2.5 Geology, Seismology, and Geotechnical Engineering

2.5 GEOLOGY, SEISMOLOGY, AND GEOTECHNICAL ENGINEERING

This section presents information on the geological, seismological, and geotechnical engineering properties of the {CCNPP} site. Section 2.5.1 describes basic geological and seismologic data, {focusing on those data developed since the publication of the Final Safety Analysis Report (FSAR) for licensing CCNPP Units 1 and 2}. Section 2.5.2 describes the vibratory ground motion at the site, including an updated seismicity catalog, description of seismic sources, and development of the Safe Shutdown Earthquake and Operating Basis Earthquake ground motions. Section 2.5.3 describes the potential for surface faulting in the site area, and Section 2.5.4 and Section 2.5.5 describe the stability of surface materials at the {CCNPP} site.

Appendix D of Regulatory Guide 1.165, "Geological, Seismological and Geophysical Investigations to Characterize Seismic Sources," (NRC, 1997) provides guidance for the recommended level of investigation at different distances from a proposed site for a nuclear facility.

- The site region is that area within 200 mi (322 km) of the site location (Figure 2.5.1-1).
- The site vicinity is that area within 25 mi (40 km) of the site location (Figure 2.5.1-2).
- The site area is that area within 5 mi (8 km) of the site location (Figure 2.5.1-3).
- The site is that area within 0.6 mi (1 km) of the site location (Figure 2.5.1-4).

These terms, site region, site vicinity, site area, and site, are used in Sections 2.5.1 through 2.5.3 to describe these specific areas of investigation. These terms are not applicable to other sections of the FSAR.

The geological and seismological information presented in this section was developed from a review of previous reports prepared for the existing units, published geologic literature, interpretation of aerial photography, and a subsurface investigation and field and aerial reconnaissance conducted for preparation of this {CCNPP Unit 3} application. {Previous site-specific reports reviewed include the Preliminary Safety Analysis Report (BGE, 1968) and the Independent Spent Fuel Storage Installation Safety Analysis Report (CEG, 2005).} A review of published geologic literature was used to supplement and update the existing geological and seismological information. In addition, relevant unpublished geologic literature, studies, and projects were identified by contacting the U.S. Geological Survey (USGS), State geological surveys and universities. The list of references used to compile the geological and seismological information is presented in the applicable section.

{Field reconnaissance of the site and within a 25 mi (40 km) radius of the site was conducted by geologists in teams of two or more. Two field reconnaissance visits in late summer and autumn 2006 focused on exposed portions of the Calvert Cliffs, other cliff exposures along the west shore of Chesapeake Bay, and roads traversing the site and a 5 mi (8 km) radius of the CCNPP site. Key observations and discussion items were documented in field notebooks and photographs. Field locations were logged by hand on detailed topographic base maps and with hand-held Global Positioning System (GPS) receivers.

Aerial reconnaissance within a 25 mi (40 km) radius of the site was conducted by two geologists in a top-wing Cessna aircraft on January 3, 2007. The aerial reconnaissance investigated the geomorphology of the Chesapeake Bay area and targeted numerous previously mapped geologic features and potential seismic sources within a 200 mi (322 km) radius of the CCNPP site (e.g., Mountain Run fault zone, Stafford fault system, Brandywine fault zone, Port Royal fault zone, and Skinkers Neck anticline). The flight crossed over the CCNPP site briefly but did not circle or approach the site closely in order to comply with restrictions imposed by the Federal Aviation Administration. Key observations and discussion items were documented in field notebooks and photographs. The flight path, photograph locations, and locations of key observations were logged with hand-held GPS receivers.

The investigations of regional and site physiographic provinces and geomorphic process, geologic history, and stratigraphy were conducted by Bechtel Power Corporation. The investigations of regional and site tectonics and structural geology were conducted by William Lettis and Associates.

This section is intended to demonstrate compliance with the requirements of paragraph c of 10 CFR 100.23, "Geologic and Seismic Siting Criteria" (CFR, 2007).

2.5.1 BASIC GEOLOGIC AND SEISMIC INFORMATION

This section of the DCD is incorporated by reference with the following departure(s) and/or supplement(s).

The U.S EPR DCD includes the following COL Item in Section 2.5.1:

A COL applicant that references the U.S. EPR design certification will use site-specific information to investigate and provide data concerning geological, seismic geophysical and geotechnical information.

The COL Item is addressed in the following sections.

This section presents information on the geological and seismological characteristics of the {CCNPP} site region (200 mi (322 km) radius), site vicinity (25 mi (40 km) radius), site area (5 mi (8 km) radius) and site (0.6 mi (1 km) radius). Section 2.5.1.1 describes the geologic and tectonic characteristics of the site region. Section 2.5.1.2 describes the geologic and tectonic characteristics of the {CCNPP} site vicinity and site location. The geological and seismological information was developed in accordance with the following NRC guidance documents:

- Regulatory Guide 1.70, Section 2.5.1, "Basic Geologic and Seismic Information," (NRC, 1978)
- Regulatory Guide 1.206, Section 2.5.1, "Basic Geologic and Seismic Information," (NRC, 2007) and
- Regulatory Guide 1.165, "Identification and Characterization of Seismic Sources and Determination of Safe Shutdown Earthquake Ground Motion," (NRC, 1997).

2.5.1.1 Regional Geology (200 mi (322 km) radius)

This section discusses the physiography, geologic history, stratigraphy, and tectonic setting within a 200 mi (322 km) radius of the {CCNPP} site. The regional geologic map and explanation as shown in Figure 2.5.1-5a and Figure 2.5.1-5b contain information on the geology, stratigraphy, and tectonic setting of the region surrounding the {CCNPP site (Schruben, 1994)}. Summaries of these aspects of regional geology are presented to provide the framework for evaluation of the geologic and seismologic hazards presented in the succeeding sections.

2.5.1.1.1 Regional Physiography and Geomorphology

{The CCNPP site lies within the Coastal Plain Physiographic Province as shown in Figure 2.5.1-1 (Fenneman, 1946). The area within a 200 mi (322 km) radius of the site encompasses parts of five other physiographic provinces. These are: the Continental Shelf Physiographic Province, which is located east of the Coastal Plain Province, and the Piedmont, Blue Ridge, Valley and Ridge and Appalachian Plateau physiographic provinces, which are located successively west and northwest of the Piedmont Province (Thelin, 1991).

Each of these physiographic provinces is briefly described in the following sections. The physiographic provinces in the site region are shown on Figure 2.5.1-1 (Fenneman, 1946). A map showing the physiographic provinces of Maryland, as depicted by the Maryland Geological Survey (MGS), is shown on Figure 2.5.1-6.

2.5.1.1.1.1 Coastal Plain Physiographic Province

The Coastal Plain Physiographic Province extends eastward from the Fall Line (the physiographic and structural boundary between the Coastal Plain Province and the Piedmont Province) to the coastline as shown in Figure 2.5.1-1. The Coastal Plain Province is a low-lying, gently-rolling terrain developed on a wedge-shaped, eastward-dipping mass of Cretaceous, Tertiary, and Quaternary age as shown in Figure 2.5.1-5a and Figure 2.5.1-5b, which are unconsolidated and semi-consolidated sediments (gravels, sands, silts, and clays), that thicken toward the coast. This wedge of sediments attains a thickness of more than 8,000 ft (2,430 m) along the coast of Maryland (MGS, 2007). In general, the Coastal Plain Province is an area of lower topographic relief than the Piedmont Province to the west. Elevations in the Coastal Plain Province of Maryland range from near sea level to 290 ft (88 m) above sea level near the District of Columbia - Prince Georges County line (Otton, 1955).

Four main periods of continental glaciation occurred in the site region during the Pleistocene. Glaciers advanced only as far south as northeastern Pennsylvania and central New Jersey as shown in Figure 2.5.1-5a and Figure 2.5.1-5b. However, continental glaciation affected sea level and both coastal and fluvial geomorphic processes, resulting in the landforms that dominate the Coastal Plain Province.

In Maryland, the MGS subdivides the Coastal Plain Physiographic Province into the Western Shore Uplands and Lowlands regions, the Embayment occupied by the Chesapeake Estuary system, and the Delmarva Peninsula Region on the Eastern Shore of the Chesapeake Bay as shown in Figure 2.5.1-6. In the site region and vicinity, geomorphic surface expression is a useful criterion for mapping the contacts between Pliocene and Quaternary units as shown in Figure 2.5.1-5a and Figure 2.5.1-5b. Constructional surface deposits define the tops of estuarine and fluvial terraces and erosional scarps correspond with the sides of old estuaries (McCartan, 1989a) (McCartan, 1989b). In some areas, the physiographic expression of terraces that might have formed in response to alternate deposition and erosion during successive glacial stages is poorly defined (Glaser, 1994) (Glaser, 2003c). Sea levels were relatively lower during glacial stages than present-day, and relatively higher than present-day during interglacial stages. Deposition and erosion during periods of higher sea levels led to the formation of several discontinuous Quaternary-age stream terraces that are difficult to correlate (McCartan, 1989a). The distribution of Quaternary surficial deposits in the CCNPP site area and site location is discussed in Section 2.5.1.2. Northeast of the Chesapeake Bay, the Western Shore Uplands Region consists of extensive areas of relatively little topographic relief, less than 100 ft (30 m). The Western Shore Lowlands Region located along the west shore of Chesapeake Bay and north of the Western Shore Uplands Region as shown in Figure 2.5.1-6 is underlain by interbedded quartz-rich gravels and sands of the Cretaceous Potomac Group and gravel, sand, silt and clay of the Quaternary Lowland deposits. During glacial retreats, large volumes of glacial melt-waters formed broad, high energy streams such as the ancestral Delaware, Susquehanna, and Potomac Rivers that incised deep canyons into the continental shelf. Southwest of the Chesapeake Bay, marine and fluvial terraces developed during the Pliocene and Pleistocene. As a result of post-Pleistocene sea level rise, the outline of the present day coastline is controlled by the configuration of drowned valleys, typified by the deeply recessed Chesapeake Bay and Delaware Bay. Exposed headlands and shorelines have been modified by the development of barrier islands and extensive lagoons (PSEG, 2002).

