantico-Ta River contact probably include strike-slip as well as near-vertical, dip-slip movement. Faults that have a similar movement sense displace rocks of the continental-margin terrane at a number of places. Northwest-trending, steeply dipping faults, probably having strike-slip and (or) dip-slip movement, offset the True Blue-Catoctin Formation contact at several places. Farther to the northeast, the contact between Mesozoic rocks and the continental-margin terrane is cut by generally northeast-trending faults. Most of these faults are probably steeply dipping normal faults, but some have a component of strike-slip movement. The northeast-trending fault separating the Mesozoic Culpeper basin terrane from rocks of the Mine Run Complex is probably a normal fault that developed by Mesozoic time along a splay of the Mountain Run fault zone. The Dumfries, Fall Hill, and Hazel Run high-angle reverse faults along the eastern side of the Piedmont (Mixon and Newell, 1977) are described in Chapter III in the section on Coastal Plain structures.

METAMORPHISM

Regional Metamorphism

The continental-margin terrane is characterized by low-grade metamorphism. The metavolcanic rocks of the Catoctin Formation have greenschist-facies mineral assemblages of chlorite, sodic plagioclase, and epidote, whereas the overlying cover rocks are at slate grade. Rocks within the Mountain Run fault zone are generally within the greenschist-facies of regional metamorphism, and at a few places some of the rocks contain two generations of chloritoid (Pavlides, 1987). The mélange deposits of the Mine Run Complex, except where contact metamorphosed, are, for the most part, also at greenschist-facies grade, and the matrix rocks of the mélanges are characterized by chlorite-muscovite or chlorite-muscovite-garnet assemblages. Mélange zone III, in particular, is characterized by late, fine-grained, euhedral magnetite that crosscuts foliation. The exotic blocks of the mélange zones also contain greenschist-facies assemblages; serpentinites contain antigorite and mafic blocks generally contain actinolitic amphibole, epidote, and albite (Pavlides, 1989). However, near the northern end of mélange zone III, an area with retrograde staurolite schist (Pavlides and others, 1994, fig. 21) that locally contains well-formed muscovite and chloritoid is enclosed by greenschist-facies phyllites. This localized area initially underwent prograde metamorphism to staurolite grade followed by retrogression that formed chloritoid and micaceous shimmer aggregate pseudomorphs of staurolite. Recrystallization accompanying a later prograde metamorphic event may have formed the coarse-grained muscovite and well-formed chloritoid that is associated with the shimmer aggregates. The abundant euhedral magnetite that truncates foliation throughout mélange zone III also may have formed during this late regional prograde event. The reason for the staurolite retrogression is not clear but it may have been related to the thrust faulting in the eastern part of this region. With local exceptions, the metavolcanic and metasedimentary rocks of the Chopawamsic Formation are at greenschist-facies grade and the mafic metavolcanic rocks contain albite-chlorite-epidote mineral assemblages.

The Garrisonville Mafic Complex contains mostly amphibolites and hornblendites; in its western part it also contains lesser amounts of enstatite-bearing metapyroxenite as well as metawebsterite and metanorite (Pavlides, 1976). Amphibolite- and pyroxene-bearing rocks within the Garrisonville may not have undergone greenschist-facies metamorphism and thus may retain some of their original mineralogy.

The Quantico Formation, north of the Accokeek fault, is at greenschist-facies grade. Where the Quantico is in unconformable contact with the Chopawamsic Formation, it generally contains white mica, chlorite, garnet (sparse), and locally rare chloritoid. Southward from the Accokeek fault, where the contact between the Chopawamsic and Quantico becomes the Long Branch thrust, the metamorphic rocks of the Quantico are at staurolite and higher grade. North of the Abel Lake reservoir, along Long Branch (see geologic map), the increase in metamorphic grade is first characterized by the appearance of quartz lumps that are mantled by kyanite and staurolite (Pavlides, 1976). To the north and south of Abel Lake, along Long Branch, the Quantico Formation is generally a biotite-muscovite-garnet-staurolite±chloritoid schist. The dimensionally aligned muscovite and biotite define the foliation. Partial to complete retrogression of staurolite to micaceous shimmer aggregates, and of garnet margins to chlorite, is common. The refolded part of the Quantico Formation immediately southeast of the Salem Church allochthon is mineralogically similar to that near Abel Lake along Long Branch, but staurolite and garnet are rarely retrograded here (Pavlides, 1976, fig. 4). Fibrolite, generally enclosed in megacrystic muscovite and in quartz, is present in the eastern part of the Quantico Formation in this region. Also, quartz-kyanite veins and veinlets are sparsely present in this part of the Quantico Formation. Along strike to the southwest, within the Quantico synclinorium, the Quantico Formation is at garnet-staurolite grade. In the vicinity of the Sturgeon Creek strike-slip fault (see fig. 1), kyanite is present as a megacrystic constituent. To the south, thin quartzitic layers along the western edge of the Quantico Formation and northeast of Lake Anna contain chloritoid, staurolite, and kyanite in mutual contact with each other. Southwest of Lake Anna the Quantico Formation contains staurolite and chloritoid megacrysts. Northeast of Mineral, and opposite the small thrust slice of Ta River Metamorphic Suite, is a thin mylonitic quartzite of the Quantico Formation, which is adjacent to the Long Branch thrust. This quartzite has a pronounced subhorizontal lineation defined by dimensionally aligned kyanite that is locally characterized by pull-apart textures, as mentioned earlier.

The Holly Corner Gneiss of the Salem Church allochthon is at amphibolite grade and is basically a hornblende-biotite-plagioclase-epidote quartz gneiss. Hornblende and biotite are aligned along foliation (Pavlides, 1976, fig. 4), which defines the refolded Pipe Dam Run foliation synform. The intrusive Falls Run Granite Gneiss has similar foliation defined by dimensionally aligned biotite and hornblende (Pavlides, 1980, fig. 8).

The Ta River Metamorphic Suite and the Po River Metamorphic Suite are also within the amphibolite facies of regional metamorphism (Pavlides, 1980; Bobyarchick and others, 1981). However the Po River is locally migmatitic and characterized by augen textures (Pavlides, 1980, figs. 2–4). The thrust-fault-bounded slivers of Ta River Metamorphic Suite (Ꞓt) on the western side of the Quantico Formation near Mineral are at amphibolite-facies or higher grade. The northern of the two thrust slivers contains pyroxene as well as amphibole.

Contact Metamorphism

Only the enclosing rocks of the Lahore, Ellisville, and Berea plutons show recognizable contact metamorphism. The contact metamorphosed rocks around the Lahore and Ellisville plutons have been described in Pavlides and others (1994). These plutons have intruded rocks of the Mine Run Complex and Chopawamsic Formation, which are within the greenschist facies of regional metamorphism (Pavlides, 1989). The phyllitic country rocks near the intrusive contacts have been metamorphosed to schists and gneissic rocks. Megacrystic muscovite and biotite are common, especially near the intrusive contact. Fibrolite needles are common in such micas, occurring mostly in aligned swarms that cut across the crystallographic axes of the micas (Pavlides and others, 1994, fig. 24). Such fibrolitic mica is attributed to late-stage potassic metasomatism by fluids emanating from the plutons. Locally, kyanite is present within the innermost part of the thermal aureole. Kyanite and staurolite are present in some of the metapelites as well as in some of the metavolcanic rocks of the Chopawamsic Formation (Pavlides and others, 1982a, fig. 4B). In order to form aluminosilicates in the volcanic rocks of the Chopawamsic Formation, it is assumed that the rocks underwent hydrothermal alteration and were residually enriched in Al_2O_3 and SiO_2 prior to contact metamorphism. Kyanite also occurs with chloritoid on the eastern side of the Ellisville pluton. There, within the Chopawamsic Formation, kyanite is present in fine-grained sulfidic schist composed of the mineral assemblage muscovite-chlorite-pyrite-kyanite-biotite-quartz. This kyanite schist apparently is interlayered with metafelsite and with chloritoid-bearing phyllite and schist.

Staurolite occurs within the contact-metamorphosed schist surrounding the Ellisville pluton. It forms euhedral as well as subhedral crystals. In some of the massive sulfide deposits near Mineral (Pavlides and others, 1994, pl. 2), the staurolite locally contains as much as 7 percent zinc (Cox, 1979; Pavlides and others, 1994).

Chloritoid occurs within the thermal aureole of the Ellisville pluton generally at greater distances from the pluton contact than the fibrolite-bearing schist described above. The chloritoid-bearing rocks commonly are phyllite and semischist and lack the close folding of the polydeformed fibrolitic schists of the innermost part of the contact aureole (Pavlides and others, 1994). Chloritoid in the phyllites and semischists is typically poikiloblastic and encloses fine-grained groundmass quartz grains. Some chloritoid-bearing contact metamorphosed rocks also contain small amounts of well-formed, fine-grained staurolite and subhedral to anhedral garnet. Chloritoid also occurs with kyanite or with kyanite and staurolite. Fine-grained intergranular chlorite is locally abundant in chloritoid phyllites. Brown biotite and subhedral to euhedral poikiloblastic chloritoid also form irregular folia in schistose rocks. In outcrop, such rocks locally have foliation planes containing chloritoid arranged in fascicular aggregates.

A number of minerals occur in the Chopawamsic Formation that may or may not have formed directly from contact metamorphism but which nonetheless are found within the thermal aureole of the Ellisville pluton. Among these are gahnite, margarite, tourmaline, allanite-clinozoisite, and chlorite whose occurrences and associations have been described in Pavlides and others (1994).

The thermal aureole on the southeastern side of the Ellisville pluton contains a variety of contact metamorphic assemblages in an irregular distribution pattern. Among these assemblages are (1) staurolite-biotite-carbonate, (2) biotite-chloritoid-staurolite-kyanite-muscovite, (3) staurolite-biotite-garnet, (4) garnet-kyanite-biotite-staurolite (Pavlides and others, 1982a, fig. 4B), and (5) staurolite-anthophyllite-biotite-chlorite (Pavlides and others, 1982a, fig. 4F).

Some of these mineral assemblages, such as 2, 4, and 5, are unusual because of the four coexisting AFM phases. Commonly during progressive metamorphism, chloritoid breaks down to form staurolite. The presence of staurolite in assemblages 2, 4, and 5 may result from the high zinc content of the rock, which stabilized the staurolite through substitution of zinc for iron in the staurolite structure. The stabilizing effect of zinc in staurolite has been described by Ashworth (1975). Alternatively, it is possible that these unusual assemblages are in that part of the aureole that overlies a near-surface cupola or extension of the Ellisville pluton. Rapid heating during pluton emplacement combined with the sluggish nature of mineral conversions may have allowed metastable associations to form. That a near-surface extension of the Ellisville pluton exists on its eastern and northern sides is suggested by a gravity survey as well as by consideration of the resulting resetting of the K-Ar isotopic system of amphiboles in exotic blocks within the Mine Run Complex on the northern side of the Ellisville pluton (Pavlides and others, 1994).

The contact metamorphic effects of the intrusion of the Berea pluton (see fig. 1) on the Holly Corner Gneiss host rock are limited in nature. One probable effect is the formation of the potassic feldspar porphyroblasts that are present in thin zones within the Holly Corner at the contact with the plutonic rocks (Falls Run Granite Gneiss). The Holly Corner Gneiss also contains similar potassic feldspar megacrysts along the western side of the Salem Church allochthon in the area between the Falls Run Granite Gneiss and the pluton composed of granitoids of the Falmouth Intrusive Suite. Since recognizable contact metamorphic effects by the Falmouth granitoids have not been recognized elsewhere in the region, the potassic metasomatism is ascribed only to the intrusion of the Berea pluton.

AGE OF DEFORMATION AND METAMORPHISM

The Cambrian (about 550 Ma) Chopawamsic Formation of the island-arc terrane underwent deformation and greenschist-facies metamorphism by about Middle Cambrian time at the beginning of the Penobscot orogeny (Pavlides, 1989). In the early phases of the orogeny, westward and northward thrusting of the island-arc terrane was initiated. At this time, blocks of foliated metamorphic rocks were shed from the Chopawamsic Formation into eastern parts of the back-arc basin where sediments of the Mine Run Complex were accumulating. With the initial westward-directed (present-day coordinates) transport during Penobscot deformation, it is possible that thrusts developed within the back-arc basin supplied some of the blocks of mafic and ultramafic rocks found in mélange zone III (Pavlides, 1989). In any event, thrusting during Middle Cambrian to Early Ordovician time is considered to be the initial phase of the telescoping of the back-arc basin. The Penobscot orogeny is thought to be a long-lived episodic event and a precursor to the Taconic orogeny. Deformation and metamorphism within the back-arc-basin terrane was essentially completed when the Lahore pluton (450 Ma) intruded the Mine Run Complex. The Lahore and Ellisville plutons were intruded at a general depth of 13 to 18 km and at a temperature of about 760 °C (Pavlides and others, 1994).

The tectonic process or processes by which the deformation of the Quantico Formation and the structurally conformable superjacent Salem Church allochthon took place are complex. The Quantico Formation has foliation essentially parallel to bedding and the folds within it are foliation folds. The geographic position of the Quantico when it was folded, however, poses problems as to the nature and timing of the deformation process as well as the grade of metamorphism.

The staurolite- and higher grade Quantico Formation bounded to the west and north by the Long Branch and Accokeek faults (see fig. 1) had previously been considered structurally and stratigraphically continuous with the low-grade Quantico Formation north of that latitude (Pavlides, 1989, 1990). It is now believed that the Long Branch and Accokeek faults separate the high- and low-grade metamorphic parts of the Quantico. The general on-strike occurrence of these Quantico segments is considered fortuitous. The Quantico Formation within the Rappahannock anticlinorium is now believed to have been deposited farther east (present-day coordinates) than the Quantico to the north of the Accokeek fault. This interpretation is required to explain the structural relations and metamorphic grade of units in the area that contains the Salem Church allochthon, the immediately enclosing Quantico Formation, the Ta River Metamorphic Suite below the Quantico Formation, and the rocks of the Quantico synclinorium with relation to the Chopawamsic Formation on its western side.

The scenario adopted herein is that the early Paleozoic folding of the Chopawamsic prior to the deposition of the Quantico Formation did not affect the coeval Ta River Metamorphic Suite (Ꞓt) or the Holly Corner Gneiss (ꞒZh), both of which lay successively eastward of the Chopawamsic terrane. The Holly Corner Gneiss is assigned a Cambrian or Proterozoic age because of its uncertain stratigraphic correlation with the Ta River and Chopawamsic rocks. The stratigraphic relation envisioned is that the Quantico Formation overlies the deformed Chopawamsic Formation with angular unconformity but disconformably overlies the protoliths of the Ta River Metamorphic Suite and the Holly Corner Gneiss to the east. Only the basal quartzite of the Quantico (Oqq) was deposited on the Holly Corner Gneiss. At about this time, the Ellisville pluton (440 Ma) had been emplaced and the trench and associated westward-dipping subduction zone had ceased to exist. By about 440 Ma the Po River terrane, composed of the recumbent westward-verging Matta nappe, was being thrust onto the eastern edge of the Holy Corner Gneiss terrane. By 440 to 408 Ma, accretion over continental crust by all metamorphic terranes is suggested by an initial $^{87}Sr/^{86}Sr$ ratio of 0.7062 from granodiorite of the Ellisville pluton and an initial ratio of 0.7070 from the Falls Run Granite Gneiss (Pavlides and others, 1994). At about 410 Ma, the Matta nappe was intruded by an elongate pluton (immediately southeast of the Massaponax synform) interpreted to be Falls Run granite.

Following intrusion of the Falls Run Granite Gneiss at 410 Ma and prior to the emplacement of the Falmouth Intrusive Suite granitoids at 320–300 Ma (Pavlides and others, 1982b), the terrane of the Matta nappe and particularly that of the Rappahannock anticlinorium underwent major additional deformation and metamorphism. In the time interval 410–300 Ma, the generally undeformed Quantico Formation, the subjacent Ta River Metamorphic Suite, the Holly Corner Gneiss, and the sill-like Berea pluton were folded into structurally conformable, essentially upright isoclinal folds having axial-plane foliation parallel to bedding along the limbs of the folds. The folds within the Quantico Formation are preserved within the La Roque synform and the Quantico synclinorium. Also at this time, along the western side of the Rappahannock anticlinorium, the unconformity between the Chopawamsic and the Quantico Formations became a décollement along which the Long Branch thrust developed. The upright isoclinal folding is inferred to be related to westward thrusting of the Matta nappe along the developing Spotsylvania thrust. With continuing deformation, the upright isoclinal folds to the east of the La Roque synform and Quantico synclinorium became westward-verging recumbent isoclinal folds. In the area now containing the Salem Church allochthon, the recumbent folds of the Quantico Formation and the Ta River were depressed into a structural basin. The Holly Corner Gneiss and the intrusive Falls Run Granite Gneiss developed into a recumbent isoclinal nappe, the Salem Church allochthon, that was thrust westward over the structural basin. Thus, locally, the Salem Church has the geometry of a doubly plunging northeast-trending synform, herein referred to as the Pipe Dam Run synform. Following these events, a combination of thrusting and dextral strike-slip faulting having a southwestern vector of motion folded the Pipe Dam Run synform, the northeast-trending folds within the Quantico Formation, and the Ta River Metamorphic Suite, including the Massaponax synform, around late folds such as the Mine Run and Rocky Pen antiforms. Locally, thrusting and dextral strike-slip movement along the Spotsylvania fault tilted the eastern part of the Rappahannock anticlinorium downward so that the overriding Spotsylvania fault block carried a thicker section of the Matta nappe over the anticlinorium. The north- to northeast-trending late folds (Mine Run and Rocky Pen antiforms) have a north to northeast plunge. The late mineral lineation, which plunges gently to the north and northeast, may have formed because of the pervasive deformation associated with the late folding. The lineation also is well developed in the granitoids of the Potomac Creek pluton of the Matta nappe.

The rocks of the Rappahannock anticlinorium and Matta nappe have undergone polymetamorphism. At least two metamorphic events can be recognized by a study of the porphyroblasts within the Quantico Formation. Early porphyroblasts of biotite and staurolite, for example, lie along foliation. Some porphyroblasts contain helicitic textures that, in places, suggest rotation of porphyroblasts during or after their formation. Generally, coarser grained porphyroblasts of staurolite, biotite, and garnet lie athwart the rock foliation and are interpreted as products of a second metamorphic event. The first metamorphism probably occurred when the upright isoclinal folds formed. The later metamorphism probably occurred when the isoclines were refolded to produce interference patterns

An $^{39}Ar/^{40}Ar$ traverse was made along the Rappahannock River across the Salem Church allochthon by Sutter and others (1985) using amphiboles from the Holly Corner Gneiss. From west to east, Carboniferous plateau ages were obtained which ranged from 321 to 307 Ma. This suggests that the closure temperature of amphibole (about 520 °C) was attained at about the time of intrusion of the Falmouth Intrusive Suite granitoids (320–300 Ma) and that the western part of the Holly Corner Gneiss passed through the 520 °C isotherm about 14 m.y. earlier than the eastern end of the Holly Corner Gneiss. This may possibly reflect the deeper burial of the eastern part of the Rappahannock anticlinorium suggested earlier in this section. Also, a thermobarometric study using garnet and coexisting biotite (Flohr and Pavlides, 1986) was made within the Quantico Formation essentially parallel to the $^{39}Ar/^{40}Ar$ traverse. On the basis of studies of staurolite-grade schists to the west and fibrolitic schists to the east, the temperature and pressure ranges respectively from 487 °C and 4.3 kb to the west to 615 °C and 4.8 kb to the east. These data are also consistent with an eastward increase in temperature and pressure because of deeper burial of the eastern part of the Rappahannock anticlinorium.

Farther to the south, in a less constrained west-to-east traverse across the Rappahannock anticlinorium and the Matta nappe (Ta River and Po River Metamorphic Suites), the $^{39}Ar/^{40}Ar$ amphibole ages range between 313 and 228 Ma (Louis Pavlides, M.J. Kunk, and Henry Cortesini, Jr., unpub. data). Also, just northeast of Mineral, Va., a small thrust slice that consists of rocks of the Ta River Metamorphic Suite has amphiboles of Carboniferous age. The amphibole ages indicate that these varied terranes, which include the Salem Church allochthon, the Rappahannock anticlinorium, and the Matta nappe, must have been metamorphosed to amphibolite grade and polydeformed during the Alleghanian orogeny. It is thought that much of the deformation and metamorphism occurred before westward thrusting along the Long Branch and related faults and the juxtaposition, probably toward the end of the Alleghanian, of the eastern, higher grade metamorphic terranes with the greenschist-facies rocks of the Chopawamsic Formation. The granitoids of the Falmouth Intrusive Suite were mostly intruded prior to the final thrusting along the Long Branch fault.

CHAPTER III

TRIASSIC AND JURASSIC SEDIMENTARY AND INTRUSIVE ROCKS OF THE CULPEPER BASIN

By Albert J. Froelich, Robert E. Weems, Ronald J. Litwin, and Joseph P. Smoot

INTRODUCTION

The early Mesozoic Culpeper basin is a deep, elongate, northeast-trending structural trough filled with Upper Triassic and Lower Jurassic continental clastic rocks, diabase dikes and sills, and basaltic flows (Roberts, 1928; Lee and Froelich, 1989). Our map encompasses much of the narrow, southern end of the basin in which only the Triassic part of the sedimentary section and the Jurassic diabase intrusions are preserved. Here, as elsewhere in the basin, the Triassic sediments are mainly red and gray mudstone, siltstone, sandstone, and conglomerate deposited in lacustrine and fluvial environments. In the western part of the basin near Culpeper, Va., these deposits are estimated to be as much as 2,180 m (7,153 ft) thick (Lee and Froelich, 1989). The thick sequences of Lower Jurassic mudstone and siltstone, which occur to the north of the map area in the much wider and deeper central part of the basin, are absent here because of erosion or nondeposition.

SEDIMENTARY ROCKS

Manassas Sandstone

The Rapidan Member of the Manassas Sandstone (Ꞧmr), which constitutes the basal deposits of the Culpeper basin, is predominantly conglomerate having pebble- to boulder-sized clasts, sandstone, and siltstone of fluvial origin. The angular to subangular conglomerate clasts consist mainly of grayish-green, grayish-olive-green, and dusky green metabasalt derived from the underlying Catoctin Formation. Clasts of quartzite, vein quartz, schist, feldspathic sandstone, and marble (Everona Limestone?) occur in lesser amounts. The conglomerate is typically poorly sorted, thin, and lenticular and grades upward into poorly sorted sandstone that locally has small-scale trough crossbedding. The greenstone conglomerates and the grayish-red to dusky red sandstones and siltstones are cemented firmly by clay and silica and, locally, by secondary calcite. In the upper part of the member, dark-red sandstone fines upward to reddish-brown, micaceous, sandy siltstone having a hackly fracture. The type section of the Rapidan Member is along the south side of the Rapidan River in the extreme southeastern corner of the Culpeper East 7.5-min quadrangle (see measured section 1, Appendix A *of* Lee and Froelich, 1989). The member pinches out to the northeast along the eastern margin of the Culpeper basin where it overlies the Catoctin and True Blue Formations with profound unconformity.

The unnamed upper member of the Manassas Sandstone (Ꞧmu) consists of upward-fining sequences of pebbly sandstone, sandstone, siltstone, and minor amounts of shale. The sandstone is dark-greenish-gray, pinkish-gray, and grayish-red, fine to coarse lithic arenite that consists of feldspar, quartz, chlorite, epidote, and comminuted greenstone rock fragments in a matrix of clay and silt. These sandstones fine upward into reddish-brown, micaceous siltstones that are calcareous and nodular, in part. Some of the siltstone sequences grade upward into dark-red and reddish-brown shale that, in turn, is overlain by lithic arenite of the next overlying sequence.

Balls Bluff Siltstone

The lower, fluvial member of the Balls Bluff Siltstone (Ꞧbf) is characterized by upward-fining successions of trough crossbedded sandstone grading to ripple cross-laminated siltstone and bioturbated silty mudstone. These successions commonly occur within inclined sets that are interpreted as lateral accretion surfaces of muddy, highly sinuous streams. The sandstone commonly contains a large component of mud intraclasts and reworked carbonate nodules suggesting periodic flash floods.

The fluvial member is gradationally overlain by the lacustrine member. The lacustrine member (Ꞧbl) consists of a thick sequence of alternating cycles of reddish-brown silty mudstone and siltstone and light- to dark-gray or black and purple shale and siltstone. These strata are best exposed in an 80-m (262-ft)-thick section in the Culpeper Crushed Stone Quarry just west of Stevensburg (fig. 9). Here, Smoot and Robinson (1988a) recognized 23 sedimentary cycles consisting of laminated to thin-bedded mudstone and siltstone that grade upward into massive, silty to sandy mudstone. They believe that these cycles represent the expansion and contraction of lakes in response to changes in the climate.

The different lacustrine rock types indicate differing subenvironments of deposition. For example, laminated, organic-rich shales are thought to have been deposited in relatively deep perennial lakes. Thin-bedded, gray to purple mudstones having oscillation ripple marks, desiccation cracks, and dinosaur footprints were deposited in shallower lakes that were intermittently dry. Thin-bedded siltstones with irregular scours and abundant cracks were deposited in ephemeral lakes. Massive, silty mudstones with abundant narrow cracks and irregular, ovate to flattened, cement-filled pores represent deposition in dry playa mudflats. Massive, silty to sandy mudstones with numerous downward tapering and branching cement-filled tubes were vegetated mudflats. The cements in deposits of vegetated mudflats and dry playas both include albite, potassium feldspar, calcite, and dolomite. The younger lacustrine deposits of the Balls Bluff Siltstone that are about a hundred meters stratigraphically above those of the Culpeper Crushed Stone Quarry have fewer intervals with characteristics of dry playa mudflats and more intervals with characteristics of vegetated mudflats. Also, the gray to black laminated shales are commonly sparsely fossiliferous (fish scales, ostracodes, and conchostrachans). These relations suggest a general trend to wetter climates up section.

Mountain Run Member of the Tibbstown Formation

Near the faulted northwestern margin of the Culpeper basin, the lacustrine member of the Balls Bluff Siltstone intertongues westward with conglomerate and sandstone of the Mountain Run

Member of the Tibbstown Formation (Ŧtmr) (R.E. Weems, unpub. data). The conglomerates of the Mountain Run are dominantly cobble- to boulder-sized clasts of greenstone, hematitic greenstone, and epidosite but also contain some quartzite, feldspathic sandstone, and vein quartz clasts. The conglomerates are characteristically very poorly sorted and have crudely layered imbricated clasts or randomly distributed clasts supported by an indurated, muddy matrix. These features suggest deposition in shallow streams and debris flows of large alluvial fans that extended southeastward from the Culpeper basin border fault and the relatively uplifted Blue Ridge terrane adjacent to the northwest. Farther to the southeast into the basin, the conglomerates grade laterally into sandstones.

The sandstones of the Mountain Run commonly occur as sheets a few meters thick and are mostly poorly sorted with small scour and fill structures, imbricated-pebble lenses, and flat laminations. They are believed to have been deposited by sheet floods at the toes of alluvial fans. But some of the sandstones are better sorted and have oscillation ripple marks indicating wave reworking in lakes, or they may have climbing ripple cross-lamination and soft-sediment deformation suggesting sheet delta deposition (Smoot, 1991).

PALEONTOLOGY AND AGE OF DEPOSITS

Because of slight to moderate thermal metamorphism associated with the extensive Early Jurassic intrusive diabase bodies, palynomorphs in most of the lacustrine beds are commonly poorly preserved. However, a Late Triassic age for much of the Balls Bluff Siltstone is indicated by palynomorphs in the Andrus core, which appears to contain the oldest productive material. The sampled interval (R395C; depth 162 m (532 ft)) is several hundred feet stratigraphically above the Balls Bluff section exposed in the Culpeper Crushed Stone Quarry (R.J. Litwin and R.E. Weems, unpub. data). The thermally altered but identifiable pollen and spore assemblage includes *Alisporites opii*, *Colpectopollis ellipsoideus*, *Cycadopites fragilis*, *Klausipollenites gouldi*, *Minutosaccus crenulatus*, *Patinasporites densus*, *Pityosporites devolvens*, *Praecirculina granifer*, and *Pyramidosporites traversei*. This assemblage indicates that the lower part of the Balls Bluff Siltstone is early Norian. Another palyniferous sample (R4128B), stratigraphically higher and geologically younger (early to middle Norian) than the Andrus sample, was recovered from a site on Mountain Run just east of Culpeper; it includes the palynomorphs *Alisporites opii*, *Colpectopollis ellipsoideus*, *Corollina meyeriana*, *Klausipollenites* sp., *Kyrtomisporis laevigatus*, *Patinasporites densus*, *Praecirculina granifer*, and *Pyramidosporites traversei* (R.J. Litwin and R.E. Weems, unpub. data). A third palyniferous sample from the Balls Bluff Siltstone, which lies stratigraphically between the other two productive samples, was recovered from a gray mudstone intercalated with siltstone and sandstone near the intertonguing contact with Mountain Run conglomerates near Route 15/29 east of Culpeper. A tentative age of Late Triassic, early to middle Norian, was determined by Bruce Cornet (oral commun., 1982).