2.5.1.1.1.2 Continental Shelf Physiographic Province

The Continental Shelf Physiographic Province is the submerged continuation of the Coastal Plain Province and extends from the shoreline to the continental slope as shown in Figure 2.5.1-1. The shelf is characterized by a shallow gradient of approximately 10 ft/mi to the southeast (Schmidt, 1992) and many shallow water features that are relicts of lower sea levels. The shelf extends eastward for about 75 to 80 mi (121 to 129 km) , where sediments reach a maximum thickness of about 40,000 ft (12.2 km) (Edwards, 1981). The eastward margin of the continental shelf is marked by the distinct break in slope to the continental rise with a gradient of approximately 400 ft/mi (Schmidt, 1992).

2.5.1.1.1.3 Piedmont Physiographic Province

The Piedmont Physiographic Province extends southwest from New York to Alabama and lies west of, and adjacent to, the Coastal Plain Physiographic Province as shown in Figure 2.5.1-1. The Piedmont is a rolling to hilly province that extends from the Fall Line in the east to the foot of the Blue Ridge Mountains in the west as shown in Figure 2.5.1-1. The Fall Line is a low east-facing topographic scarp that separates crystalline rocks of the Piedmont Province to the west from less resistant sediments of the Coastal Plain Province to the east (Otton, 1955) (Vigil, 2000). The Piedmont Province is about 40 mi (64 km) wide in southern Maryland and narrows northward to about 10 mi (16 km) wide in southeastern New York.

Within the site region, the Piedmont Province is generally characterized by deeply weathered bedrock and a relative paucity of solid rock outcrop (Hunt, 1972). Residual soil (saprolite) covers the bedrock to varying depths. On hill slopes, the saprolite is capped locally by colluvium (Hunt, 1972).

In Maryland, the Piedmont Province is divided into the Piedmont Upland section to the east and the Piedmont Lowland section to the west, which is referred to as a sub-province in some publications as shown in Figure 2.5.1-6. The Piedmont Upland section is underlain by metamorphosed sedimentary and crystalline rocks of Precambrian to Paleozoic age. These lithologies are relatively resistant and their erosion has resulted in a moderately irregular surface. Topographically higher terrain is underlain by Precambrian crystalline rocks and Paleozoic quartzite and igneous intrusive rocks. The Piedmont Lowland section is a less rugged terrain containing fault-bounded basins filled with sedimentary and igneous rocks of Triassic and Early Jurassic age.

2.5.1.1.1.4 Blue Ridge Physiographic Province

The Blue Ridge Physiographic Province is bounded on the east by the Piedmont Province and on the west by the Valley and Ridge Province as shown in Figure 2.5.1-1. The Blue Ridge Province, aligned in a northeast-southwest direction, extends from Pennsylvania to northern Georgia. It varies in approximate width from 5 mi (8 km) to more than 50 mi (80 km) (Hunt, 1967). This province corresponds with the core of the Appalachians and is underlain chiefly by more resistant granites and granitic gneisses, other crystalline rocks, metabasalts (greenstones), phyllites, and quartzite along its crest and eastern slopes.

2.5.1.1.1.5 Valley and Ridge Physiographic Province

The Valley and Ridge Physiographic Province lies west of the Blue Ridge Province and east of the Appalachian Plateau Province as shown in Figure 2.5.1-1. This is designated as the Valley and Ridge Province in Maryland as shown in Figure 2.5.1-6. Valleys and ridges are aligned in a northeast-southwest direction in this province, which is between 25 and 50 mi (40 and 80 km) wide. The sedimentary rocks underlying the Valley and Ridge Province are tightly folded and, in some locations, faulted. Sandstone units that are more resistant to weathering are the ridge formers. Less resistant shales and limestones underlie most of the valleys as shown in Figure 2.5.1-5a and Figure 2.5.1-5b. The Great Valley Section of the province as shown in Figure 2.5.1-6, to the east, is divided into many distinct lowlands by ridges or knobs, the largest lowland being the Shenandoah Valley in Virginia. This broad valley is underlain by shales and by limestones that are prone to dissolution, resulting in the formation of sinkholes and caves. Elevations within the Shenandoah Valley typically range between 500 and 1,200 ft (152 and 366 m) msl. The western portion of the Valley and Ridge Province is characterized by a series of roughly parallel ridges and valleys, some of which are long and narrow (Lane, 1983). Elevations within the ridges and valleys range from about 1,000 to 4,500 ft (305 to 1,372 m) msl (Bailey, 1999).

2.5.1.1.1.6 Appalachian Plateau Physiographic Province

Located west of the Valley and Ridge Province, the Appalachian Plateau Physiographic Province includes the western part of the Appalachian Mountains, stretching from New York to Alabama as shown in Figure 2.5.1 1. The Allegheny Front is the topographic and structural boundary between the Appalachian Plateau and the Valley and Ridge Province (Clark, 1992). It is a bold, high escarpment, underlain primarily by clastic sedimentary rocks capped by sandstone and conglomerates. In eastern West Virginia, elevations along this escarpment reach 4,790 ft (1,460 m) (Hack, 1989). West of the Allegheny Front, the Appalachian Plateau's topographic surface slopes gently to the northwest and merges imperceptibly into the Interior Low Plateaus. Only a small portion of this province lies within 200 mi (322 km) of the CCNPP site as shown in Figure 2.5.1-1.

The Appalachian Plateau Physiographic Province is underlain by sedimentary rocks such as sandstone, shale, and coal of Cambrian to Permian age as shown in Figure 2.5.1-5a and Figure 2.5.1-5b. These strata are generally subhorizontal to gently folded into broad synclines and anticlines and exhibit relatively little deformation. These sedimentary rocks differ significantly from each other with respect to resistance to weathering. Sandstone units tend to be more resistant to weathering and form topographic ridges. The relatively less resistant shales and siltstones weather preferentially and underlie most valleys. The Appalachian Plateau is deeply dissected by streams into a maze of deep, narrow valleys and high narrow ridges (Lane, 1983). Limestone dissolution and sinkholes occur where limestone units with high karst susceptibility occur at or near the ground surface.}

2.5.1.1.2 Regional Geologic History

{The geologic and tectonic setting of the CCNPP site region is the product of a long, complex history of continental and island arc collisions and rifting, which spanned a period of over one billion years and formed the Appalachian Mountains (Appalachian Orogen) extended continental crust and coastal plain as shown in Figure 2.5.1-7. This history of deformation imparts a pre-existing structural grain in the crust that is important for understanding the current seismotectonic setting of the region. Episodes of continental collisions have produced a series of accreted terranes separated, in part, by low angle detachment faults. Sources of seismicity may occur in the overlying, exposed, or buried terranes or may occur along structures within the North American basement buried beneath the accreted terranes or overthrust plates. That is, regional seismicity may not be related to any known surface structure. Intervening episodes of continental rifting have produced high angle normal or transtensional faults that either sole downward into detachment faults or penetrate entirely through the accreted terranes and upper crust. Understanding the history of the evolution and the geometry of these crustal faults, therefore, is important for identifying potentially active faults and evaluating the distribution of historical seismicity within the tectonic context of the site region.

Major tectonic events in the site region include five compressional orogenies and two extensional episodes (Faill, 1997a). While direct evidence of these deformational events is visible in the Blue Ridge and Piedmont provinces, it is buried beneath the coastal plain sediments in the site region but is inferred from geophysical data, as described in Section 2.5.1.1.4.3, and borehole data as described in Section 2.5.1.1.3. The site region is located currently on the passive, divergent trailing margin of the North American plate following the last episode of continental extension and rifting. Each of these tectonic events is described in the following paragraphs.

2.5.1.1.2.1 Grenville Orogeny

The earliest of the compressional deformational events (orogenies) recorded in the rocks of North America is the Grenville orogeny that occurred during Middle to Late Precambrian (Proterozoic) time, approximately one billion years ago, as a result of the convergence of the ancestral North American and African tectonic plates. During this orogeny, various terranes were accreted onto the edge of the ancestral North American plate, forming the Grenville Mountains (Faill, 1997a), which were likely the size of the present day Himalayas (Fichter, 2000). The Grenville orogeny was followed by several hundred million years of tectonic quiescence, during which time the Grenville Mountains were eroded and their basement rocks exposed. In Virginia and Maryland, the Grenville basement rocks are exposed in the Blue Ridge Province and portions of the Piedmont Province (Fichter, 2000). This appears to be represented in Maryland by the Middletown Valley biotite granite gneiss in the Blue Ridge Province and the Baltimore Gneiss in the eastern Piedmont Province.