Although no skeletal remains of dinosaurs have been reported from the southern part of the Culpeper basin, abundant saurischian dinosaur trackways occur at two separate horizons in the Balls Bluff Siltstone section exposed at the Culpeper Crushed Stone Quarry (fig. 10). The upper horizon of trackways has been described by Weems (1987). Subsequently, the laterally more extensive and better preserved lower horizon has been described by Weems (1992) and R.J. Litwin and R.E. Weems (unpub. data). To the north of the map area, near Dulles International Airport, the skeletal remains of a large parasuchian dinosaur were excavated from calcareous red mudstone of the lower part of the Balls Bluff Siltstone (Weems, 1979). This discovery supports a Late Triassic age for the Balls Bluff as indicated by the palynomorph studies.

STRUCTURE

In the map area, the dominant structure of the early Mesozoic Culpeper basin is a gentle northwest-dipping homocline. At Stevensburg, the Culpeper basin is bisected by a major north-trending horst of pre-Mesozoic crystalline rocks that forms "The Ridge." The part of the basin lying west of "The Ridge" and north of Mount Pony is mainly underlain by a broad, gentle, north-plunging syncline, whereas the basinal areas to the south and east are extensively intruded by the Rapidan and Germanna Bridge diabase sheets.

Whereas the canoe-shaped Rapidan sheet (Jdqh) is largely orthopyroxene-bearing cumulate diabase, the sill-like, northwest-dipping Germanna Bridge sheet (Jdqh and Jdqhg) contains abundant evolved (mainly granophyric) rocks. The sheets are probably cogenetic but are apparently separated by one or more poorly exposed, largely inferred, north-striking faults downthrown to the east. Parts of the hornfels roof of both sheets appear to be preserved in this area. The hornfels roof of the Germanna Bridge sheet, which is exposed along Mountain Run in the north-central part of the Germanna Bridge 7.5-min quadrangle, may mark the trough of a longitudinal syncline. Here, the parallel belts of granophyric rocks to the north and south may be duplicated by obscure parallel folds or faults.

The basal Mesozoic strata along much of the linear southeastern flank of the basin are cut by a poorly exposed, northeast-striking, northwest-dipping normal fault. Near the northern edge of the map and northeast of the Rappahannock River, this longitudinal normal fault is clearly offset by an east-northeast-striking fault. To the south, along the Rapidan River, the basal Mesozoic strata are cut by poorly exposed, steeply dipping, north-, northwest-, and northeast-striking normal faults that form a series of minor grabens and horsts along the scalloped margin of the basin.

Numerous minor longitudinal and transverse flexures, faults, and intrusive diabase dikes, plugs, and sheets are associated with the major structures, thus even the complex structure portrayed is probably oversimplified. Faulting probably began in the Late Triassic, but at least some post-Early Jurassic faulting is suggested.

GEOCHEMISTRY AND PETROLOGY OF DIABASE

Regional geologic, petrologic, and geochemical relations between the high-titanium, quartz-normative (HTQ) diabase sheets in the Culpeper basin are documented in Froelich and Gottfried (1988). Suites of geochemical samples collected along Virginia Routes 655 and 661 show that the Rapidan sheet is a quartz-normative tholeiite with average chilled margin contents of 7.8 weight percent[1] MgO, 1.1 percent TiO_2, and 52 percent SiO_2. The thick intrusive sheet consists mostly of cumulus orthopyroxene (bronzite), clinopyroxene (augite and pigeonite), and plagioclase (labradorite-bytownite). Major and trace element geochemistry of all diabase samples in the map area is documented in Gottfried and others (1991). The content of MgO increases to a maximum of 12.4 percent in the near basal orthopyroxene-rich norite but averages about 10 percent MgO across the sheet. The geochemistry of the Rapidan sheet along Virginia Route 655 is discussed in detail by Tollo and others (1988; field trip stops 4A–C).

The ferrogabbro, ferrodiorite, and pegmatoid of the Germanna Bridge sheet are medium to coarse crystalline, with plagioclase,

[1]All geochemical analyses of oxides in this chapter are given in weight percent.

large blades of clinopyroxene, and minor quartz, orthoclase, and ilmenomagnetite. Granophyre and syenite contain abundant orthoclase and albite and prominent stubby to acicular phenocrysts of hornblende and actinolite, with accessory ilmenite, magnetite, quartz, biotite, and apatite. Minor aplite pods and dikes consist of finely crystalline feldspar and quartz, and sparse hornblende and (or) biotite. Geochemical samples collected from the granophyre and ferrogabbro and the enclosing diabase are partially documented in Tollo and others (1988; field trip stop 6). Granophyre and ferrogabbro differentiates of the HTQ diabase samples contain from 0.77 to 4.28 percent MgO, 1.6 to 1.8 percent TiO_2, and 52.2 to 59 percent SiO_2, values characteristic of such evolved HTQ rocks elsewhere in the Culpeper basin. Samples from medium-grained diabase enclosing these evolved rocks contain 5.5 to 6.07 percent MgO, 0.82 to 1.21 percent TiO_2, and 52.0 to 52.4 percent SiO_2 (Gottfried and others, 1991).

The low-titanium, quartz-normative (LTQ) diabase dikes in the map area were not sampled for geochemistry. However, nearby related dikes are quartz-normative tholeiite with chilled margin contents of about 7.5 percent MgO, 0.75 percent TiO_2, and 51.5 percent SiO_2, as documented by Gottfried and others (1991). The tholeiite is porphyritic, fine- to medium-crystalline diabase consisting mainly of intergrown clinopyroxene (augite) and calcic plagioclase (labradorite-bytownite) as groundmass with clusters of centimeter-size calcic plagioclase (labradorite-bytownite) phenocrysts. Wherever seen, the LTQ dikes cut the HTQ sheets, and dike margins are aphanitic and chilled against medium- to coarse-crystalline HTQ diabase.

ECONOMIC GEOLOGY

Potential rock and mineral resources in the Culpeper basin are chiefly those useable for construction materials such as crushed stone, dimension stone, common brick, and tile. The thick, orthopyroxene-bearing diabase of the Rapidan sheet is a major potential source for crushed stone or ornamental stone; however, active quarries in this unit are situated in the adjacent Rapidan and Culpeper West quadrangle areas and apparently supply all of the current needs. Rock for crushed stone of similar quality also is available in the Germanna Bridge diabase sheet. A large quarry in indurated Balls Bluff Siltstone is operating west of Stevensburg and south of Virginia Route 3; immense reserves of similar layered rock material are present to the south and west of the active quarry. Formerly, the high-grade hornfels beneath the Rapidan diabase sheet was extensively quarried from a pit east of U.S. Route 522 and north of the Rapidan River, and large reserves of similar material are present adjacent to the Rapidan and Germanna Bridge sheets. According to Calver and others (1961), weathered clay, shale, and mudstone from the Balls Bluff Siltstone and Manassas Sandstone are suitable for common brick and tile. Base-metal sulfides (Pb, Zn, Cu, Ag, and so on) are present in significant but subeconomic abundances in some of the lacustrine shales and sandy siltstones of the Balls Bluff Siltstone (Smoot and Robinson, 1988b). Ground-water resources of good quality in volumes adequate for domestic purposes are available at depths of less than 90 m (295 ft) in most parts of the Culpeper basin that are underlain by fractured sedimentary rocks.

ENVIRONMENTAL GEOLOGIC AND HYDROLOGIC FACTORS

A variety of regional geologic and hydrologic factors that bear on land-use planning and decisionmaking in the Culpeper basin area are discussed in U.S. Geological Survey Miscellaneous Investigations Series Map I–1313, which comprises nine maps published at a scale of 1:125,000 (1 in=2 mi). A list of the maps and their main topics follows:

I–1313–A	Morsches and Zenone (1981) Flood studies
I–1313–B	Froelich and Leavy (1981) Mineral resources
I–1313–C	Leavy and others (1983) Bedrock map and geotechnical properties of rocks
I–1313–D	Posner and Zenone (1983) Chemical quality of ground water
I–1313–E	Froelich (1985) Geotechnical properties of surface materials
I–1313–F	Laczniak and Zenone (1985) Ground-water resources
I–1313–G	Leavy (1984) Lineaments and planar and linear structural features
I–1313–H	Lynch and others (1987) Low-flow characteristics and chemical quality of streams
I–1313–J	Froelich (1989) Geologic and hydrologic factors affecting land-use planning

CHAPTER IV

UPPER MESOZOIC AND CENOZOIC DEPOSITS OF THE ATLANTIC COASTAL PLAIN

By Robert B. Mixon, David S. Powars, Robert E. Weems, J. Stephen Schindler, Wayne L. Newell, Lucy E. Edwards, and Lauck W. Ward

STRATIGRAPHY

Potomac Formation

The Lower Cretaceous Potomac Formation (Kp), consisting mainly of unlithified sand, gravel, silt, and clay of fluvial and deltaic origin, is the basal Coastal Plain formation in much of northeastern Virginia. In the map area, the formation's generally wide outcrop belt parallels the Fall Line[1] from the Ni River valley northward to Fredericksburg and hence northeastward through Stafford County (Mixon and others, 1989a; Glaser, 1969). In both the outcrop and the subsurface, the Potomac strata overlie, with great angularity, the crystalline rocks of the Proterozoic-Paleozoic basement complex and red beds and volcanic rock of the buried part of the early Mesozoic Taylorsville basin (Weems, 1980). Near the Fall Line, the Potomac Formation thins westward in a steplike manner across the faulted Coastal Plain margin (cross sections *A–A'*, *D–D'*; Mixon and Newell, 1977, 1978). The Potomac beds thicken to the east and northeast and are thought to be as much as 500 m (1640 ft) thick in the section penetrated by Texaco's W.B. Wilkins No. 1 oil test well near Maple Grove, Va., in the easternmost part of the map area (Milici and others, 1991). In the deeper parts of the Salisbury structural embayment on the Eastern Shore of Maryland and Virginia, the Lower Cretaceous section may be as thick as 1000 m (3281 ft), or more (Spangler and Peterson, 1950; Brown and others, 1972).

In Maryland, equivalent Lower Cretaceous deposits are assigned to the Potomac Group, which includes, from bottom to top, the Patuxent Formation, Arundel Clay, and Patapsco Formation (see Cleaves and others, 1968). The names Potomac Group, Patuxent Formation, and Patapsco Formation were formerly used in Virginia, but are no longer in use there by the U.S. Geological Survey because the Lower Cretaceous beds are not readily divisible into formational units (Seiders and Mixon, 1981; Meng and Harsh, 1988; Drake and Froelich, 1986).

Numerous excellent exposures of the Potomac Formation show that, in updip areas, the formation consists mainly of lenticular bodies of trough-crossbedded, poorly sorted, medium to very coarse gravelly sand and sandstone (fig. 11). Cut and fill structures and intraformational clay-clast conglomerates are common. Locally, these coarse sediments grade upward into fine-grained, flaggy sandstone, laminated and micaceous siltstone, and silty clay that contains abundant carbonized plant fragments and, more rarely, stem and leaf impressions of ferns, cycads, conifers, and flowering plants. Lesser amounts of poorly sorted, greenish-gray, silty clay and clayey fine sand fill small to large channels of former streams; rarely, red-mottled sediment of this type occurs in thin, tabular bodies that may represent overbank deposits. In the innermost Coastal Plain, the preponderance of coarse, sandy and gravelly sediment in upward-fining channel-lag and channel-bar sequences, the high-energy bedforms, the abundant plant material, and the lack of marine fossils indicate that the Potomac Formation was deposited as a gently seaward-sloping alluvial plain formed by coalescing, braided-river systems.

The lithology and paleontology of the Potomac Formation in the subsurface of the eastern part of the map area is best known from studies of the core from a USGS stratigraphic test hole drilled on the upland, 3.7 km (2.3 mi) west-southwest of Oak Grove, Va. (Reinhardt and others, 1980a). The test hole penetrated 284 m (931 ft) of the Potomac Formation between greensand of the upper Paleocene Aquia Formation and the bottom of the hole at a depth of 422 m (1385 ft). In this core section, Reinhardt and others (1980a, fig. 2) informally divided the Potomac Formation (still known as the Potomac Group at that time) into two major upward-fining sequences. Their lower Potomac Group, which is 125.6 m (412.1 ft) thick, consists of a lower unit of mostly thick-bedded sand and clay-clast conglomerate and an upper unit composed of thinner bedded sand interbedded with laminated carbonaceous clay. Their upper Potomac group is 158 m (518 ft) thick of which the lower 96.6 m (316.9 ft) consists of upward-fining, dominantly sandy sequences that grade upward to laminated or massive clay. The upper 61 m (200 ft) of the upper Potomac is dominantly greenish-gray silty clay and clayey silt, in part highly oxidized and mottled yellowish and reddish brown. Some beds contain clay-filled, vertical to horizontal burrows possibly made by worms; radiating patterns of very small tubules may be traces of plant rootlets (Reinhardt and others, 1980a, figs. 17, 18). These clayey and silty sediments were thought to be indicative of deposition in subaerial flood plain areas.

In the map area an Early Cretaceous age for the Potomac Formation is based on studies of angiosperm megafossils (leaves of flowering plants) from a few outcrop localities (Doyle and Hickey, 1976) and palynomorphs (spores and pollen) from outcrop localities and from the Oak Grove core (Doyle and Robbins, 1977; Reinhardt and others, 1980a). Brenner's (1963) zonation of the Potomac Group in Maryland provides the framework for these palynomorph studies.

In the map area, the most notable fossil plant locality is in an abandoned sand and gravel pit east of Aquia, Va., and just south of the northern edge of the map sheet (see Widewater 7.5-min quadrangle). The pit, located on Engineers Road about 0.8 km (0.5 mi) west of the Potomac River and 0.8 km (0.5 mi) south of Chopawamsic Creek, exposes the uppermost part of the Potomac Formation and the basal beds of the overlying Aquia Formation.