2.5.1.1.2.2 Late Precambrian Rifting

Following the Grenville orogeny, crustal extension and rifting began during Late Precambrian time, which caused the separation of the North America and African plates and created the proto-Atlantic Ocean (Iapetus Ocean). Rifting is interpreted to have occurred over a relatively large area, sub-parallel to the present day Appalachian mountain range (Faill, 1997a) (Wheeler, 1996). This period of crustal extension is documented by the metavolcanics of the Catoctin, Swift Run, and Sams Creek formations (Schmidt, 1992). During rifting, the newly formed continental margin began to subside and accumulate sediment. Initial sedimentation resulted in an eastward thickening wedge of clastic sediments consisting of graywackes, arkoses, and shales deposited unconformably on the Grenville basement rocks. In the Blue Ridge and western Piedmont, the Weverton and Sugarloaf Mountain quartzites represent late Precambrian to early Cambrian fluvial and beach deposits. Subsequent sedimentation included a transgressive sequence of additional clastic sediments followed by a thick and extensive sequence of carbonate sediments. Remnants of the rocks formed from these sediments can be found within the Valley and Ridge Province and Piedmont Province (Fichter, 2000). In the western Piedmont, the sandy Antietam Formation was deposited in a shallow sea. In the Valley and Ridge Province, a carbonate bank provided the environment of deposition for the thick carbonates ranging from the Cambrian Tomstown Dolomite through the Ordovician Chambersburg Formation. In the eastern Piedmont, the Setters Formation (quartzite and interbedded mica schist) and the Cockeysville Marble have been interpreted as metamorphosed beach and carbonate bank deposits that can be correlated from Connecticut to Virginia. Accumulation of this eastward thickening wedge of clastic and carbonate sediments is thought to have occurred from the Middle to Late Cambrian into Ordovician time (PSEG, 2002).

2.5.1.1.2.3 Late Precambrian to Early Cambrian Orogenies (Potomac/Penobscot Orogency)

Fossil fauna, detailed geologic mapping, petrologic investigations, and radiometric age dates indicate that the Virgilina orogeny is a Late Proterozoic-earliest Cambrian compressional deformation event that may have involved the accretion of a crustally juvenile Carolina zone to a more crustally evolved Goochland zone in the Carolinas and southern Virginia (Hibbard, 1995) as shown in Figure 2.5.1-7. Island arc rifting in the Carolina zone might have been associated with the Virgilina orogeny. It is possible that the Virgilina orogeny deformed the Mather Gorge Formation in the central Piedmont of Maryland and northern Virginia. The Sykesville Formation in the same area contains olistoliths of Mather Gorge phyllonite (Drake, 1999). Because the Sykesville Formation was folded prior to the emplacement of the Early Ordovician Falls Church Intrusive Suite and Occoquan Granite, that folding, originally interpreted as a result of the Penobscot orogeny, is now believed to have formed as a result of the Cambrian to earliest Ordovician Potomac orogeny. The deformation, metamorphism and west-directed thrusting affected the western portion of the Piedmont in the Potomac River Valley (Hibbard, 1995) (Drake, 1999).

During Late Cambrian time, as the now tectonically stable continental margin continued to subside, micro-continents and volcanic arcs, characteristic of an intra-oceanic island-arc terrane, began to develop in the proto-Atlantic Ocean as a result of east-directed oceanic subduction and initial closing of the proto-Atlantic. The Penobscot orogeny (documented in the Maritime Provinces of Canada) is thought to have been caused by crustal convergence and accretion of these volcanic arcs thrust over micro-continents along the North American plate margin as shown in Figure 2.5.1-7. This orogeny is considered to represent the beginning of the convergent phase in the closing of the proto-Atlantic Ocean (Fichter, 2000). Subsequent convergent phases in the closing of the proto-Atlantic include the Taconic and Acadian orogenies and the Allegheny orogeny that finally closed the proto-Atlantic in the Permian.

2.5.1.1.2.4 Taconic Orogeny

The Taconic orogeny occurred during Middle to Late Ordovician time and was caused by continued collision of micro-continents and volcanic arcs with eastern North America along an eastward dipping subduction zone during progressive closure of the proto-Atlantic Ocean as shown in Figure 2.5.1-7. Taconic terranes are preserved today in the Piedmont in a series of belts representing island-arcs and micro-continents. They include the Chopawamsic belt, the Carolina Slate belt, the Eastern Slate belt, the Goochland-Raleigh belt as shown in Figure 2.5.1-8 (Bledsoe, 1980) (Fichter, 2000), and the Sussex Terrane, directly west of the CCNPP site. These Taconic terranes are considered to have collided with, and accreted to, eastern North America at different times during the orogeny (Fichter and Baedke, 2000). Closer to the CCNPP site, the central Piedmont in Northern Virginia, Maryland, and Pennsylvania contains several belts of rocks whose age is unknown and/or whose relation to the pre- or synorogenic rocks of the Taconic Orogen is uncertain (Drake, 1999). These stratigraphic units include the Wissahickon Formation, which is now recognized in the Potomac Valley as three distinct lithotectonic assemblages (Drake, 1999). Other stratigraphic units, whose ages range from Late Proterozoic to Late Ordovician and contain indications of Taconic deformation, include various units in the Ijamsville Belt, the Glenarm Group Belt, which includes the Baltimore Gneiss, the Potomac terrane that was thrust over the Glenarm Group belt, and the Baltimore mafic complex to the east as shown in Figure 2.5.1-8 (Horton, 1989) (Bledsoe, 1980) (Fichter, 2000). Additional details on the complex stratigraphy of the Taconic orogen in the Piedmont are contained in Drake (Drake, 1999).

Accretion of the island-arcs and micro-continents to the eastern margin of North America created a mountain system, the Taconic Mountains, that became a major barrier between the proto-Atlantic to the east and the carbonate platform to the west. The growth of this barrier transformed the area underlain by carbonate sediments to the west into a vast, elongate sedimentary basin, the Appalachian Basin. The present day Appalachian Basin extends from the Canadian Shield in southern Quebec and Ontario Provinces, Canada, southwestward to central Alabama, approximately parallel to the Atlantic coastline (Colton, 1970). The formation of the Appalachian Basin is one of the most significant consequences of the Taconic orogeny in the region defined by the Valley and Ridge Province and Appalachian Plateau Province. The Taconic mountain system was the source of most of the siliclastic sediment that accumulated in the Appalachian Basin during Late Ordovician and Early Silurian time. Many of these units are preserved closest to the CCNPP site in the Valley and Ridge Province. A continent-wide transgression in Early Silurian time brought marine shales and carbonate sedimentation eastward over much of the basin, and a series of transgressions and regressions thereafter repeatedly shifted the shoreline and shallow marine facies. Carbonate deposition continued in the eastern part of the basin into Early Devonian time (Faill, 1997b).

2.5.1.1.2.5 Acadian Orogeny

The Acadian orogeny (Figure 2.5.1-7) was caused by the collision of the micro-continent Avalon with eastern North America during the Middle to Late Devonian Period. At its peak, the orogeny produced a continuous chain of mountains along the east coast of North America and brought with it associated volcanism and metamorphism. Remnants of the Avalon terrane (the Acadian Mountains) can be found in the Piedmont Province within the pre-existing Taconic Goochland belt, Carolina Slate belt, and the Chopawamsic belt (Fichter, 2000). The Acadian orogeny ended the largely quiescent environment that dominated the Appalachian Basin during the Silurian, as vast amounts of terrigenous sediment from the Acadian Mountains were introduced into the basin and formed the Catskill clastic wedge in Pennsylvania and New York as shown in Figure 2.5.1-5a, Figure 2.5.1-5b, and Figure 2.5.1-7. Thick accumulations of clastic sediments belonging to the Catskill Formation are spread throughout the Valley and Ridge Province (Faill, 1997b). During the Mississippian Period, the Acadian Mountains were completely eroded, and the basement rocks of the Avalon terrane were exposed (Fichter, 2000).

2.5.1.1.2.6 Allegheny Orogeny

The Allegheny orogeny occurred during the Late Carboniferous Period and extended into the Permian Period. The orogeny represents the final convergent phase in the closing of the proto-Atlantic Ocean in the Paleozoic Era (Figure 2.5.1-7). Metamorphism and magmatism were significant events during the early part of the Allegheny orogeny. The Allegheny orogeny was caused by the collision of the North American and African plates, and it produced the Allegheny Mountains. As the African continent was thrust westward over North America, the Taconic and Acadian terranes became detached and also were thrust westward over Grenville basement rocks (Fichter, 2000). The northwest movement of the displaced rock mass above the thrust was progressively converted into the deformation of the rock mass, primarily in the form of thrust faults and fold-and-thrust structures, as seen in the Blue Ridge and Piedmont Plateau Provinces. The youngest manifestation of the Allegheny orogeny was northeast-trending strike-slip faults and shear zones in the Piedmont Province. The extensive, thick, and undeformed Appalachian Basin and its underlying sequence of carbonate sediments were deformed and a fold-and-thrust array of structures, long considered the classic Appalachian structure, was impressed upon the basin. The tectonism produced the Allegheny Mountains and a vast alluvial plain to the northwest. The Allegheny Front along the eastern margin of the Appalachian Plateau Province is thought to represent the westernmost extent of the Allegheny orogeny. Rocks throughout the Valley and Ridge Province are thrust faulted and folded up to this front, whereupon they become relatively flat and only slightly folded west of the Allegheny Front (Faill, 1998).