[1]Differential erosion along the contact between the soft Coastal Plain strata and the hard Piedmont crystalline rocks commonly causes falls or rapids to form in streams at or slightly west of the boundary between the two terranes. The Fall Line, a term applied to the imaginary line connecting the falls or rapids in successive streams crossing the Piedmont-Coastal Plain boundary, is the upstream limit of navigability on major rivers and consequently the site of important colonial cities such as Richmond, Washington, and Baltimore (Mixon and others, 1981).

The fossil plants occur in silty clay beds about 15 m (49 ft) above the base of the pit. The diverse flora at this locality, including one species of horsetail (*Equisetum*), one species of cycadophyte, nine species of conifers, and twelve species of flowering plants, are described in detail in Upchurch and others (1994). The most abundant plant species in the fossil flora is *Nelumbites extenuinervis*, an extinct relative of the water lotus; the presence of this fossil and other aquatic angiosperms together with the clayey sediment and excellent preservation led Upchurch and his co-workers to conclude that the fossil-bearing beds represent deposition in a pond or swale. The floral assemblage is considered to be of middle to early late Albian age.

Pamunkey Group

Over much of the map area, a thin to thick blanket of glauconitic sand and silt of marine origin angularly overlies the eroded, irregular, upper surface of the Lower Cretaceous Potomac Formation. These glauconitic strata were named the Pamunkey Formation by N.H. Darton (1891) for the extensive exposures along the Pamunkey River to the south of the map area in Caroline, Hanover, and King William Counties, Virginia. Subsequently, Clark and Martin (1901) raised the Pamunkey Formation to group status and divided it into the Aquia Formation, named for Aquia Creek in Stafford County, Virginia, and the Nanjemoy Formation, named for Nanjemoy Creek in Charles County, Maryland. These streams are tributaries of the Potomac River in the northeastern part of the Fredericksburg map area. Clark and Martin (1901, p. 65) also recognized a thin but laterally very extensive pink to red clay, referred to informally as the Marlboro Clay, which separates the Aquia and Nanjemoy greensand sequences. Aquia and Nanjemoy sections exposed along the Potomac River from Glymont, Md., to Popes Creek, Md., were further divided into 17 zones or beds by Clark and Martin (1901, p. 57–71) based on lithology and biostratigraphy. The Piney Point Formation (Otton, 1955), which unconformably overlies the Nanjemoy in the Pamunkey River outcrops and elsewhere in the outer Coastal Plain (Ward, 1985; Mixon and others, 1989b; Gibson and Bybell, 1984, 1994), also is included in the Pamunkey Group but has not been recognized in the map area.

Aquia Formation

In the map area, the basal formation of the Pamunkey Group is the upper Paleocene Aquia Formation (Ta), a variably shelly, glauconitic quartz sand interbedded with a few thick to very thick beds of limestone and carbonate-cemented quartz arenite. The formation is as much as 35 m (115 ft) thick. In the Fredericksburg area, as in much of the northern Virginia Coastal Plain, the Aquia can be divided into two lithic units: the more poorly sorted, muddier, and more calcareous Piscataway Member (lower) and the better sorted, much less calcareous Paspotansa Member (upper) (see Clark and Martin, 1901; Ward, 1985; Powars, 1987). These members are not differentiated on the geologic map or cross sections. In the absence of an adequately described type section, the fossiliferous greensand outcrops in bluffs on the western side of the Potomac River estuary, 2.4 km (1.5 mi) south of the mouth of Aquia Creek in Stafford County, Virginia, have been designated the principal reference section for the Aquia Formation and its two constituent members (Ward, 1985, loc. 3).

Piscataway Member (Revised)

The Piscataway Member of the Aquia Formation was named for exposures of glauconitic sand along Piscataway Creek, Prince Georges County, Maryland (Clark and Martin, 1901). As originally defined by Clark and Martin (1901, p. 60), the Piscataway Member included the lower seven lithic "zones" of the seventeen "zones" or beds composing the Aquia-Nanjemoy sequence. The boundary between the Piscataway and the overlying Paspotansa beds was moved lower in the section by Ward (1985, p. 9) who placed the contact at the top of "zone" 5 to agree more closely with the upward change from poorly sorted, clayey sand to well sorted, cleaner sand. More recently, studies of the cores from the USGS's Lake Jefferson and Lake Jefferson East stratigraphic test holes, about 1.6 km (1 mi) northwest of Comorn, Va., suggest that a more natural and more useful boundary between the two members would be slightly lower—at the base rather than the top of "zone" 5 (Powars, 1987; Edwards, 1989). In the Lake Jefferson East core, the most conspicuous lithic break in the Aquia greensand section occurs at an altitude of -12.2 m (-40 ft) where a one-meter-thick lag of fine to coarse, shelly and pebbly sand at the base of the Paspotansa coarsens downward and is burrowed into very fine, muddy sand of the Piscataway. Studies of dinocyst assemblages from the Aquia Formation in the nearby Lake Jefferson corehole (L.E. Edwards, oral commun., 1991) indicate that a floral break coincides with the change in lithology across the burrowed contact and suggest that it may represent a minor unconformity. Edwards finds that a relatively low-diversity dinocyst assemblage in Aquia strata below the pebbly sand and burrowed contact is dominated by the *Glaphyrocosta exuberans* complex. Above the burrowed contact, a higher diversity dinocyst assemblage is marked by an influx of *Impagidinium* sp. of Edwards and others (1984). *Kallosphaeridium brevibarbatum* de Coninck and, less consistently, *Apectodinium homomorphum* (Deflandre & Cookson) Lentin & Williams. In some cores, a transition occurs between these two assemblages, marked by the lowest occurrences of *Adnatosphaeridium robustum* (Morgenroth) de Coninck, *Cassidium* sp. of Edwards and others (1984), and *Eocladopyxis peniculata* Morgenroth.

Over much of the type area of the Aquia Formation, the Piscataway Member is commonly 9 to 12 m (30–39 ft) thick. In both cores and newly exposed outcrop sections, the abrupt contact between the dark, glauconitic shelf sand of the basal Piscataway and the directly underlying, much lighter colored and partly oxidized fluvial-deltaic sand and clay of the Lower Cretaceous Potomac Formation is easily recognized and is a major erosional unconformity. Upper Cretaceous strata and the thin Brightseat Formation (Hazel, 1969; Hazel and others, 1984) of early Paleocene age, which occur below the Aquia Formation at many localities in Maryland, do not appear to be present in the map area. The base of the Piscataway is irregular to planar and, in updip areas, is commonly marked by 0.3 to 1.0 m (1–3.3 ft) of pebbly sand or sandy gravel composed mainly of well-rounded, fine to coarse pebbles of quartz. In the furthest updip areas to the west, the glauconitic sand fill of shallow channels cut into the top of the Potomac beds includes cobble- to boulder-sized clasts of quartz and, less commonly, metamorphic rocks. A narrow, 8-m (26-ft)-deep channel at the base of the Aquia, exposed in the Richmond, Fredericksburg, and Potomac railroad cuts south of Aquia Creek, was noted by N.H. Darton. This sharply incised drainage feature is on the uplifted, western side of a main strand of the Brooke fault zone—suggesting that topographic relief along a preexisting fault-line scarp may have accentuated the depth of incision.

The coarse basal lag of the Piscataway grades abruptly upward to fine to medium glauconitic quartz sand in thick to massive beds. Green and black, sand-sized pellets of glauconite commonly constitute 20 to 80 percent of the sediment; most pellets are polylobate. Mollusks are concentrated in thick to very thick beds; locally, cementation by calcium carbonate and silica forms indurated, ledge-forming beds of sandy limestone. The dominant taxa include large bivalves (clams and oysters) and the high-spired gastropods *Turritella humerosa* and *Turritella mortoni* (Ward, 1985). Shark's teeth, fish scales and bones, and the remains of sea turtles (Weems, 1988) are minor constituents.

In western, updip areas of outcrop, the Piscataway Member contains less clay and silt and much less glauconite. Black heavy-mineral cross-laminations, clay-lined burrows of the *Ophiomorpha nodosa* type, and small bars composed of clean, relatively well-sorted, crossbedded sand become more common upward in the section—indicating deposition in shoaling-upward waters (fig. 12).

Paspotansa Member (Revised)

The Paspotansa Member was named for Paspotansa (now Passapatanzy) Creek, a tributary of the Potomac River in Stafford County, Virginia (Clark and Martin, 1901). As defined by Clark and Martin (1901), the Paspotansa Member comprised fine to very fine glauconitic sand containing abundant *Turritella mortoni* (their lithic "zones" 8 and 9). In this report, we follow Ward's (1985) revision of the Piscataway and Paspotansa Members, which places lithic "zones" 6 and 7 of Clark and Martin (1901) in the lower part of the Paspotansa. As discussed in the section on the Piscataway Member, we suggest that the definition of the Paspotansa Member be further modified to include lithic "zone" 5 of Clark and Martin (1901) as the lowermost part of the Paspotansa.

The Paspotansa Member is extensively exposed along the Virginia shore of the Potomac River estuary in Stafford and northwestern King George Counties. The principal reference section for the Paspotansa Member is the high river bluff 2.4 km (1.5 mi) below the mouth of Aquia Creek and Youbedamn Landing (see Widewater and Passapatanzy 7.5-min quadrangles). There, the Paspotansa consists of 19.4 m (63.6 ft) of thick-bedded to massive, variably shelly, very fine to fine, micaceous, glauconitic quartz sand. Gray- to olive-black sand in fresh exposures weathers to yellowish gray and grayish orange. A few medium to thick beds of sandy limestone and boulder-sized concretions are interbedded with the dominantly unconsolidated sands. Here, as elsewhere, discontinuous, lenticular beds of *Turritella mortoni* are characteristic of the member. At this locality, the contact between the Paspotansa and the overlying thin (0.23 m (0.8 ft)) remnant of Marlboro Clay is sharp but conformable. Excellent exposures of the Paspotansa Member also occur in Bull Bluff at the mouth of Potomac Creek.

At the principal reference section and elsewhere in the type area of the Aquia Formation (see Lake Jefferson East and Caledon State Park coreholes; Powars, 1987), the contact between the Paspotansa Member and the underlying Piscataway Member appears to be disconformable. A similar sharp, burrowed contact between the basal pebbly sand of the Paspotansa and the much clayier and more calcareous sands of the underlying Piscataway was also reported in a study of the Haynesville cores in Richmond County, Virginia, about 62 km (38.5 mi) to the southeast of the Lake Jefferson East corehole (Mixon and others, 1989b).

Marlboro Clay

The term "Marlboro clay" was first used by Clark and Martin (1901) to refer to a thin (±6 m (±20 ft)), but conspicuous, unit of red, pink, and gray clay that occurs above the Aquia Formation greensands and below similar greensands of the Nanjemoy Formation. The clay is typically exposed near the town of Upper Marlboro in eastern Prince Georges County, Maryland. Probably because of its clayey nature, early workers considered the Marlboro to be the basal bed of the overlying Nanjemoy Formation and commonly lumped it in measured sections with the lower Potapaco clay member of the Nanjemoy (Clark and Miller, 1912; Darton, 1951). Subsequently, Glaser (1971) described and mapped the Marlboro Clay over a wide area in southern Maryland and elevated the unit to formational rank. The Marlboro is now known to occur throughout much of the Maryland and Virginia Coastal Plain west of the Chesapeake Bay (Cederstrom, 1957; Sinnott, 1969; Brown and others, 1972; Hansen, 1974; Powars and others, 1992; Gibson and Bybell, 1994).

In the Fredericksburg map area, the Marlboro Clay (Tmc) consists of as much as 8.5 m (28 ft) of light- to dark-gray and light-brown to pale-reddish-brown kaolinitic clay and clayey silt. Rounded and irregularly shaped concretions of marcasite(?) are common in outcrops along the Potomac River near Fairview Beach. In most outcrops, the unit appears to be massive; however, cores and dewatered outcrops show the laminated to thin-bedded character of the unit. Silty clay is commonly interbedded with laminated and ripple-cross-laminated fine silt—suggesting a low-energy depositional environment. Locally, in updip areas, small scour channels filled with sandy silt indicate weak bottom currents. In these areas, molds of small mollusks, agglutinated foraminifers, a low-diversity dinocyst assemblage, and the freshwater alga *Pseudoschizaea* suggest deposition in brackish water or a very restricted shelf area near the mouth of a river. However, in the Loretto core from the southeasternmost part of the map area, the Marlboro contains a diverse dinocyst assemblage including *Adnatosphaeridium multispinosum* Williams & Downie, *Apectodinium augustum* (Harland) Lentin & Williams, *A. homomorphum* (Deflandre & Cookson), *A. parvum* (Alberti) Lentin & Williams, *Catillopsis abdita* Drugg, *Fromea fragilis* (Cookson & Eisenack) Stover & Evitt, *Muratodinium fimbriatum* (Cookson & Eisenack) Drugg, *Phthanoperidinium crenulatum* (de Coninck) Lentin & Williams, and *Sengalinium? dilwynense* (Cookson & Eisenack) Stover & Evitt. These taxa indicate a less restrictive depositional environment for the Marlboro in downdip areas and suggest an age near the Paleocene-Eocene boundary.

The conformable contact between the Marlboro Clay and the underlying Aquia Formation is commonly marked by interbedding, over a 0.5 to 1.0 m (1.6–3 ft) interval, of thin beds of kaolinitic Marlboro Clay and fine glauconitic sand of the uppermost Paspotansa Member of the Aquia. Less commonly, the contact may be an abrupt gradation from fine sand (below) to gray clay and silt of the Marlboro (above). An unconformable relation with the overlying Potapaco Member of the Nanjemoy Formation is indicated by the abrupt, uneven contact and by large, sand-filled burrows extending from the base of the Potapaco downward for 1 to 2 m (3–6.5 ft), or more, into the Marlboro Clay. The burrow fill is poorly sorted, muddy, glauconitic quartz sand commonly containing shells or fragments of shells and scattered very fine pebbles.