2.5.1.1.2.7 Early Mesozoic Extensional Episode (Triassic Rifting)

Crustal extension during Early Mesozoic time (Late Triassic and Early Jurassic) marked the opening of the Atlantic Ocean (Figure 2.5.1-7). This extensional episode produced numerous local, closed basins ("Triassic basins") along eastern North America (Faill, 1998). The elongate basins generally trend northeast, parallel to the pre-existing Paleozoic structures (Figure 2.5.1-9). The basins range in length from less than 20 mi (32 km) to over 100 mi (161 km) and in width from less than 5 mi (8 km) to over 50 mi (80 km) . The basins are exposed in the Piedmont Lowland of Maryland and Northern Virginia (Gettysburg and Culpeper Basins) and are also buried beneath sediments of the Coastal Plain. The closest exposed basin to the site, the Gettysburg Basin, extends northeast from the Frederick Valley at the south end of the basin into Pennsylvania. Valleys in these Mesozoic basins are developed on sandstone and shale units and trend northeast-southwest, parallel to the strike of the bedrock. Generally, the basins are asymmetric half-grabens with principal faults located along the western margin of the basins. Triassic and Jurassic rocks that fill the basins primarily consist of conglomerates, sandstones, and shales interbedded with basaltic lava flows. At several locations, these rocks are cross-cut by basaltic dikes. The basaltic rocks are generally more resistant to erosion and form local topographically higher landforms. In the Frederick Valley, the younger Mesozoic units are deposited on Ordovician age limestone units subject to dissolution and karst development. Areas in the Fredrick Valley underlain by limestone subject to dissolution have relatively low relief compared to the higher and more rugged terrain underlain by intrusive and extrusive rocks consisting predominantly of diabase and basalt (Brezinski, 2004).

2.5.1.1.2.8 Cenozoic History

The Early Mesozoic extensional episode gave rise to the Cenozoic Mid-Atlantic spreading center. The Atlantic seaboard presently represents the trailing passive margin related to the spreading at the Mid-Atlantic ridge. Ridge push forces resulting from the Mid-Atlantic spreading center are believed to be responsible for the northeast-southwest directed horizontal compressive stress presently observed along the Atlantic seaboard.

During Cenozoic time, as the Atlantic Ocean opened, the newly formed continental margin cooled and subsided, leading to the present day passive trailing divergent continental margin. As the continental margin developed, continued erosion of the Appalachian Mountains produced extensive sedimentation within the Coastal Plain. The Cenozoic history of the Atlantic continental margin, therefore, is preserved in the sediments of the Coastal Plain Province, and under water along the continental shelf. The geologic record consists of a gently east-dipping, seaward-thickening wedge of sediments, caused by both subsidence of the continental margin and fluctuations in sea level. Sediments of the Coastal Plain Province cover igneous and metamorphic basement rocks and Triassic basin rift deposits.

During the Quaternary Period much of the northern United States experienced multiple glaciations interspersed with warm interglacial episodes. The last (Wisconsinan) Laurentide ice sheet advanced over much of North America during the Pleistocene. The southern limit of glaciation extended into parts of northern Pennsylvania and New Jersey, but did not cover the CCNPP site vicinity (Figure 2.5.1-5a and Figure 2.5.1-5b). South of the ice sheet, periglacial environments persisted throughout the site region (Conners, 1986). Present-day Holocene landscapes, therefore, are partially the result of geomorphic processes, responding to isostatic uplift, eustatic sea level change, and alternating periglacial and humid to temperate climatic conditions (Cleaves, 2000).}

2.5.1.1.3 Regional Stratigraphy

{This section contains information on the regional stratigraphy within each of the physiographic provinces. The regional geology and generalized stratigraphy within a 200 mi (322 km) radius of the CCNPP site is shown on Figure 2.5.1-5a and Figure 2.5.1-5b.

2.5.1.1.3.1 Coastal Plain Physiographic Province

2.5.1.1.3.1.1 Pre-Cretaceous Basement Rock

As described in the subsection on Cenozoic History (Section 2.5.1.1.2.7), early Mesozoic rifting and opening of the Atlantic Ocean was followed by the sea floor spreading and the continued opening of the Atlantic Ocean during the Cenozoic time. Continued erosion of the Appalachian Mountains and the exposed Piedmont produced extensive sedimentation within the Coastal Plain Province that includes the CCNPP site region.

The non-marine and marine sediments deposited in the Coastal Plain Physiographic Province overlie what are most likely foliated metamorphic or granitic rocks, similar to those cropping out in the Piedmont approximately 50 mi (80 km) to the northwest (Figure 2.5.1-5a and Figure 2.5.1-5b). The Pre-Cretaceous basement bedrock is only encountered in the Coastal Plain Province by borings designed to characterize deep aquifers above the underlying basement rock. The closest borehole to the CCNPP site that penetrates the basement rock is located in St. Mary's County about 13 mi (21 km) south of the site (Figure 2.5.1-10). It has been indicated (Hansen, 1986) that most of the borings that penetrate coastal plain sediments and extend to the underlying basement have encountered metamorphic or igneous rocks. For example, well DO-CE 88 in Dorchester, County located approximately 24 mi (39 km) east of the CCNPP site was drilled into gneissic basement rock at 3,304 ft (1,007 m) in depth (Figure 2.5.1-10). Well QA-EB 110, in Queen Anne's County, located 38 mi (61 km) north of the CCNPP site, was drilled to explore for deep freshwater aquifers. This well was drilled into basement at a depth of 2,518 ft (767 m). The basement rock was only sampled in the drill cuttings and suggests a gneiss/schist from the mineralogy present, (i.e., biotite, chlorite, and clear quartz).

Regional geophysical and scattered borehole data indicate that a Mesozoic basin might be present in the site vicinity, buried beneath Coastal Plain sediments. Triassic clastic deposits, indicative of a possible rift basin, were penetrated in Charles County (well CH-CE 37), located over 20 mi (32 km) west of the site, for an interval of 99 ft (30 m), returning samples of weathered brick red clay and shale.

Diabase was cored in the closest deep boring (SM-DF 84) to the CCNPP site that penetrated the Pre-Cretaceous basement. The boring is located in Lexington Park, St. Mary's County, about 13 mi (21 km) south of the CCNPP site (Hansen, 1984) (Figure 2.5.1-10). A statement regarding the presence of the diabase was made (Hansen, 1984):

As no other basement lithologies were encountered, it is presently not known whether the diabase is from a sill or dike associated with the rift-basin sediments or whether it is cross-cutting the crystalline rocks. The diabase is apparently a one-pyroxene (augite) rock, which Fisher (1964, p. 14) suggests is evidence of rapid, undifferentiated crystallization in a relatively thin intrusive body, such as a dike.

The occurrence of Mesozoic rift-basin rocks in St. Mary's and Prince George's County are further discussed (Hansen, 1986): "The basins that occur in Maryland are all half-grabens with near-vertical border faults along the western sides. The strata generally strike north-easterly, but, in places, particularly in the vicinity of cross-faults, strike may diverge greatly from the average."

Because of the depth of Coastal Plain sediments, the basement rock type beneath the CCNPP site must be inferred based on surrounding borings and geophysical data. The presence and character of basement rock beneath the CCNPP site is discussed further in Section 2.5.1.2.

2.5.1.1.3.1.2 Cretaceous Stratigraphic Units

Regionally, coastal plain deposits lap onto portions of the eastern Piedmont. In Stafford, Prince William, and Fairfax counties in Virginia Lower Cretaceous Potomac Formation sediments were deposited unconformably on a narrow belt of Ordovician Quantico Slate and on the Cambrian Chopawamsic Formation (Mixon, 2000). The Potomac Formation occurs on Proterozoic to Cambrian metamorphic and igneous rocks in the Washington DC area (McCartan, 1990).

The Lower Cretaceous Potomac Group overlies a complex suite of basement rocks that includes strata as young as Triassic. Jurassic units appear to be missing north of the Norfolk Arch (Hansen, 1978) (Figure 2.5.1-11). The undulatory and east-dipping basement surface that underlies the Coastal Plain resulted from a combination of downwarping, erosion, and faulting. This has led to local variations in the slope of the bedrock surface. The Coastal Plain sediments deposited east of the Fall Line, range from Early Cretaceous to Quaternary in age and consist of interbedded silty clays, sands, and gravels that were deposited in both marine and non-marine environments. These sediments dip and thicken toward the southeast. Whereas the basement surface dips southeast at about 100 ft/mi in Charles County, west of the CCNPP site, a marker bed in the middle of the Cretaceous Potomac Group dips southeast at about 50 ft per mile (McCartan, 1989a). This wedge of unlithified sediments consists of Early Cretaceous terrestrial sediments and an overlying sequence of well-defined, Late Cretaceous, marine stratigraphic units. These units from oldest to youngest are summarized in the following paragraphs.

The Lower Cretaceous strata of the Potomac Group consists of a thick succession of variegated red, brown, maroon, yellow, and gray silts and clays with interstratified beds of fine to coarse gray and tan sand. In the Baltimore-Washington area, the Potomac Group is subdivided from oldest to youngest into the Patuxent, Arundel, and Patapsco Formations. This subdivision is recognizable in the greater Washington-Baltimore area where the clayey Arundel Formation is easily recognized and separates the two dominantly sandy formations (Hansen, 1984). This distinction is less pronounced to the east and southeast where the Potomac Group is divided into the Arundel/Patuxent formations (undivided) and the overlying Patapsco Formation. At Lexington Park, Maryland, the clayey beds that dominate the formation below a depth of 1,797 ft (548 m) are assigned to the Arundel/Patuxent Formations (undivided) (Hansen, 1984).