The thickest known occurrence of the Marlboro Clay in the map area is in the northernmost part of the Dahlgren 7.5-min quadrangle where the Ashton corehole, sited on the high ridge just east of the mouth of Chotank Creek, penetrated an 8.5-m (27.9-ft)-thick section of the clay. The Marlboro thins generally westward to eastern Stafford County where the Brooke fault zone (fig. 2) is the present-day western limit of outcrop. The Marlboro Clay is absent because of erosion and (or) nondeposition over a wide area including the structural high along the Skinkers Neck anticline and the relatively downwarped area adjacent on the west of the anticline (much of Bowling Green and Rappahannock Academy 7.5-min quadrangles, eastern part of Passapatanzy quadrangle, and western King George quadrangle). That the absence of the Marlboro in this area is mainly the result of erosion is suggested by outcrops near the mouth of Potomac Creek on the ridge just west of Bull Bluff (see Passapatanzy 7.5-min quadrangle). There, the basal, poorly sorted sand of the Nanjemoy Formation, containing well-rounded pebble- and cobble-sized clasts of the Marlboro Clay, directly overlies the Aquia greensand. The absence of the Marlboro Clay in both the structurally high and low areas along the Skinkers Neck anticline suggests that some arching of the Paleocene beds along this structure occurred in latest Paleocene or earliest Eocene time—before the main faulting or folding took place.

Nanjemoy Formation

In the Fredericksburg map area, the upper part of the Pamunkey Group is the lower Eocene Nanjemoy Formation (Tn), a generally poorly sorted, clayey and silty, glauconitic sand as much as 64 m (210 ft) thick. The Nanjemoy unconformably overlies the Marlboro Clay or the Aquia Formation and, in turn, is overlain with great unconformity by the lower and middle Miocene Calvert Formation. The Nanjemoy-Calvert unconformity is a 35- to 40-Ma hiatus that includes much of middle and late Eocene, Oligocene, and early Miocene time. However, in the more complete stratigraphic sections of the middle Coastal Plain of Virginia and Maryland, glauconitic sands of the middle Eocene Piney Point Formation and the upper Oligocene Old Church Formation are present between the Nanjemoy and the Calvert (Mixon and others, 1989b; Bybell and Gibson, 1994).

In the inner Coastal Plain of Virginia and in parts of Charles and Prince Georges Counties, Maryland, the Nanjemoy Formation can be subdivided into a lower, very clayey and very glauconitic quartz sand (the Potapaco Member, Tnp) and an upper, somewhat less clayey, very micaceous, glauconitic quartz sand containing scattered very fine pebbles of quartz (the Woodstock Member, Tnw). The distribution of the two members in the subsurface of the map area is shown by cross sections *A–A′* and *D–D′*. The members are not differentiated on the map itself. The lithology and paleontology of the Nanjemoy Formation and the Potapaco and Woodstock Members have been discussed by various workers (Clark and Martin, 1901; Clark and Miller, 1912; Darton, 1951; Glaser, 1971; Frederiksen, 1979; Gibson and others, 1980; Reinhardt and others, 1980b; Ward and Krafft, 1984; Ward, 1985; Powars, 1987).

Potapaco Member

The Potapaco Member consists of very poorly sorted, clayey and silty, glauconitic sand and sandy clay-silt that are sufficiently cohesive to allow good core recovery from the unit. In the Virginia part of the map area, the member ranges in thickness from about 25 m (82 ft) in northeastern King George County (Ashton core, Powars, 1987) to about 8 m (26 ft), or less, along the Brooke fault zone, which is the present-day updip limit of outcrop of the Potapaco in eastern Stafford County. The Potapaco Member is probably thicker in Maryland in the northeastern corner of the map area, but there the section has not been studied in detail and, consequently, is poorly known. In updip areas, Paleogene units thin westward to a feather edge as the result of truncation by onlapping, successively younger formations. For example, at the Caroline Stone Quarry near the edge of the Coastal Plain just south of the map area, a 1.5-m (5-ft)-thick remnant of the Nanjemoy (equivalent in part to the Potapaco Member?) laps over an equally thin Aquia section onto crystalline rocks of the Piedmont. The Nanjemoy strata, in turn, are strongly truncated by the overlying middle Miocene beds of the Calvert and Choptank(?) Formations. At the western end of the quarry, the Miocene beds lap over the Nanjemoy onto the Piedmont crystalline rocks.

The Potapaco Member was originally described by Clark and Martin (1901, p. 65–73) who subdivided the unit into their lithic "zones" 10 to 15. Their "zone" 10 included the Marlboro Clay at the base. As their "zones" were poorly described and difficult to correlate with sections away from the Potomac River, later workers have developed a more regionally useful subdivision based mainly on lithology and sequence concepts (Ward, 1985; Powars, 1987; Mixon and others, 1989b). Thus, within the Fredericksburg map area, the Potapaco Member can be divided into five upward-fining sequences which are herein designated Units A (oldest) to E (youngest).

Unit A.—Unit A of the Potapaco Member consists of 2 to 8 m (7–26 ft) of olive-black, olive-gray, and greenish-gray, sparsely glauconitic quartz sand containing scattered small mollusks and shell fragments. One or two beds of shelly limestone (or concretion zones?) as much as 0.3 m (1.0 ft) thick occur in the unit in western and northern King George County, Virginia. The dominantly fine to very fine sand is variably muddy and very micaceous. Obscure, small-scale, wavy bedding is evident in some outcrops. The unit commonly coarsens downward to a basal bed of fine to coarse sand containing scattered very fine to medium pebbles of quartz and phosphate and rip clasts of the underlying Marlboro Clay.

The abrupt, burrowed contact between unit A and the underlying Marlboro Clay, observed in outcrops and cores in much of the map area, is a regional unconformity also noted in the Richmond and Hopewell areas of central Virginia and in Prince Georges and Calvert Counties, Maryland. Where the Marlboro Clay is absent in parts of King George and Caroline Counties, Virginia, unit A commonly directly overlies the Paspotansa Member of the Aquia Formation.

The upper part of unit A and the lowermost part of unit B are exposed in the southern bank of the Rappahannock River, opposite Goat Island, Caroline County, Virginia. The low diversity of the molluscan assemblage in unit A at this locality (*Venericardia potapacoensis*, *Macrocallista* sp., *Corbula* sp., *Lucina* sp., and *Cadulus* sp.) and elsewhere suggests deposition in waters of less than normal salinity. The dinocyst assemblage, which is dominated by a single taxon, also suggests restricted marine conditions (L.E. Edwards, oral commun., 1984).

Unit B.—Unit B of the Potapaco Member consists of 4 to 9 m (13–30 ft) of light-olive-gray to olive-black, very clayey and silty, fine glauconitic quartz sand and sandy clay that disconformably overlies unit A. One or two zones of concretions up to boulder size that occur near the base of unit B may be mappable locally. Unit B sediments also contain abundant small concretions or nodules of iron sulphide and variable amounts of mica and carbonaceous material. Unit B, equivalent to Bed B of Ward (1985), is overall the muddiest part of the Potapaco Member and is characterized by numerous, thin, discontinuous shell beds; here, the small bivalve *Venericardia potapacoensis* is the dominant mollusk. The thicker unit B sections in downdip areas commonly include two upward-fining sequences that are lithically similar and contain the same dinocyst assemblage (see the Ashton, Lake Jefferson, and Lake Madison cores; Powars, 1987).

In the map area, the better exposed, more complete sections of unit B occur in high bluffs and low banks along the Rappahannock River in Caroline and King George Counties, Virginia (see also Ward, 1985; and Rappahannock Academy and Port Royal 7.5-min quadrangles). The Gera and Lake Jefferson East cores (Powars, 1987) provide excellent reference sections for unit B in which its sharp, disconformable contacts with unit A (below) and unit C (above) are clearly defined.

Unit C.—Unit C of the Potapaco consists of 2 to 4.3 m (6.5–14 ft) of dark-greenish-gray to dark-olive-gray, poorly sorted, very clayey and silty, glauconitic quartz sand. The abundance of fine to coarse, dark-green to black glauconite grains commonly gives the unit a darker and coarser grained appearance than the underlying and overlying units. Locally, thin beds and burrow fills are as much as 90 percent glauconite—giving the sediment a completely black aspect. Generally, unit C, which is at least partly equivalent to Ward's Bed C (Ward, 1985), is thick bedded to massive and intensely bioturbated. The very abundant burrows, including both clay-filled burrows and clay-lined, sand-filled burrows similar to *Ophiomorpha*, led Ward to informally name his Bed C the "burrowed Potapaco." One or two thin shell beds, characterized by the small bivalve *Venericardia potapacoensis*, are commonly present in the lower part of the unit.

In our cores, the abrupt contact between unit C and the directly underlying unit B is marked by dark-greenish-gray to olive-black, glauconite-rich, fine to medium sand that is burrowed deeply into light-olive-gray, silty fine sand of unit B. An exposure of the contact may be seen in a steep bank on the southeastern side of Mill Creek at the outlet of Millers Pond (now drained) in Caroline County, Virginia. This site is about 55 m (180 ft) upstream from the U.S. Route 17 bridge over the creek (see Port Royal 7.5-min quadrangle). One of the better exposures of unit C is the 3-m (10-ft)-thick section in banks along the northern side of the Rappahannock River, 0.6 km (0.4 mi) above the mouth of Gingoteague Creek, King George County, Virginia.

Units D and E.—Units D and E are upward-fining sequences of glauconitic quartz sand that constitute the upper part of the Potapaco Member in eastern King George and eastern Caroline Counties and in the northwestern parts of Essex and Westmoreland Counties. These strata are generally very poorly exposed; however, the Gera and Ashton cores, from the southern and northeastern parts, respectively, of King George County, provide excellent sections for study (Powars, 1987). In these cores, units D and E are each 5 to 6 m (16.4–19.7 ft) thick.

Units D and E are very similar, lithologically, to the underlying units B and C but are sandier and, thus, are overall coarser grained than the very clayey and silty strata of the lower Potapaco. Units D and E both have poorly sorted, fine to medium basal sands that are burrowed deeply into the finer grained, upper parts of subjacent units. For example, the irregular surface between unit E (above) and unit D is commonly marked by olive-black or greenish-black, fine to medium sand of unit E burrowed into very light gray to olive-gray, clayey and silty, fine to very fine sand of unit D. This disconformable contact and similar unit contacts in the Potapaco Member may represent minor hiatuses. The hiatuses, which are breaks in the continuity of the stratigraphic sequence, are thought to be caused by transgressive pulsations of the sea (Powars, 1987).

The clayey and silty, micaceous, very fine to fine sand of the middle and upper parts of units D and E are interbedded with very thin to thin (2 to 8 cm) beds of shell dominated by *Venericardia potapacoensis*. *Corbula*, *Lucina*, *Cadulus*, and small *Turritella* are also present.

Woodstock Member

The Woodstock Member of the Nanjemoy Formation was named by Clark and Martin (1901) for exposures in the Potomac River bluffs near the old Woodstock Plantation about 2.5 km (1.6 mi) upstream from Mathias Point, King George County, Virginia (fig. 13). The Woodstock outcrops at this locality, as much as 15 m (49 ft) thick, have been designated the principal reference section (lectostratotype) for the Woodstock Member (Ward, 1985). Clark and Martin (1901, p. 66) describe the Woodstock in the Potomac River area as "fine, homogeneous greensands and greensand marls, that are less argillaceous than the underlying Potapaco beds."

In the Potomac River bluffs and elsewhere in the eastern part of the Fredericksburg map area, the Woodstock is overlain with great unconformity by the lower and middle Miocene Calvert Formation. Middle and upper Eocene, Oligocene, and lowermost Miocene strata are absent as the result of uplift and erosion prior to deposition of the Calvert. In the subsurface a few miles to the east of the map area and in outcrops along the Pamunkey and James Rivers, the Woodstock is overlain by the middle Eocene Piney Point Formation. In updip areas of the Coastal Plain, including the Fredericksburg map area, the contact between the Woodstock and the underlying Potopaco Member of the Nanjemoy is believed to be a minor unconformity, representing a relatively brief hiatus (Ward, 1985; Powars, 1987).

The Woodstock Member consists of 9 to 15 m (30–49 ft) of light- to dark-olive-gray and greenish-black, variably muddy, very fine to coarse, glauconitic quartz sand that is coarser, less glauconitic, and more micaceous than the underlying Potapaco Member. The unit is commonly poorly sorted, sparsely to abundantly shelly, and contains abundant small fragments of woody material. In the Ashton corehole (Powars, 1987) near the lectostratotype section, the base of the Woodstock is marked by a 15-cm (6-in)-thick bed of calcareous concretions. In general, the Woodstock sands contain less clay and silt matrix than the sands of the Potapaco Member. Therefore, the Woodstock tends to be more friable than the Potapaco and easily washes out of the core barrel. Only about 50 percent core recovery was obtained.

An abundant and diverse molluscan fauna has been reported in outcrops of the Woodstock along the Potomac and Pamunkey Rivers (Clark and Martin, 1901; Ward, 1985). The large bivalve *Venericardia ascia* is characteristic and locally abundant and is easily distinguished from the smaller *Venericardia potapacoensis* that is found in the underlying Potapaco Member (Ward, 1985). Other common taxa include *Cubitostrea* sp., *Pitar ovata*, *Macrocallista subimpressa*, *Corbula aldrichi*, *Lucina dartoni*, *Glycymeris* sp., *Turritella* sp., *Lunatia* sp., and *Cadulus* sp. The Woodstock beds also contain abundant benthic foraminifers and a sparse planktonic assemblage (Gibson and others, 1980). Well-preserved and diverse dinocyst floras are also present (Gibson and others, 1980; Edwards, 1989; Edwards and others, 1984; and this report). *Wetherellia*, a spherical fruit (nut) that is locally abundant in the Woodstock, is thought to be an indicator of a warm climate and a nearshore-marine depositional environment (Mazer and Tiffney, 1982).

Calvert Formation

Overlying the Pamunkey Group, but separated from it by a major regional unconformity, is the lower and middle Miocene Calvert Formation (Tc), a shelly, fine quartz sand and diatomaceous clayey silt and silty clay as much as 40 m (131 ft) thick. The Calvert Formation was named for Calvert County in southern Maryland, where the unit is extensively exposed in a series of high cliffs along the Chesapeake Bay (Shattuck, 1902, 1904). The Calvert Formation is widely distributed from Maryland and Delaware southward to southern Virginia (fig. 14). In North Carolina, equivalent strata are named the Pungo River Formation; in New Jersey, correlative beds are included in the Kirkwood Formation.

In its type area in Calvert County, Maryland, the Calvert Formation is subdivided, from bottom to top, into the Fairhaven, Plum Point Marl, and Calvert Beach Members, which are named for communities along the Chesapeake Bay shore in southern Anne Arundel and Calvert Counties. The Fairhaven and Plum Point Marl Members were named and described by Shattuck (1904). The Calvert Beach Member was defined and described by Gernant

(1970), who considered it to be the lowest member of the overlying Choptank Formation. More recently, however, Ward (1984a,b) demonstrated that the Calvert Beach beds are partly equivalent to Shattuck's (1904) "zones" 14–15 of the upper Calvert Formation and recommended that the Calvert Beach be retained as the uppermost member of the Calvert. Although the three members of the Calvert are similar lithically, they occupy different, though overlapping, depositional basins and are separated by unconformities of at least subregional extent. Ward (1992) has proposed a new molluscan biozonation of the Miocene beds in the middle Atlantic Coastal Plain and has described the molluscan assemblages that characterize each biozone.