At the Lexington Park well, located about 13 mi (21 km) south of the CCNPP site (Figure 2.5.1-10), about 30 ft (9 m) of a denser, acoustically faster, light gray, fine to medium clayey sand occurs at the base of the Potomac Group and might represent an early Cretaceous, pre-Patuxent Formation. These sediments might correlate with the Waste Gate Formation encountered east of Chesapeake Bay in the DOE Crisfield No. 1 well (Hansen, 1984).

The Patapsco Formation contains interbedded sands, silts, and clays, but it contains more sand than the overlying Arundel/Patuxent Formations (undivided). The contact is marked by an interval dominated by thicker clay deposits. The Arundel/Patuxent Formations (undivided) are marked by the absence of marine deposits. The Mattaponi Formation was proposed (Cederstrom, 1957) for the stratigraphic interval immediately above the Patapsco Formation. An identified interval (Hansen, 1984) as the Mattaponi (?) is now recognized as part of the upper Patapsco Formation. In general, it appears that downwarping associated with the Salisbury Embayment (Figure 2.5.1-11) began early in the Cretaceous and continued intermittently throughout the Cretaceous and Tertiary periods. Deposition apparently kept pace, resulting in a fluvial-deltaic environment. Biostratigraphic data from test wells on the west side of Chesapeake Bay indicate that Upper Cretaceous sediments reach maximum thickness in Anne Arundel County and show progressive thinning to the south. This appears to reflect deposition within the downwarping, northwest-trending Salisbury Embayment during the Cretaceous (Hansen, 1978). In southern Calvert County, the Upper Cretaceous Aquia Formation rests unconformably on Lower Cretaceous sediments (Figure 2.5.1-12). Thinning and overlapping within the Upper Cretaceous interval suggests that the northern flank of the Norfolk Arch was tectonically active during late Cretaceous time (Hansen, 1978) (Figure 2.5.1-11).

The Upper Cretaceous Magothy Formation is approximately 200 ft (61 m) thick in northern Calvert County but becomes considerably thinner southward at the CCNPP site and pinches out south of the site and north of wells in Solomons and Lexington Park, Maryland (Hansen, 1996) (Achmad, 1997) (Figure 2.5.1-12). This pattern also appears to reflect thicker deposition in the Salisbury Embayment. The Magothy Formation is intermittently exposed near Severna Park, Maryland, and in the interstream area between the Severn and Magothy Rivers. This outcrop belt becomes thinner to the south in Prince Georges County. The Magothy consists mainly of lignitic or carbonaceous light gray to yellowish quartz sand interbedded with clay layers. The sand is commonly coarse and arkosic and in many places is cross bedded or laminar. Pyrite and glauconite occur locally (Otton, 1955).

The upper Cretaceous Matawan and Monmouth formations are exposed in Anne Arundel County, Maryland. While the Matawan is absent in Prince Georges County, the Monmouth crops out in a narrow belt near Bowie, Maryland. Exposures of these formations have not been identified in Charles County. These formations are inseparable in sample cuttings and drillers' logs and are undifferentiated in southern Maryland (Otton 1955) (Hansen, 1996). They consist mainly of gray to grayish-black micaceous sandy clay and weather to a grayish brown. Glauconite is common in both formations and fossils include fish remains, gastropods, pelecypods, foraminifera, and ostracods. The presence of glauconite and this fossil fauna indicate that the Matawan and Monmouth are the oldest in a sequence of marine formations. These formations range in thickness from a few feet or less in their outcrop area to more than 130 ft (40 m) at the Annapolis Water Works (Otton, 1955). The formations thin to the west and average about 45 ft (14 m) in Prince Georges County. The combined formations along with the Brightseat Formation form the Lower Confining Beds (Section 2.4.12) that become progressively thinner from southern Anne Arundel County through Calvert County to St. Mary's County where this hydrostratigraphic unit appears to consist mainly of the Brightseat Formation (Hansen, 1996).

2.5.1.1.3.1.3 Tertiary Stratigraphic Units

The Brightseat Formation is exposed in a few localities in Prince Georges County and contains foraminifera of Paleocene age. This unit is relatively thin (up to about 25 ft (8m)) but occurs widely in Calvert and St. Mary's counties. It is generally medium and olive gray to black, clayey, very fine to fine sand that is commonly micaceous and /or phosphatic (Otton, 1955) (Hansen, 1996). It can be distinguished from the overlying Aquia Formation by the absence or sparse occurrence of glauconite. It generally contains less fragmental carbonaceous material than the underlying Cretaceous sediments (Otton, 1955). The Brightseat Formation is bounded by unconformities with a distinct gamma log signature that is useful for stratigraphic correlation (Hansen, 1996).

The Late Paleocene Aquia Formation was formerly identified as a greensand due to the ubiquitous occurrence of glauconite. This formation is a poorly to well sorted, variably shelly, and glauconitic quartz sand that contains calcareous cemented sandstone and shell beds. The Aquia Formation was deposited on a shoaling marine shelf that resulted in a coarsening upward lithology. This unit has been identified in the Virginia Coastal Plain and underlies all of Calvert County and most of St. Mary's County, Maryland (Hansen, 1996). The Aquia Formation forms an important aquifer as discussed in Section 2.4.12.

The Late Paleocene Marlboro Clay was formerly considered to be a lower part of the early Eocene Nanjemoy Formation but is now recognized as a widely distributed formation. The Marlboro Clay extends approximately 120 mi (193 km) in a northeast-southwest direction from the Chesapeake Bay near Annapolis, Maryland to the James River in Virginia. Micropaleontological data indicate a late Paleocene age although the Eocene-Paleocene boundary may occur within the unit (Hansen, 1996). The Marlboro Clay is one of the most distinctive stratigraphic markers of the Coastal Plain in Maryland and Virginia. It consists chiefly of reddish brown or pink soft clay that changes to a gray color in the subsurface of southern St. Mary's and Calvert Counties. Its thickness ranges from 40 ft (12 m) in Charles County to about 2 ft (60 cm) in St. Mary's County (Otton, 1955). However, the thickness is relatively constant from Anne Arundel County south through the CCNPP site to Solomons and Lexington Park, Maryland (Figure 2.5.1-12). The apparent localized thickening in Charles County might represent a local depocenter rather than a broader downwarping of the Salisbury Embayment relative to the Norfolk Arch (Figure 2.5.1-11).

The lower part of the overlying Early Eocene Nanjemoy Formation is predominantly a pale-gray to greenish gray, glauconitic very fine muddy sand to sandy clay. This formation becomes coarser upward from dominantly sandy silts and clays to dominantly clayey sands. The gradational contact between the two parts of the Nanjemoy is defined on the basis of geophysical log correlations (Hansen, 1996). In southern Maryland the Nanjemoy Formation ranges in thickness from several ft in its outcrop belt to as much as 240 ft (73 m) in the subsurface in St Mary's County (Otton, 1955) (Figure 2.5.1-12).

The Middle Eocene Piney Point Formation was recognized (Otton, 1955) as a sequence of shelly glauconitic sands underlying the Calvert Formation in southern Calvert County. The contact with the underlying Nanjemoy Formation is relatively sharp on geophysical logs, implying a depositional hiatus or unconformity (Hansen, 1996). The Piney Point Formation ranges in thickness from 0 ft (0 m) in central Calvert County to about 90 ft (27 m) at Point Lookout at the confluence of the Potomac River and Chesapeake Bay (Hansen, 1996). The Piney Point Formation contains distinctive carbonate-cemented interbeds of sand and shelly sand that range up to about 5 ft (1.5 m) in thickness (Hansen, 1996) and a characteristic fauna belonging to the Middle Eocene Jackson Stage (Otton, 1955). This unit is recognizable in the subsurface in Charles, Calvert, St. Marys, Dorchester, and Somerset Counties in Maryland and in Northumberland and Westmoreland Counties in Virginia but has not been recognized at the surface (Otton, 1955).

The work of several investigators were summarized (Hansen, 1996) who identified a 1 to 4 ft (30 to 122 cm) thick interval of clayey, slightly glauconitic, fossiliferous olive-gray, coarse sand containing fine pebbles of phosphate. This thin interval of late Oligocene (?) age occurs near the top of the Piney Point Formation and appears to correlate with the Old Church Formation in Virginia. This formation appears to thicken downdip between Piney Point and Point Lookout (Hansen, 1996). The absence of middle Oligocene deposits in most of the CCNPP site region indicates possible emergence or non-deposition during this time interval. Erosion or non-deposition during this relatively long interval of time produced an unconformity on the top of the Piney Point Formation that is mapped as a southeast dipping surface in the CCNPP site vicinity (Figure 2.5.1-13).

Renewed downwarping within the Salisbury Embayment resulted in marine transgression across older Cretaceous and Eocene deposits in Southern Maryland. The resulting Miocene-age Chesapeake Group consists of three marine formations; from oldest to youngest these are the Calvert, Choptank and St. Marys Formations. The basal member of the group, the Calvert Formation, is exposed in Anne Arundel, Calvert, Prince Georges, St. Mary's and Charles Counties. Although these formations were originally defined using biostratigraphic data, they are difficult to differentiate in well logs (Hansen, 1996) (Glaser, 2003a). The basal sandy beds are generally 10 to 20 ft (3 to 6 m) thick and consist of yellowish green to greenish light gray, slightly glauconitic fine to medium, quartz sand. The basal beds unconformably overlie older Oligocene and Eocene units and represent a major early Miocene marine transgression (Hansen, 1996). The overlying Choptank and St. Marys formations are described in greater detail in Section 2.5.1.2.3.