Fairhaven beds.—In the Fredericksburg map area, beds equivalent to the lower Miocene Fairhaven Member of the Calvert have been recognized only in the Oak Grove core (Gibson and others, 1980) and at two outcrops on the northern side of the Rappahannock River about 2.5 km (1.6 mi) south of Rollins Fork (see map and localities 70 and 81 of Ward, 1985, p. 72, 74). In the Oak Grove core, diatomaceous strata of the Fairhaven are more than 5 m (16 ft) thick and contain the marker diatoms *Delphineis ovata*, *Sceptroneis caduceus*, and *Rhaphoneis scalaris* of the East Coast Diatom Zone 2 of Andrews (1988). The outcrops on the Rappahannock River expose as much as 7.5 m (24.6 ft) of the Fairhaven Member and its unconformable contact with the underlying Nanjemoy Formation. There, the lowermost 1.4 m (4.6 ft) of the Fairhaven, separated by a disconformity from the overlying Fairhaven beds, contains a sparse assemblage of siliceous microfossils including *Delphineis ovata* and the silicoflagellate *Naviculopsis navicula* (Barrett, 1977). These taxa suggest a late early or, possibly, early middle Miocene age for at least the lower part of the Fairhaven in the map area.

Plum Point beds.—Beds equivalent to the middle Miocene Plum Point Member are thicker in the eastern part of the map area where the unit includes two or more upward-fining sequences consisting mainly of fine quartz sand and diatomaceous clayey silt and silty clay. Thickness of the Plum Point is commonly 15 to 18 m (49–59 ft). A diverse and well-preserved diatom assemblage, including *Rhaphoneis magnapunctata*, *Rhaphoneis parilis*, *Delphineis penelliptica*, *Delphineis angustata*, *Delphineis novaecaesaraea*, and *Actinoptychus virginicus*, is indicative of East Coast Diatom Zone 3–4 of Andrews (1988) and suggests correlation with the upper part of the Plum Point Marl Member in its type area near Plum Point, Md. A section of similar thickness, but including beds equivalent to the lower part of the Plum Point Marl Member, was encountered in the Haynesville corehole on the lower Northern Neck of Virginia, about 30 km (19 mi) southeast of the map area (Mixon and others, 1989b). The Plum Point beds thin northwestward through the map area, extending to the Brooke fault zone which appears to mark the updip limit of outcrop of the unit. The thin (7 m (23 ft)) section of Plum Point beds, preserved on the relatively downthrown, eastern side of the Brooke fault zone, is truncated across the structure by the Calvert Beach beds, which directly overlie the Aquia Formation west of the fault zone (see cross section *A–A'*).

Calvert Beach beds.—The Calvert Beach beds, which commonly constitute the upper 9 to 15 m (30–49 ft) of the Calvert Formation in the map area, are similar lithically to the underlying Plum Point beds but have thicker and more extensive beds of sand (Mixon and others, 1989b). The basal one to two meters (3–6.5 ft) of the Calvert Beach is a poorly sorted sand commonly containing abundant small to medium pebbles of quartz and phosphate and bones of vertebrates including whale vertebrae. Rarely, some chalky shell material is preserved. The basal sand causes a very pronounced deflection of the single-point electrical resistivity curve and, based on available well log data, appears to be mappable in the subsurface through much of the Virginia Coastal Plain. The unconformable nature of the contact between the Calvert Beach and the underlying Plum Point beds is supported by the fairly abundant reworked planktonic and benthic foraminifers of Paleogene age found in the lowermost Calvert Beach—indicating erosion and wide dispersal of sediment from areas of Eocene and Paleocene outcrop in the Coastal Plain in early Calvert Beach time. From east to west across the inner Coastal Plain, the Calvert Beach strata lap over the Plum Point beds onto the Nanjemoy, Aquia, and Potomac Formations and crystalline rocks of the Piedmont (see cross sections *A–A'*, *D–D'*; Marr and Ward, 1987). These relations record a major transgression of the sea in the late middle Miocene.

The basal transgressive sand of the Calvert Beach grades abruptly upward into 3 to 8 m (10–26 ft) of diatomaceous clayey silt, silty clay, and clayey very fine sand. Commonly, in updip areas, dehydration and oxidation of the clays and silts brings out their laminated to very thin bedded character. The clayey and silty section is overlain, either gradationally or very sharply, by very thick bedded to massive, very fine to fine sand as much as 9 m (30 ft) thick. This sandy, upper part of the unit may represent a regressive phase of the Calvert Beach (see also Kidwell, 1988). West of the Brooke fault zone, the upper, sandy part of the Calvert Beach was largely removed by erosion both prior to and accompanying the maximum Yorktown transgression. The thin remnant of the Calvert Beach preserved in this structurally uplifted area, consisting mainly of the pebbly basal sand and a few meters of deeply weathered, red and gray mottled clay and silt, provided an easily recognized mapping horizon used during the Stafford fault system study (Mixon and Newell, 1977, 1978, 1982).

The well-preserved diatom assemblage in the Calvert Beach, including *Actinoptychus marylandicus*, *Coscinodiscus plicatus*, *Delphineis biseriata*, *Delphineis novaecaesaraea*, *Rhaphoneis amphiceros*, *Rhaphoneis clavata*, *Rhaphoneis gemmifera*, *Rhaphoneis lancettula*, and *Rhaphoneis scutula*, is typical of East Coast Diatom Zones 5–6 of Andrews (1988) and indicates correlation with the Calvert Beach Member (Shattuck's zones 14–15) in its type area in Maryland. Diatomaceous beds in the innermost Coastal Plain in Stafford and Spotsylvania Counties also contain *Rhaphoneis diamantella*, long considered to be a marker diatom for the Choptank Formation (East Coast Diatom Zone 7 of Andrews (1988)), which overlies the Calvert Formation in its type area in Calvert County, Maryland. In Virginia, the beds containing *Rhaphoneis diamantella* do not appear to be separable from the Calvert Beach beds and are here mapped with that unit. Abbott (1978) has also discussed the diatom zonation and the correlation of lower and middle Miocene strata in the inner Coastal Plain area of Virginia.

Eastover Formation

The upper Miocene Eastover Formation (Te) is a variably muddy, fine to very fine quartz sand and silty clay that is generally thick- to very thick bedded or massive and contains sparse to abundant molluscan shell material. Where fresh, the Eastover beds are dark gray to bluish gray and greenish gray; weathered outcrops are yellowish gray, greenish yellow, and yellowish brown. In the map area, the Eastover is as much as 15 m (49 ft) thick. The formation was named for the Eastover Plantation in Surry County in the south-central Virginia Coastal Plain (Ward and Blackwelder, 1980). Abundantly fossiliferous Eastover strata are well exposed at the type section, which is about 5 km (3 mi) east of Eastover in wave-cut cliffs on the southern side of the James River. The Eastover

Formation is widely distributed in the Virginia and Maryland Coastal Plain, extending from the subsurface of the Eastern Shore westward to the Virginia Fall Line. At the Fall Line, south of the map area, the Eastover laps over older Coastal Plain formations onto crystalline rocks of the Piedmont and continental red beds of the early Mesozoic Taylorsville basin. In Virginia, the Eastover Formation has been divided into a lower member, the Claremont Manor, consisting of slightly to moderately shelly, poorly sorted, clayey and silty fine sand and silty clay, and an upper member, the Cobham Bay, which is relatively well sorted, very shelly fine sand (Ward and Blackwelder, 1980; Ward, 1984b). The type sections of the Claremont Manor and Cobham Bay Members are also along the James River. The Eastover beds of the Fredericksburg map area (see also Weems and others, 1996) appear to be roughly equivalent to the Claremont Manor Member.

In the map area, the best reference sections for the Eastover Formation are cores obtained from the Wilcox Camp corehole in the Bowling Green 7.5-min quadrangle about 1.2 km (0.75 mi) northwest of Delos, Va., and the Loretto corehole in the Loretto 7.5-min quadrangle about 1.6 km (1.0 mi) southeast of Hustle, Va. The Eastover core section at Wilcox Camp on the Fort A.P. Hill Military Reservation consists of two lithic units separated by a pebbly, deeply burrowed surface of disconformity, which is thought to be a ravinement. The lower unit, which is about 2 m (7.0 ft) thick, is very clayey and silty, medium- to dark-gray fine sand. Abundant *Turritella plebia* and the large bivalves *Placopecten principoides* and *Cyrtopleura* sp. were collected from temporary exposures of the lower unit in the wall of a settling pond for a nearby sewage treatment plant. The muddy sand of the lower unit, the low-diversity molluscan assemblage, and the articulated valves of *Cyrtopleura* (commonly called angel wing) suggest a restricted marine depositional environment such as a bay or lagoon. The upper unit, which is about 6 m (20 ft) thick in the core, consists of better sorted, less muddy, fine to very fine sand. The settling pond exposure of the upper unit contained well-preserved *Chesapecten middlesexensis*, *Mercenaria* sp., *Ostrea "compressirostra"*, and *Turritella plebia* that suggests a somewhat less restricted, shallow-shelf environment of deposition. At this locality the Eastover Formation unconformably overlies the Calvert Formation and, in turn, is overlain unconformably by the Yorktown Formation of Pliocene age.

In downdip parts of the map area (northwestern Essex and eastern Caroline Counties), where the Eastover section is thicker and more typically developed, beds of medium-grained, crossbedded sand (sand bars) occur in the upper part of the formation, suggesting deposition in upward-shoaling waters. In central Essex County, just southeast of the map area, the upper Eastover includes a very distinctive 5- to 6-m (16- to 20-ft)-thick unit consisting of alternating, thin to very thin beds of sand and clay. Bedding types include wavy bedding and lenticular bedding exhibiting both connected and isolated sand lenses. In the Fredericksburg map area, these tidally influenced deposits and the overlying Eastover section are missing because of truncation by the transgressing Yorktown sea. In a similar manner, the absence of the Eastover in the area between the Rappahannock and Potomac Rivers is the result of the differing alignments of the Pliocene and late Miocene depositional basins in northern Virginia. Because of the superposition of the north-trending Pliocene basin (see fig. 3) on the northeast-trending late Miocene basin of the Eastover (R.B. Mixon, unpub. data), the truncation of the Eastover by younger deposits was greater in the area to the northeast of the Rappahannock River.

Abundantly fossiliferous Eastover beds crop out at four localities in the southeastern part of the map area, including one locality in each of the Rappahannock Academy, Woodford, Bowling Green, and Loretto 7.5-min quadrangles (see map). A composite list of mollusks includes

Dallarca virginiae (Dall)
Dallarca sp.
Chesapecten middlesexensis (Mansfield)
Placopecten principoides (Emmons)
Ostrea "compressirostra"
Anadara sp.
"Spisula" rappahannockensis (Gardner)
Dosinia acetabulum (Conrad)
Mercenaria sp.
Nucula sp.
Corbula sp.
Cyrtopleura sp.
Turritella plebia Say
Ecphora sp.

The small bivalve, *"Spisula" rappahannockensis*, and the large, strongly ribbed pecten, *Chesapecten middlesexensis*, are characteristic species of the Eastover Formation (Ward and Blackwelder, 1975, 1980).

The late Miocene age of the Eastover Formation is based on benthic and planktonic foraminiferal data from outside the map area. Gibson (1983) and P.F. Huddlestun (oral commun., 1986) have placed the Eastover beds in Virginia in the uppermost part of foraminiferal zone N16, and in N17 and N18 which are of late Tortonian, Messinian, and early Zanclian ages.

Yorktown Formation and Unit Tps

The Yorktown Formation was named by Clark and Miller (1906) for exposures of shelly sand, silt, and clay in cliffs along the York River near Yorktown in southeastern Virginia. There, the formation is 18 m (59 ft) thick and contains fossil assemblages indicating a shallow-shelf depositional environment. On the basis of well-exposed sections along the James, York, and Rappahannock Rivers in southeastern and central Virginia, Ward and Blackwelder (1980) revised the Yorktown Formation and divided it into four members: a basal Sunken Meadow Member, and the overlying Rushmere, Morgarts Beach, and Moore House Members. Differences in lithology, fossil content, and geographic distribution of these members suggest deposition in three transgressive pulses. The Sunken Meadow Member, representing an early Pliocene transgression, is thin or absent in updip areas of the northern Virginia Coastal Plain. In central and southern Virginia the Rushmere and Morgarts Beach Members, which represent the second (maximum) transgression during Yorktown deposition (Ward and Blackwelder, 1980; Ward and Strickland, 1985), extend westward to the Fall Line and, at least locally, lap onto crystalline rocks of the outermost Piedmont. The Moore House Member, representing a possible third transgressive pulse of minor extent, overlies the Rushmere and Morgarts Beach Members but appears to be restricted to the outer Coastal Plain of southeastern Virginia and northern North Carolina. Except for the Rushmere, the members have not been recognized in updip areas of northern Virginia where the Yorktown consists of poorly fossiliferous to nonfossiliferous, fine to coarse sand and sandy gravel deposited in various marine and marginal-marine environments. In northern Virginia and in southern Maryland west of the Chesapeake Bay, some workers have extended the original definition of the Yorktown Formation to include the marginal-marine and (or) fluvial-deltaic equivalents of the fossiliferous shallow-shelf deposits of central and southeastern Virginia (see Stephenson and MacNeil, 1954; Newell and Rader, 1982; Johnson and others, 1987; Powars, 1987; Pavich and others, 1989, fig. 18; McCartan and others, 1995). In this report, however, the fluvial-deltaic equivalents of the Yorktown are mapped separately as Unit Tps (see geologic map). The Pliocene age of the Yorktown Formation is based on studies of planktonic foraminifers, ostracodes, and nannofossils from shelf deposits in southeastern Virginia and North Carolina (Akers, 1972; Gibson, 1983; Hazel, 1983; Cronin and others, 1984). These workers placed the Yorktown in foraminiferal zones N18, N19, and N20 of early and early late Pliocene age. The most recent detailed study of the Virginia section (Dowsett and Wiggs, 1992) assigns the Yorktown to zones N19–N20.

Stratigraphic and geomorphic relations

In the map area, the transgressive deposits of the Yorktown Formation (Ty) and the overlying, regressive sands and gravels of Unit Tps are a thick sheet of fine to coarse clastic sediment that caps the uplands of the inner Coastal Plain (Mixon and others, 1994). From east to west across the map area, the Yorktown unconformably overlies the Eastover Formation, the Calvert Formation, Bon Air gravel equivalents (unit Tms), and crystalline rocks of the Piedmont. The thickness of the Yorktown Formation ranges from about 25 m (82 ft) in the Supply and Loretto 7.5-min quadrangles, where the Yorktown constitutes most of the Pliocene section, to a feather edge at the north-trending Thornburg scarp near the western margin of the Coastal Plain (see fig. 3 and Mixon, 1978). This coastwise scarp is believed to be a wave-cut paleoshoreline formed during a high stand of the sea associated with the maximum transgression during Yorktown deposition (Ward and Blackwelder, 1980; Dowsett and Cronin, 1990). Thus, the Thornburg scarp is the western, updip limit of both the marginal-marine strata of the Yorktown Formation and the delta-plain deposits of Unit Tps (see geologic map). Along the Broad Rock scarp in the eastern part of the map area (fig. 3), the beds of the upper Yorktown are truncated by the Bacons Castle Formation of latest Pliocene age. East of the scarp, the Bacons Castle unconformably overlies the lower part of the Yorktown Formation.

Lithology

In the map area, the Yorktown Formation consists of thin- to very thick bedded, fine to coarse quartz and feldspar sand and sandy gravel; minor amounts of clay and silt occur as thin to thick beds interbedded with the coarser materials. Pebbles and cobbles are dominantly vein quartz, quartzite, and sandstone. In the downdip direction (eastern and southeastern parts of the map area), sand of the Yorktown becomes more quartzose and is overall finer grained and better sorted. In these areas, the basal beds of the Yorktown contain some glauconite, sand- to pebble-sized phosphate and, locally, shelly material. Because the Yorktown Formation consists of highly permeable materials, and because it and the overlying Unit Tps constitute the higher interfluve areas, the formation is commonly deeply oxidized to yellowish gray, yellowish orange, and reddish brown. Where fresh, the strata are light to dark gray or bluish gray.

Over much of the area, the base of the Yorktown Formation is marked by a pebbly lag ranging in thickness from less than 0.3 m (1.0 ft) to as much as 4.6 m (15 ft) in lows on the pre-Yorktown erosion surface (see fig. 3). These basal lag deposits commonly grade upward into fine to coarse, massive to crossbedded sand characterized by sparse to very abundant *Ophiomorpha nodosa*, a clay-lined, noded burrow commonly filled with sand. The burrows are thought to be made by a marine crustacean similar to the present-day mud shrimp, *Callianassa major*. The shrimp occurs in various marine environments—including barriers, backbarriers, and the nearshore sublittoral zone—wherever salinities and current energy are moderately high (Frey and others, 1978). In updip areas, crossbedded, pebbly sands containing *Ophiomorpha* commonly grade upward, or laterally, into alternating beds of fine sand, silt, and clay that exhibit flaser and lenticular bedding. These beds, believed to represent shoreface and intertidal deposits, form a distinctive 2- to 5-m (6.5- to 16-ft)-thick unit that can be mapped throughout much of the map area and adjacent parts of the northern Virginia and Maryland Coastal Plain.

Bowling Green paleovalley and fluvial-deltaic complex

Structure contours at the base of the Yorktown Formation in parts of six counties in northern Virginia (fig. 3) show an irregular surface that dips generally eastward from altitudes of about 82 m (269 ft) at the toe of the Thornburg scarp to about 30 m (98 ft) near Tappahannock, Va., a few kilometers southeast of the Fredericksburg map area. The average dip of the erosion surface over this distance is 0.8 m/km (4 ft/mi). In the western part of the Coastal Plain area, the contours delineate a shallow, sinuous, 6-km (3.7-mi)-wide paleotopographic low incised about 10 to 15 m (33 to 49 ft) below the general level of the pre-Yorktown erosion surface. This feature, which intersects the present-day Rappahannock River valley at the edge of the Piedmont near Embrey Hill just west of Fredericksburg, is believed to be a Rappahannock River paleovalley of early(?) Pliocene age. From Embrey Hill, the paleovalley trends southeastward through the Guinea 7.5-min quadrangle to the vicinity of Bowling Green (fig. 3). Just northwest of Bowling Green, the paleovalley veers abruptly to the northeast before resuming a southeasterly course through the eastern Bowling Green and Supply 7.5-min quadrangle areas. The northeastward deflection of the paleovalley mimics the deflection of the present-day Rappahannock River at Skinkers Neck, suggesting that the abrupt change in direction of both the modern and ancient river courses may be influenced by the Skinkers Neck anticline (fig. 5).

Outcrops along the upper part of the paleovalley in the Salem Church, Spotsylvania, and Guinea 7.5-min quadrangle areas show that the valley fill is mainly medium to coarse gravelly sand and pebble and cobble gravel of fluvial origin. Some of the better exposures of fluvial deposits are in the high cuts for the Richmond, Fredericksburg, and Potomac Railroad about 1 mi north of Summit (see Guinea quadrangle) where the railroad crosses the area of the paleovalley fill. There, the railroad cuts show large-scale trough crossbedding in coarse sand and gravels that fill a series of channels up to 20 m (65 ft) or more in width. The fluvial channel fills appear to grade upward and laterally into a large fluvial-deltaic complex that constitutes the surficial deposits of the inner Coastal Plain in a multicounty area east and south of Fredericksburg. The coarse fluvial fill of the Bowling Green paleovalley and the interfingering and overlying deposits of the delta-plain complex constitute Unit Tps (Pliocene sand and gravel).

Bacons Castle Formation

Sand, gravel, silt, and clay of the uppermost Pliocene Bacons Castle Formation constitute the surficial deposits of the extensive coastwise terrace between the highly dissected Broad Rock scarp to the west and the Surry scarp to the east (Oaks and Coch, 1973; Johnson and Peebles, 1984). The formation was named by Coch (1965, 1968) for the community of Bacons Castle in Surry County, southeastern Virginia. Subsequently, Oaks and Coch (1973, p. 119) designated type sections in Surry County for both the sand and the silt facies of the Bacons Castle. Much later, after detailed study and mapping of the Bacons Castle in the Coastal Plain area east of Richmond, Va., the formation was subdivided into the updip, relatively coarse-grained Varina Grove Member and the downdip, finer grained Barhamsville Member (Johnson and Ramsey, 1987; Ramsey, 1988).

Table 1.—Principal structures of the Stafford fault system, inner Coastal Plain of Virginia.
[Numbers keyed to map]

Name and map designation	Extent and location	Best exposure	Rock units affected
1 Dumfries fault zone	40.5 km (25.2 mi) through parts of Stafford, Prince William, and Fairfax Counties, Virginia	Trench about 3 km (1.9 mi) west-northwest of Stafford, Va.	Ordovician Quantico Formation is thrust at high angles over Lower Cretaceous Potomac Formation. Fault is updip limit of the upper Paleocene Aquia Formation.
2 Fall Hill fault	20.1 km (12.5 mi) from Fall Hill in the City of Fredericksburg to Aquia Creek northeast of Stafford, Va.	Trenches along Virginia Power electrical transmission line in northern Falmouth, Va. Formerly well exposed on Fall Hill Avenue in Fredericksburg, Va. Now obscured by road improvement.	High-angle reverse and vertical faults juxtapose early Paleozoic and (or) Late Proterozoic gneiss and schist of the Po River Metamorphic Suite and sand and clay of the Potomac Formation. In downthrown fault block at Falmouth study site, high-angle reverse faults and a low-angle east-dipping thrust having minor offset displace beds of both the Potomac Formation and the upper Paleocene Aquia Formation.
3 Hazel Run fault	16.0 km (9.95 mi) from Fredericksburg, Va., southwestward to Fraziers Gate near the Ni River, east-central Spotsylvania County, Virginia.	Hazel Run and tributary ravine about 150 and 90 m (492 and 295 ft), respectively, west of Mary Washington Wayside roadside park on U.S. Highway 1 Bypass.	Schist and biotite augen gneiss of Po River Metamorphic Suite are thrust at a high-angle over the Lower Cretaceous Potomac Formation. Surface mapping and a structure contour map on base of the Aquia Formation indicate as much as 18.5 m (60.7 ft) of fault displacement of the Aquia.
4 Brooke fault zone	About 40 km (25 mi) from near Massaponax in Spotsylvania County, Virginia, northeastward to the Quantico Marine Base in eastern Prince William County, Virginia.	Tank Creek fault, a main strand of Brooke fault zone, crops out in wave-cut cliffs on the Potomac River at northern edge of map about 365 m (1197 ft) north of mouth of Tank Creek.	The arcuate, high-angle Tank Creek fault of uncertain classification and an intersecting high-angle reverse fault, which occur within a thick section of the Potomac Formation, indicate down-to-the-coast displacement of the Potomac strata and a strike-slip component of movement. Water well data on Quantico Marine Base adjacent to north of map suggest 50 to 60 m (164–197 ft) of down-to-the-coast displacement of the crystalline basement rock surface across the Brooke fault zone.

The Varina Grove Member is an upward-fining fluvial-estuarine sequence, as much as 22 m (72 ft) thick, that consists of sandy gravel, pebbly sand, and sandy mud. The outcrop and subsurface sections at the West Sand and Gravel Company pit, which is about 1.5 km (0.9 mi) southwest of Varina Grove in Henrico County, Virginia, show typical lithologies and considerable vertical and lateral variation of the unit. In the downdip direction, the Varina Grove interfingers with the Barhamsville Member, which consists of as much as 17 m (56 ft) of interbedded, fine to coarse sand, silt, and clay characterized by flaser, lenticular, and wavy bedding. The Barhamsville is believed to have been deposited as a tidal flat and tidal channel complex (Johnson and Ramsey, 1987). The type section of the Barhamsville Member is exposed in roadcuts along Virginia Route 632 about 2 km (1.2 mi) west of the community of Barhamsville, New Kent County, Virginia.

In the map area, the Bacons Castle (Tb) occurs mainly as high-level, fluvial terrace deposits along major drainages such as the Potomac, Rappahannock, and Mattaponi Rivers. These deposits are inset below the level of the adjacent, highly dissected upland areas underlain by gravelly sands of Unit Tps and the Yorktown Formation. Here, the Bacons Castle commonly overlies the lower Yorktown or diatomaceous strata of the Calvert Formation. Along the Rappahannock River, in the relatively uplifted, innermost part of the Coastal Plain, fluvial beds of the Bacons Castle successively overlie, from east to west, strata of the Nanjemoy, Aquia, and Potomac Formations. Locally, the Bacons Castle directly overlies crystalline rocks of the Piedmont. The Bacons Castle is very poorly exposed in the map area; thus, the two members of the formation have not been differentiated. However, the Bacons Castle deposits mapped along the main rivers are upward-fining sequences that are very similar lithically to the Varina Grove Member as described above.

COASTAL PLAIN STRUCTURES

Stafford Fault System

The Stafford fault system consists of en echelon, northeast-striking, high-angle reverse faults extending from the vicinity of Spotsylvania and Massaponax in northern Spotsylvania County, Virginia, northeastward along the Fall Line for 68 km (42 mi) to Newington in southern Fairfax County, Virginia (see geologic map, table 1, and Washington West 30′ x 60′ quadrangle, adjacent to the north). The dominantly compressional faulting involves both the ancient metamorphic rocks of the easternmost Piedmont and the overlying, much younger sedimentary formations of the Virginia Coastal Plain (Mixon and Newell, 1977, 1978, 1982). The few outcrops of the faults commonly show steeply northwest-dipping to vertical fault planes along which the Piedmont metamorphic rocks are thrust at high angles over unconsolidated Lower Cretaceous sands of the Potomac Formation. Although vertical displacement of beds by individual faults is moderate (10–60 m (33–197 ft)), the Stafford faults appreciably affect the present distribution and thickness of Coastal Plain formations. For example, areas along the relatively downthrown, southeastern side of faults preserve thicker and more complete stratigraphic sections (fig. 2). In relatively upthrown areas along the western side of the faults, the Coastal Plain formations are commonly abruptly thinner or absent because of partial or complete truncation by younger stratigraphic units.

The Stafford fault system includes four discrete mappable faults or zones of faults: the Dumfries fault zone, the Fall Hill fault, the Hazel Run fault, and the Brooke fault zone (see map and fig. 2). These faults were recognized initially during the mapping of the Coastal Plain in Prince William and Stafford Counties, Virginia, and in the city of Fredericksburg (Mixon and others, 1972; Mixon and Newell, 1977, 1978). The generally poor exposures of the faults and the subtle differences between Coastal Plain formations involved in the faulting make the Stafford faults difficult to find and map. Recently, intensive development in some parts of the map area has completely obscured any surface expression of the faults. As knowledge of the nature and timing of faulting is necessary to evaluate earthquake hazards and understand the movement of ground water and possible subsurface contaminants, the locations and descriptions of the main fault structures are summarized in the following text.

Dumfries fault zone.—The Dumfries fault zone, the northwesternmost structure of the Stafford fault system, has been mapped along the Fall Line from Accotink Creek in Fairfax County, Virginia, southwestward for about 45 km (28 mi) to the vicinity of Abel Lake Reservoir southwest of Stafford in Stafford County, Virginia. The fault zone was first delineated near the town of Dumfries just north of the Fredericksburg map area. There, structure contours drawn on the top of the Piedmont crystalline rocks show an oversteepened southeast-dipping gradient as a pronounced contour lineament trending southwestward through Dumfries, Triangle, and Boswells Store (see geologic map of the Quantico quadrangle by Mixon and others, 1972). The contour lineament defines both the southeasternmost outcrops of Piedmont crystalline rocks and the updip limit of the thicker (>90 m (295 ft)) part of the superjacent Potomac Formation. Geomorphically, the structure-contour lineament marks the point where wide, flat-bottomed valleys, characterized by tidal creeks and marshes, narrow abruptly to V-shaped headwaters partly incised into the Piedmont crystalline rocks. These stratigraphic, structural, and geomorphic relations in the Dumfries area strongly suggested faulting of the crystalline basement rock and the overlying Coastal Plain strata and led to recognition of the Stafford fault zone.