The Upper Miocene Eastover Formation and the Lower to Upper Pliocene Yorktown Formation occur in St. Mary's County and to the south in Virginia (McCartan, 1989b) (Ward, 2004). These units appear to have not been deposited to the north of St. Mary's County and that portion of the Salisbury Embayment may have been emergent (Ward, 2004).

2.5.1.1.3.1.4 Plio-Pleistocene and Quaternary Deposits

Surficial deposits in the Coastal Plain consist, in general, of two informal stratigraphic units: the Pliocene-age Upland deposits and the Pleistocene to Holocene Lowland deposits. These deposits are mapped (Mc Cartan, 1989a) (MaCartan, 1989b) as two units of Upper Pliocene fluvial Upland Gravels. It was recognized (McCartan, 1989b) that an Upper Pliocene sand with gravel cobbles and boulders that blankets topographically high areas in the southeast third of St. Mary's County. The Upland Deposits are areally more extensive in St. Mary's County than in Calvert County (Glaser, 1971). The map pattern has a dendritic pattern and since it caps the higher interfluvial divides, this unit is interpreted as a highly dissected sediment sheet whose base slopes toward the southwest (Glaser, 1971) (Hansen, 1996). This erosion might have occurred due to differential uplift during the Pliocene or down cutting in response to lower base levels when sea level was lower during period of Pleistocene glaciation.

McCartan (1989b) differentiates three Upper Pleistocene estuarine deposits, Quaternary stream terraces, Holocene alluvial deposits and colluvium in St. Mary's County. The Lowland deposits in southern Maryland were laid down in fluvial to estuarine environments (Hansen, 1996) and are generally found along the Patuxent and Potomac River valleys and Chesapeake Bay. These deposits occur in only a few places along the eastern shore of Chesapeake Bay. The Lowland deposits extend beneath Chesapeake Bay and the Potomac River filling deep, ancestral river channels with 200 ft (61 m) or more of fluvial or estuarine sediments (Hansen, 1996). These deep channels and erosion on the continental slope probably occurred during periods of glacial advances and lower sea levels. Deposition most likely occurred as the glaciers retreated and melt waters filled the broader ancestral Susquehanna and Potomac Rivers.

2.5.1.1.3.2 Piedmont Physiographic Province

There are two distinct divisions to the rocks of the Piedmont Physiographic Province. The first is a set of predominantly Late Precambrian and Paleozoic age crystalline rocks and the second is a set of Early Mesozoic (Triassic) age sedimentary rocks deposited locally in down-faulted basins within the crystalline rocks (Section 2.5.1.1.1) (Fichter, 2000) (Figure 2.5.1 5a, Figure 2.5.1-5b, and Figure 2.5.1-9).

2.5.1.1.3.2.1 Crystalline Rocks (Late Precambrian and Paleozoic)

Crystalline rocks of the Piedmont Province primarily occur within the Piedmont Upland section. The crystalline rocks consist of deformed and metamorphosed meta-sedimentary, metaigneous, and meta-volcanic rocks intruded by mafic dikes and granitic plutons (Markewich, 1990). The rocks belong to a number of northeast-trending belts that are defined on the basis of rock type, structure and metamorphic grade (Bledsoe, 1980) and are interpreted to have formed along and offshore of ancestral North America (Pavlides, 1994). From east to west the main lithotectonic belts are: the Goochland-Raleigh belt; the Carolina and Eastern slate belts; the Chopawamsic and Milton belts; and the Western/Inner Piedmont belt (Bledsoe, 1980) (Fichter, 2000) (Figure 2.5.1-8). The stratigraphy of the crystalline rock in these lithotectonic belts are discussed in the following paragraphs.

2.5.1.1.3.2.1.1 Goochland-Raleigh Belt

The Goochland-Raleigh belt stretches southward from Fredericksburg, Virginia, to the North Carolina state line east of the Spotsylvania fault (presented in Section 2.5.1.1.4.4.2) (Frye, 1986) (Figure 2.5.1-8). The Goochland belt (Virginia) is composed predominantly of granulite facies (high grade) metamorphic rocks and the Raleigh belt (North Carolina) is composed of sillimanite (very high grade) metamorphic rocks (Fichter, 2000). The Goochland-Raleigh belt is interpreted to be a microcontinent that was accreted to ancestral North America during the Taconic orogeny. Some geologists believe that the micro-continent was rifted from ancestral North America during the proto-Atlantic rifting while others believe that it formed outbound of ancestral North America (exotic or suspect terrane). Rocks of the Goochland-Raleigh belt are considered to be the oldest rocks of the Piedmont Province and bear many similarities to the Grenville age rocks of the Blue Ridge Province (Spears, 2002).

The Po River Metamorphic Suite and the Goochland terrane, that lie southeast of the Spotsylvania fault, make up the easternmost part of the Goochland-Raleigh belt. The Po River Metamorphic Suite was named after the Po River in the Fredericksburg area and comprises amphibolite grade (high grade) metamorphic rocks, predominantly biotite gneiss and lesser amounts of hornblende gneiss and amphibolite (Pavlides, 1989). The age of this unit is uncertain, but it has been assigned a provisional age of Precambrian to Early Paleozoic (Pavlides, 1980). The Goochland terrane was first studied along the James River west of Richmond, Virginia, and contains the only dated Precambrian rocks east of the Spotsylvania fault. It is a Precambrian granulite facies (high grade) metamorphic terrane.

2.5.1.1.3.2.1.2 Carolina Slate and Eastern Slate Belts

The Carolina Slate belt extends southward from southern Virginia to central Georgia, while the Eastern Slate belt is located predominantly in North Carolina, east of the Goochland-Raleigh belt (Figure 2.5.1-8). Both the Carolina and Eastern Slate belts are composed of greenschist facies (low grade) metamorphic rocks (Fichter, 2000), including meta-graywacke, tuffaceous argillites, quartzites, and meta-siltstones (Bledsoe, 1980). The Carolina and Eastern Slate belts are interpreted to be island-arcs that were accreted to ancestral North America during the Taconic orogeny. The island-arcs are interpreted to have been transported from somewhere in the proto-Atlantic Ocean, and are therefore considered to be exotic or suspect terranes. Rocks of the Carolina and Eastern Slate belts generally are considered to be Early Paleozoic in age. Granitic and gabbro-rich plutons that intrude the belts generally are considered to be Middle to Late Paleozoic in age (Bledsoe, 1980).

2.5.1.1.3.2.1.3 Chopawamsic Belt, including Milton and Charlotte Belts

The Chopawamsic belt, and its southeastward extensions, the Milton and Charlotte belts comprise a broad central part of the Piedmont Province from Virginia to Georgia (Figure 2.5.1-8). The belt is interpreted to be part of an island-arc and consist predominantly of meta-sedimentary and meta-volcanic rocks.

The Chopawamsic belt, also referred to as the "Chopawamsic Volcanic Belt" (Bailey, 1999) and the "Central Virginia Volcanic-Plutonic Belt (Rader, 1993) takes its name from exposures along Chopawamsic Creek in northern Virginia. The belt trends northeastward from the North Carolina state line, crosses the James River between Richmond and Charlottesville and continues northeastward to south of Washington D.C., where it is covered by Coastal Plain deposits. The Chopawamsic belt is bounded on the west by the Chopawamsic fault and on the east by the Spotsylvania fault (Section 2.5.1.1.4). The Chopawamsic belt is interpreted to be an island-arc that was accreted to ancestral North America during the Taconic orogeny (Figure 2.5.1-7). The Chopawamsic belt is regarded as an exotic or suspect terrain. Rocks in the Chopawamsic belt are Early Paleozoic in age. Recent U-Pb studies consistently yield Ordovician ages for Chopawamsic volcanic rocks and Rb-Sr and U-Pb dating of granite rocks give late Ordovician ages (Spears, 2002).

The Chopawamsic belt is comprised of the Chopawamsic Formation and the Ta River Metamorphic Suite. The Chopawamsic Formation and the Ta River Metamorphic Suite are interpreted to have formed as an island-arc. The Chopawamsic Formation is interpreted to have formed as the continent-ward side of the island-arc and the Ta River Metamorphic Suite as the ocean-ward side (Pavlides, 2000). The Chopawamsic Formation consists of a sequence of felsic, intermediate and mafic meta-volcanic rocks with subordinate meta-sedimentary rocks. The Ta River Metamorphic Suite consists of a sequence of amphibolites and amphibole-bearing gneisses with subordinate ferruginous quartzites and biotite gneiss. Rocks of the Ta River Metamorphic Suite are generally thought to be more mafic and to have experienced higher-grade regional metamorphism than the rocks of the Chopawamsic Formation (Spears, 2002).

The Chopawamsic Formation and Ta River Metamorphic Suite are unconformably overlain by the Quantico and Arvonia Formations. The Quantico and Arvonia Formations consist of meta-sedimentary rocks including slates, phyllites, schists, and quartzites. These meta-sedimentary rocks are considered to have been deposited in successor basins after the subjacent terranes were eroded and formed depositional troughs. Rocks of the Arvonia Formation are exposed in the Arvonia and Long Island synclines, while rocks of the Quantico Formation are exposed in the Quantico syncline. Rocks of the Arvonia, Long Island, and Quantico synclines form three belts across the central Virginia Piedmont, the Quantico synclines to the southeast and the Arvonia and Long Island synclines to the north (Spears, 2002).

The Chopawamsic Formation and the Ta River Metamorphic Suite are intruded by a number of granite plutons. The number of plutons and their relation to one another, however, remains uncertain (Spears, 2002). Rocks of the Falmouth Intrusive Suite intrude the Ta River Metamorphic Suite and Quantico Formation in the form of dikes, sills, and small irregular intrusions (Pavlides, 1980).

2.5.1.1.3.2.1.4 Western/Inner Piedmont Belt/Baltimore Terrane

The Western Piedmont belt, referred to as the Inner Piedmont belt in some publications, extends southward from Pennsylvania, where it has been designated as part of the Baltimore Gneiss and Glenarm Group (Baltimore terrane) through North Carolina and into Georgia (Figure 2.5.1-8). It is composed of greenschist facies (low grade) and amphibolite facies (high grade) meta-sedimentary rocks. These meta-sedimentary rocks enclose blocks of meta-basalt, ultramafic rocks, granite and other quasi-exotic lithologies and are called mélanges (Pavlides, 2000). These mélanges are interpreted to have formed in a Cambrian-Ordovician back-arc or marginal basin that lay on the continent-ward side of an island-arc terrane (Pavlides, 1989). The Baltimore terrane, a Middle Proterozoic metamorphosed sequence of felsic to intermediate rocks (Horton, 1989), consists of the Baltimore Gneiss and its cover sequence, the Glenarm Group, which consists of the basal Setters Formation, the Cockeysville

Marble and the pelitic Loch Raven Schist. Mineral assemblages within the Glenarm Group indicate that it was metamorphosed during the Paleozoic (Horton, 1989). The Potomac terrane (not shown on Figure 2.5.1-8 due to scale) was thrust upon the Baltimore terrane during the Taconic orogeny.

Two distinct types of mélange deposits occur within a collage of thrust slices in the Western Piedmont belt. The first type is a block-in-phyllite mélange that constitutes the Mine Run Complex of Virginia. It consists of a variety of meta-plutonic, meta-volcanic, mafic, and ultramafic blocks enclosed within a matrix of phyllite or schist and meta-sandstones of feldspathic or quartz meta-graywacke. The Mine Run complex is interpreted to consist of four imbricated thrust slices, each with its own distinctive exotic block content (Pavlides, 1989).

The second mélange type within the Western Piedmont belt is a meta-diamictite and contains a less extensive variety of exotic blocks, the most common being mafic and ultramafic blocks. The exotic blocks are enclosed in a micaceous quartzofeldspathic matrix, which has contemporaneously deposited schist and quartz-lump fragments as its characterizing features. Several varieties of meta-diamictite have been recognized in Virginia and described as the Lunga Reservoir and Purcell Branch Formations (Pavlides, 1989).

The mélanges of the Western Piedmont are overlain unconformably by Ordovician age meta-sedimentary rocks and are intruded by Ordovician age and Late Ordovician or Early Silurian age felsic plutons, such as the Lahore and Ellisville plutons (Pavlides, 1989).

2.5.1.1.3.2.1.5 Ijamsville Belt/Westminster Terrane

The Ijamsville-Pretty Boy-Octoraro terrane is more currently known as the Westminster terrane (Horton, 1989). This belt consists of pelitic schist or phyllite characterized by albite porphyroblasts and a green and purple phyllite unit. Rocks of the Ijamsville/Westminster terrane were interpreted to comprise a tectonic assemblage of undated rocks of the rise and slope deepwater deposits of the Iapetus Ocean that were thrust onto the Grenville-age Blue Ridge Province along the Martic overthrust during the Taconic orogeny (Drake, 1989) (Horton, 1989).

2.5.1.1.3.2.2 Sedimentary Rocks (Early Mesozoic)

Mesozoic sedimentary rocks of the Piedmont Province occur primarily within the Piedmont Lowland section (Figure 2.5.1-9). The sediments were deposited in a series of northeast-trending basins described below in Section 2.5.1.1.4.4.3. Sediments filling the basins include intermontane fanglomerates, fresh-water limestone, mudstones, siltstones and sandstones, and basic igneous intrusive dikes and sills and lava flows (Markewich, 1990). The Lower Mesozoic sediments deposited in these basins usually are referred to as Triassic basin deposits, although the basins are now known to also contain Lower Jurassic rocks.

2.5.1.1.3.2.3 Surficial Sediments (Cenozoic)

Surficial sediments in the Piedmont Province consist of residual and transported material. The residual soils have developed in place from weathering of the underlying rocks, while the transported material – alluvium and colluvium – has been moved by water or gravity and deposited as unconsolidated deposits of clay, silt, sand, and gravel (Carter, 1976). Surficial sediments in the Piedmont Upland section are interpreted to be the product of Cenozoic weathering, Quaternary periglacial erosion and deposition, and recent anthropogenic activity (Sevon, 2000).

Residual soil in the Piedmont Province consists of completely decomposed rock and saprolite. Residual soils occur almost everywhere, except where erosion has exposed the bedrock on ridges and in valley bottoms. Saprolite comprises the bulk of residual soil in the Piedmont

Province and is defined as an earthy material in which the major rock-forming minerals (other than quartz) have been altered to clay but the material retains most of the textural and structural characteristics of the parent rock. The saprolite forms by chemical weathering, its thickness and mineralogy being dependent on topography, parent rock lithology and the presence of surface and/or groundwater (Cleaves, 2000).

Relief affects the formation of soils by causing differences in internal drainage, runoff, soil temperatures, and geologic erosion. In steep areas where there is rapid runoff, little percolation of water through the soil and little movement of clay, erosion is severe and removes soil as rapidly as it forms. Gently sloping areas, on the other hand, are well drained and geologic erosion in these areas is generally slight. The characteristics of the underlying rock strongly influence the kind of changes that take place during weathering. Because of differences in these characteristics, the rate of weathering varies for different rock types. The igneous, metamorphic and sedimentary rocks of the Piedmont Province are all sources of parent material for the soils.

Colluvium in the Piedmont Province occurs discontinuously on hilltops and side slopes, while thicker colluvium occurs in small valleys lacking perennial streams. Alluvium is present in all valleys with perennial streams (Sevon, 2000).

2.5.1.1.3.3 Blue Ridge Physiographic Province

The Blue Ridge Physiographic Province is underlain by a broad, northeast-trending, structurally complex metamorphic terrane (Mixon, 2000). In the site region, the Blue Ridge occurs southward from south-central Pennsylvania through Virginia (Figure 2.5.1-1). The Blue Ridge terrain consists of stratified meta-sedimentary rocks and meta-basalts of Early Paleozoic and Late Precambrian age and an underlying gneissic and granitic basement-rock complex of Middle to Late Precambrian age (Figure 2.5.1-5a and Figure 2.5.1-5b).

2.5.1.1.3.4 Valley and Ridge Physiographic Province

The Valley and Ridge Physiographic Province is underlain primarily by layered sedimentary rock that has been intensely folded and locally thrust faulted. The sedimentary rocks range in age from Cambrian to Pennsylvanian. The valley areas within the Great Valley (Figure 2.5.1-6) are underlain predominantly by thick sequences of limestone, dolomite and shale. The upland areas of the Valley and Ridge Province (Appalachian Mountains) to the west are underlain predominantly by resistant sandstones and conglomerates, while the lowland areas are underlain predominantly by less resistant shale, siltstone, sandstone and limestone (Colton, 1970) (Figure 2.5.1-5a and Figure 2.5.1-5b).

2.5.1.1.3.5 Appalachian Plateau Physiographic Province

The Appalachian Plateau Physiographic Province is underlain by rocks that are continuous with those of the Valley and Ridge Province, but in the Appalachian Plateau the layered rocks are nearly flat-lying or gently tilted and warped, rather than being intensely folded and faulted. Rocks of the Allegheny Front along the eastern margin of the province consist of thick sequences of sandstone and conglomerate, interbedded with shale, ranging in age from Devonian to Pennsylvanian. Rocks of the Appalachian Plateau west of the Allegheny Front are less resistant and consist of Permian age sandstone, shale and coal (Lane, 1983) (Hack, 1989) (Figure 2.5.1-5a and Figure 2.5.1-5b).}

2.5.1.1.4 Regional Tectonic Setting

{In 1986, the Electric Power Research Institute (EPRI) developed a seismic source model for the Central and Eastern United States (CEUS), which included the CCNPP site region

(EPRI, 1986). The CEUS is a stable continental region (SCR) characterized by low rates of crustal deformation and no active plate boundary conditions. The EPRI source model included the independent interpretations of six Earth Science Teams and reflected the general state of knowledge of the geoscience community as of 1986. The seismic source models developed by each of the six teams were based on the tectonic setting and the occurrence, rates, and distribution of historical seismicity. The original seismic sources identified by EPRI (1986) are thoroughly described in the EPRI study reports (EPRI, 1986) and are summarized in Section 2.5.2.2.

Since 1986, additional geological, seismological, and geophysical studies have been completed in the CEUS and in the CCNPP site region. The purpose of this section is to summarize the current state of knowledge on the tectonic setting and tectonic structures in the site region and to highlight new information acquired since 1986 that is relevant to the assessment of seismic sources.