The Dumfries structure is more conspicuous and more easily studied in the Fredericksburg map area northwest of Stafford where differential erosion of the Coastal Plain strata in fault contact with more resistant Piedmont metamorphic rocks has formed an 8-km (5-mi)-long fault-line scarp (see this map and the Stafford 7.5-min quadrangle). A trench (fig. 15*A*, *B*) excavated across a fault contact on the fault-line scarp shows that the main structure is a northwest-dipping, high-angle reverse fault thrusting slate, phyllite, and schist of the Ordovician Quantico Formation over feldspathic sand of the Lower Cretaceous Potomac Formation (Newell and others, 1976). The main reverse fault in the trench strikes N. 50° E. and dips 68° NW. As is commonly the case with strike measurements of reverse faults of the Stafford system, the strike of this fault is at a considerable angle to the more northerly N. 35° E. trend of the structure-contour lineament delineating the Dumfries fault zone. Thus, we infer that the Dumfries fault zone is a series of en echelon and interconnecting faults that strike more easterly than the main structure—a pattern suggesting a dextral strike-slip component of movement. Although displacement on the fault is dominantly dip-slip, slickenside measurements also support a dextral component. Vertical separation of the unconformable contact between the Potomac Formation and the underlying Quantico Formation is about 35 m (115 ft), as indicated by detailed mapping and borehole data.

In the innermost Coastal Plain, the trace of the Dumfries fault zone follows closely the outcrop-subcrop belt of the relatively incompetent slates and phyllites of the Quantico Formation. The structure crosses at a high angle the southeast-trending pre-Cretaceous and Cretaceous paleodrainage developed on the top of the underlying crystalline rock terrane. The doubling in thickness of the Potomac Formation across the fault zone and the more steeply dipping, coarse, poorly sorted, feldspathic Potomac sands in the vicinity of the structure suggest that in this area part of the Potomac was deposited as broad, gently sloping, coalescing alluvial fans adjacent on the east of a low fault scarp marking the edge of the Piedmont upland.

In the map area and in the Washington West map area adjacent to the north (Lyttle and others, in press), the Dumfries fault zone is the present-day updip limit of outcrop of the upper Paleocene Aquia Formation (see fig. 2; cross section *A–A′* and map). From east to west across the fault zone, sandy and silty Miocene strata of the Calvert(?) Formation truncate the Aquia beds and, on the western, upthrown side, directly overlie either Lower Cretaceous sand of the Potomac Formation or crystalline rocks of the Piedmont. The abrupt truncation of the Aquia beds along the Dumfries fault zone and the reversal of regional dip (normally to the southeast) of the Aquia-Potomac unconformity near the structure indicate faulting of the Aquia prior to deposition of the Miocene beds. Whether or not the Miocene and younger strata are also faulted has not been determined as yet.

Fall Hill fault.—The Fall Hill structure is a narrow zone of dominantly high-angle reverse faults mapped from the vicinity of Coal Landing on Aquia Creek, northeast of Stafford, Va., southwestward for about 18 km (11 mi) to the edge of the Piedmont in Spotsylvania County, adjacent on the western side of the city of Fredericksburg. The fault zone crosses the Rappahannock River just upstream from Laucks Island, near the upper end of the long series of rapids that constitute the Fall Line on the Rappahannock. For much of its length, the fault trace is marked by a topographic lineament consisting of squared-off, east-facing ridge spurs, aligned ravines, and abrupt changes in the courses of small streams. The fault is named for exposures on Fall Hill, a 76-m (250-ft)-high ridge of Piedmont crystalline rock, capped by thin Coastal Plain deposits. Fall Hill is situated at the edge of the Piedmont upland at Fredericksburg, near the upper end of the wide alluvial valley that characterizes the Coastal Plain part of the Rappahannock River.

The more accessible outcrops of the Fall Hill fault are in low roadcuts along Virginia Route 639 in west Fredericksburg, where the highway ascends the eastern slopes of Fall Hill (fig. 16). At the break-in-slope near the top of the hill, at an altitude of about 60 m (197 ft), coarse-grained biotite gneiss of the Po River Metamorphic Suite is thrust at a high angle over sand of the Lower Cretaceous Potomac Formation. The main fault exposed in the roadcut strikes N. 39° E. and dips about 78° NW. A clay gouge as much as 15 cm (6 in) thick is present along the fault plane. Vertical separation of the crystalline rock-Potomac Formation contact is at least 35 m (115 ft).

The base of a thin gravelly sand unit, which unconformably overlies the Potomac Formation and the Po River gneiss at this locality, is offset a minimum of 28 cm (11 in) by the main reverse fault. The lithology and the topographic and stratigraphic position of the gravelly sand unit, which contains cobbles and boulders of crystalline rocks, suggest that it is a high-level Rappahannock River terrace deposit equivalent, in part, to the Pliocene Yorktown Formation.

At the Falls Run locality in northwest Falmouth, Va., trench studies of the Fall Hill fault, conducted by Dames and Moore for the Potomac Electric Power Company (PEPCO) of Maryland, exposed a 9-m (30-ft)-wide shear zone in the Potomac Formation that includes both vertical faults and steeply northwest-dipping reverse faults (figs. 17, 18). In this area, the strike of the main Fall Hill structure is about N. 35° E., but the strike of individual faults in the narrow zone of faulting exposed by the trenching ranges from north to N. 57° E. (Potomac Electric Power Company, 1976, pl. 9). At the southeastern end of PEPCO's "trench number 1", the Aquia Formation unconformably overlies the more steeply dipping Potomac strata. Here, some of the subsidiary high-angle reverse faults and a low-angle, southeast-dipping thrust showing minor offset displace beds of both the Potomac Formation and the upper Paleocene Aquia Formation—clearly indicating that the faulting affected Tertiary strata. Toward the northwestern end of the trench, the base of the capping Pliocene sandy gravel (Unit Tps) is offset about 2.5 cm (1 in) by one of the main faults (fig. 18*B*). Borehole data for the PEPCO study indicate that, across the fault, vertical separation of the Potomac Formation-crystalline rock unconformity is about 36 m (118 ft).

Hazel Run fault.—The Hazel Run structure is a linear, high-angle reverse fault mapped from the vicinity of Mary Washington Wayside, a roadside park on U.S. Highway 1 Bypass in south Fredericksburg, southwestward for about 12 km (7.5 mi) through the communities of Leavells and Fraziers Gate to the Vulcan Materials quarry on the Ni River (see map). The structure is named for natural exposures at two sites along Hazel Run, a short distance upstream from the U.S. Highway 1 Bypass bridge and west of Mary Washington Wayside Park (fig. 4). The west-dipping fault plane of the main reverse fault crops out in the steep streambank on the south side of Hazel Run about 150 m (492 ft) west-southwest of the highway bridge. At this locality, biotite augen gneiss of the Po River Metamorphic Suite is thrust at a high angle over greenish-gray sand of the Potomac Formation. A borehole on the relatively downthrown side of the fault, sited on the abandoned Civil War era railroad south of Hazel Run (fig. 4), demonstrates about 37 m (121 ft) of vertical separation of the Potomac Formation-crystalline rock contact across the fault. Vertical separation of the base of the Aquia Formation by the fault is only about half as much (18.5 m (60.7 ft)), as indicated by mapping and contouring the basal Aquia unconformity in the immediate area. The best exposures of the Hazel Run fault and related subsidiary faults in the Potomac Formation occur near the mouth of a narrow ravine on the northern side of Hazel Run, about 90 m (295 ft) west of the picnic tables at Mary Washington Wayside Park (figs. 4, 19).

Brooke fault zone.—The Brooke fault zone was originally mapped in reconnaissance as the Brooke monocline (Mixon and Newell, 1977). Subsequently, detailed mapping disclosed a 1.5-km (1-mi)-wide zone of en echelon faults and flexures extending from the vicinity of Massaponax in Spotsylvania County northeastward for about 40 km (25 mi) to the Quantico Marine Corps Base in eastern Prince William County, Virginia (see map and Washington West 30′ x 60′ quadrangle adjacent to the north) (Lyttle and others, in press). At Quantico, the Brooke fault zone projects northeastward beneath the upper Potomac River estuary toward Alexandria, Va., and Washington, D.C. The position of this northeasterly reach of the Potomac River is thought to be controlled by the relative down-to-the-coast displacement of the Coastal Plain strata along the Brooke fault zone and other structures (Seiders and Mixon, 1981, fig. 1) along the northern part of the Stafford fault system.

The Brooke fault zone was defined and mapped by structure contouring control points on the top of the Lower Cretaceous Potomac Formation (which is equivalent to contouring the base of the directly overlying Aquia Formation of Paleocene age). Most altitudinal data were obtained by altimeter surveys and recorded on 1:24,000-scale topographic maps. Major faults are evident as well-defined contour lineaments on structure-contour maps. The maps indicate that maximum displacement of the top of the Potomac Formation by an individual fault is generally 15 to 18 m (49–59 ft). Displacement of the Potomac-Aquia contact along a northwest-trending transect of the entire Brooke fault zone is as much as 30 m (98 ft).

Although fault displacements are small, these structures appreciably affect the distribution and thickness of Coastal Plain formations. For example, between Aquia and Potomac Creeks, the southeasternmost fault of the Brooke fault zone (see map) marks the present-day updip limit of the Marlboro Clay and Nanjemoy Formations (fig. 2). On the southeastern side of the fault, the middle Miocene Calvert Formation unconformably overlies the lower Eocene Nanjemoy Formation. On the northwestern, upthrown side of the fault, the Miocene beds directly overlie the upper Paleocene Aquia Formation. Thus, the Marlboro and the Nanjemoy beds were eroded from the upthrown side of the fault during middle Eocene to early Miocene time. These relations indicate a middle Tertiary episode of deformation in this region.

An excellent natural exposure (fig. 20) of one of the main strands of the Brooke fault zone is found in low, wave-cut cliffs bordering the Potomac River near the northern edge of the map area and about 365 m (1198 ft) north of the mouth of Tank Creek (see map and Widewater 7.5-min quadrangle). The section exposed in the cliffs consists of the Potomac Formation and thin Potomac River terrace deposits of late Pleistocene age (Sedgefield Member of Tabb Formation). The Tank Creek structure includes a high-angle fault of uncertain classification and an intersecting, small, high-angle reverse fault showing relative up-to-the-Piedmont displacement of the Potomac Formation beds (Mixon and Newell, 1982, fig. 15). The subsidiary reverse fault strikes N. 35° E., parallel to the main Stafford faults, and dips 56° NW. A strike-slip component of movement is indicated by the marked difference in thickness of a crossbedded sand bed directly across the plane of the reverse fault. The exposed part of the larger fault has an arcuate fault plane that is convex to the northwest and strikes N. 66° E. at a considerable angle to the Brooke structure-contour lineament. Vertical separation of Potomac Formation beds by the larger fault can only be estimated as greater than 3 m (10 ft), the height of the cliff exposure. However, lithic logs of water wells at Quantico, Va. (Mixon and others, 1972), indicate as much as 50 to 60 m (164–197 ft) of up-to-the-Piedmont displacement of the crystalline basement rock surface across the Brooke fault zone.

Skinkers Neck Anticline

The Skinkers Neck anticline is a broad, low-amplitude, gently northeast-plunging anticline that extends from the edge of the Coastal Plain near Bowling Green, Va., north-northeastward for more than 40 km (25 mi) to the vicinity of Fairview Beach on the Potomac River (Mixon and Powars, 1984; Mixon and others, 1992). The form and extent of the fold is defined by structure-contour and isopach maps of Cretaceous and Tertiary formations (see geologic map and figs. 5, 6). The axis of the anticline, which crosses the Rappahannock River at Skinkers Neck, closely parallels the faulted western edge of the early Mesozoic Taylorsville rift basin present in the subsurface beneath the thick Coastal Plain deposits. In a like manner, but farther to the northeast in Charles and Prince Georges Counties, Maryland, the western edge of the Taylorsville basin is paralleled by the Brandywine fault zone and associated folds (Jacobeen, 1972; Potomac Electric Power Company, 1973). Additionally, cataclastic rocks of the Hylas fault, which marks a major zone of late Paleozoic ductile shearing and thrusting along the western sides of the Richmond and Taylorsville basins in the Virginia Piedmont, are believed to project northeastward into the map area along the western side of the buried part of the Taylorsville basin. The superposition of late Mesozoic and Cenozoic folds and faults of the Coastal Plain above early Mesozoic and Paleozoic fault zones in the deep subsurface indicates continued release of compressional stress along the Hylas zone and (or) reactivation and reversal of movement on the extensional faults of the early Mesozoic rift basins (Mixon and Newell, 1977; Mixon and Powars, 1984; Mixon and others, 1992).

Port Royal Fault Zone

A few miles to the east and downdip of the Skinkers Neck anticline, structure contours on the top of the Lower Cretaceous Potomac Formation delineate a 3-km (2-mi)-wide zone of downwarping of the Coastal Plain strata that extends from central Caroline County, Virginia, northeastward through the town of Port Royal on the Rappahannock River and Mathias Point Neck on the Potomac River (fig. 5). The zone of deformation, named the Port Royal fault zone (Mixon and Powars, 1984; Mixon and others, 1992) is marked by dominantly down-to-the-coast displacement and thickening of the Coastal Plain sedimentary section; thus, the sense of movement is the opposite of that shown by the Skinkers Neck anticline. Seismic reflection surveys across the Port Royal fault zone indicate that the Coastal Plain structure parallels the subsurface trace of one or more major listric faults within the subjacent Taylorsville rift basin. These intrabasinal faults appear to separate relatively thin synrift deposits (red beds) to the west from substantially thicker synrift deposits to the east—mirroring the fairly abrupt eastward thickening of the overlying Coastal Plain beds.

Near the town of Port Royal and to the northeast, where some borehole data are available, the Port Royal fault zone is tentatively interpreted as a shallow graben-like structure having greater structural relief along its northwestern side (fig. 5). There, vertical displacement of the top of the Cretaceous strata is about 15 m (49 ft), or more, and is thought to have occurred mainly by faulting. Displacement of the top of the Paleocene section (Aquia Formation and Marlboro Clay) is as much as 10 m (33 ft). The overlying Eocene and Miocene strata also are deformed, but much of the deformation may have occurred by flexuring rather than faulting.

The southeastern side of the Port Royal structure is poorly defined but at least some up-to-the-coast displacement of the Cretaceous strata is suggested by water well data which include both lithic and geophysical logs. A structural high along the southeastern side of the Port Royal fault zone also is suggested by seismic reflection data and the relatively high post-rift unconformity (top of Triassic red beds) encountered by the Texaco Butler No. 1 corehole (Milici and others, 1991).

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