A global review of earthquakes in SCRs shows that areas of Mesozoic and Cenozoic extended crust are positively correlated with large SCR earthquakes. Nearly 70% of SCR earthquakes with $M \geq 6$ occurred in areas of Mesozoic and Cenozoic extended crust (Johnston, 1994). Additional evidence shows an association between Late Proterozoic rifts and modern seismicity in eastern North America (Johnston, 1994) (Wheeler, 1995) (Ebel, 2002). Paleozoic and Mesozoic extended crust underlies the entire 200 mi (322 km) CCNPP site region (Figure 2.5.1-14). However, as discussed in this section, there is no evidence for late Cenozoic seismogenic activity of any tectonic feature or structure in the site region (Crone, 2000) (Wheeler, 2005). Although recent characterization of several tectonic features has modified our understanding of the tectonic evolution and processes of the mid-Atlantic margin, no structures or features have been identified in the site region since 1986 that show clear evidence of seismogenic potential greater than what was recognized and incorporated in the EPRI study (EPRI, 1986) seismic source model.

The following sections describe the tectonic setting of the site region by discussing the: (1) plate tectonic evolution of eastern North America at the latitude of the site, (2) origin and orientation of tectonic stress, (3) gravity and magnetic data and anomalies, (4) principal tectonic features, and (5) seismic sources defined by regional seismicity. Historical seismicity occurring in the site region is described in Section 2.5.2.1. The geologic history of the site region was discussed in Section 2.5.1.1.2.

2.5.1.1.4.1 Plate Tectonic Evolution of the Atlantic Margin

The Late Precambrian to Recent plate tectonic evolution of the site region is summarized in Section 2.5.1.1.2 and in Figure 2.5.1-7. Most of the present-day understanding of the plate tectonic evolution comes from research performed prior to the 1986 EPRI report (EPRI, 1986). Fundamental understanding about the timing and architecture of major orogenic events was clear by the early 1980's, after a decade or more of widespread application of plate tectonic theory to the evolution of the Appalachian orogenic belt (e.g., (Rodgers, 1970) (Williams, 1983)). Major advances in understanding of the plate tectonic history of the Atlantic continental margin since the EPRI study report (EPRI, 1986) include the organization of lithostratigraphic units and how they relate to the timing and kinematics of Paleozoic events (e.g., Hatcher, 1989) (Hibbard, 2006) (Hibbard, 2007) and the refinement of the crustal architecture of the orogen and passive margin (e.g., (Hatcher, 1989) (Glover, 1995b) (Klitgord, 1995)).

The following subsections divide the regional plate tectonic history into: (1) Late Proterozoic and Paleozoic tectonics and assembly of North American continental crust, (2) Mesozoic rifting and passive margin formation, and (3) Cenozoic vertical tectonics associated with exhumation, deposition, and flexure.

2.5.1.1.4.1.1 Late Proterozoic and Paleozoic Plate Tectonic History

Although details about the kinematics, provenance, and histories of lithostratigraphic units within the Appalachian orogenic belt continue to be debated and reclassified (e.g., (Hatcher, 1989) (Horton, 1991) (Glover, 1995b) (Hibbard, 2006)), it is well accepted that plate boundary deformation has occurred repeatedly in the site region since late Precambrian time. Suturing events that mark the welding of continents to form supercontinents and rifting events that mark the breakup of supercontinents to form ocean basins have each occurred twice during this interval. Foreland strata, deformation structures, and metamorphism associated with the Grenville (Middle Proterozoic) and Allegheny (Late Paleozoic) orogenies record the closing of ocean basins and welding of continents to form the supercontinents Rodinia and Pangaea, respectively (Figure 2.5.1-7). Synrift basins, normal faults, and postrift strata associated with the opening of the Iapetus (Late Proterozoic to Early Cambrian) and Atlantic (Early Mesozoic) Ocean basins record the break-up of the supercontinents. The principal structures that formed during the major events are salient to the current seismic hazards in that: (1) they penetrate the seismogenic crust, (2) they subdivide different crustal elements that may have contrasting seismogenic potential, and (3) their associated lithostratigraphic units make up the North American continental crust that underlies most of the site region. Many of the principal structures are inherited faults that have been reactivated repeatedly through time. Some are spatially associated with current zones of concentrated seismic activity and historical large earthquakes. For example, the 1811-1812 New Madrid earthquake sequence ruptured a failed Late Proterozoic rift that also may have been active in the Mesozoic (Ervin, 1975).

During the interval between opening of the Iapetus Ocean and opening of the Atlantic Ocean, the eastern margin of the ancestral North America continent was alternately (1) an active rift margin accommodating lithospheric extension with crustal rift basins and synrift strata and volcanism; (2) a passive continental margin accumulating terrestrial and shallow marine facies strata; and (3) an active collisional margin with accretion of microcontinents, island arcs, and eventually the African continent. Major Paleozoic mountain building episodes associated with the collision and accretion events included the Taconic, Acadian, and Allegheny Orogenies. More localized collisional events in the site region include the Avalon, Virgilina and Potomac (Penobscot) orogenies (Hatcher, 1987) (Hatcher, 1989) (Glover, 1995b) (Hibbard, 1995) (Drake, 1999) (Figure 2.5.1-7). The geologic histories of these orogenies are described in Section 2.5.1.1.2.

Tectonic structures developed during the interval between the Late Proterozoic and Triassic Periods are variable in sense of slip and geometry. Late Proterozoic and early Cambrian rifting associated with the breakup of Rodinia and development of the Iapetus Ocean formed east-dipping normal faults through Laurentian (proto-North American) crust (Figure 2.5.1-15 and Figure 2.5.1 16). Late Proterozoic extended crust of the Iapetan margin probably underlies the Appalachian fold belt southeastward to beneath much of the Piedmont Province (Wheeler, 1996). Paleozoic compressional events associated with the Taconic, Acadian, and Allegheny orogenies formed predominantly west-vergent structures that include (1) Valley and Ridge Province shallow folding and thrusting within predominantly passive margin strata, (2) Blue Ridge Province nappes of Laurentian crust overlain by Iapetan continental margin deposits, (3) Piedmont Province thrust-bounded exotic and suspect terranes including island arc and accretionary complexes interpreted to originate in the Iapetan Ocean, and (4) Piedmont Province and sub-Coastal Plain Province east-dipping thrust, oblique, and reverse fault zones that collectively are interpreted to penetrate much of the crust and represent major sutures that juxtapose crustal elements (Hatcher, 1987) (Horton, 1991) (Glover, 1995b) (Hibbard, 2006) (Figure 2.5.1-15 and Figure 2.5.1-16). Many investigators recognize significant transpressional components to major faults bounding lithostratigraphic units (Hatcher, 1987) (Glover, 1995b) (Hibbard, 2006) (Figure 2.5.1-7 and Figure 2.5.1-15).

2.5.1.1.4.1.2 Mesozoic and Cenozoic Passive Margin Evolution

At the time of the EPRI (1986) study much was published about the structure and crustal elements of the Mesozoic to Cenozoic Atlantic passive margin (e.g., (Klitgord, 1979)). However, it was not until the Geological Society of America's Decade of North American Geology (DNAG) volume on the U.S. Atlantic continental margin (Sheridan, 1988), seminal papers within it (e.g., (Klitgord, 1988)), and later summary publications (e.g., (Klitgord, 1995) (Withjack, 1998)) that the current understanding of the margin structure and tectonic history was formulated comprehensively.

The current Atlantic passive continental margin has evolved since rifting initiated in the Early Triassic. The progression from active continental rifting to sea-floor spreading and a passive continental margin included: (1) initial rifting and hot-spot plume development, (2) thinning of warm, buoyant crust with northwest-southeast extension, normal faulting and deposition of synrift sedimentary and volcanic rocks, and (3) cooling and subsidence of thinned crust and deposition of postrift sediments on the coastal plain and continental shelf, slope, and rise (Klitgord, 1988) (Klitgord, 1995). The transition between the second (rifting) and third (drifting) phases during the Early Jurassic marked the initiation of a passive margin setting in the site region, in which active spreading migrated east away from the margin. As the thinned crust of the continental margin cooled and migrated away from the warm, buoyant crust at the mid-Atlantic spreading center, horizontal northwest-southeast tension changed to horizontal compression as gravitational potential energy from the spreading ridge exerted a lateral "ridge push" force on the oceanic crust. Northwest-southeast-directed postrift shortening, manifested in Mesozoic basin inversion structures, provides the clearest indication of this change in stress regime (Withjack, 1998). The present-day direction of maximum horizontal compression—east-northeast to west-southwest—is rotated from this hypothesized initial postrift direction.

The crustal structure of the passive continental margin includes areas of continental crust, (Iapetan-extended crust (Wheeler, 1996)), rifted continental crust, rift-stage (transitional) crust, marginal oceanic crust, and oceanic crust (Klitgord, 1995) (Figure 2.5.1-17 and Figure 2.5.1-18). Rifted continental crust is crust that has been extended, faulted, and thinned slightly. In the site region, rifted-continental crust extends from the western border faults of the exposed synrift Danville, Scottsville, Culpeper, Gettysburg, and Newark basins to the basement hinge zone, approximately coincident with the seaward edge of the continental shelf (Klitgord, 1995) (Figure 2.5.1-11 and Figure 2.5.1-18). Rifted crust also includes exposed and buried Upper Triassic to Lower Jurassic basins within the eastern Piedmont and Coastal Plain Provinces, including the Richmond, Taylorsville, and Norfolk basins (Figure 2.5.1-9). Several additional basins with poorly defined extent also underlie the Coastal Plain and Continental Shelf and are shown directly east and northeast of the site (Figure 2.5.1-9). Buried synrift basins are delineated based on sparse drillhole data, magnetic and gravity anomalies, and seismic reflection data (e.g., (Benson, 1992)). Figure 2.5.1-18 shows east-dipping basin-bounding faults that penetrate the seismogenic crust and have listric geometries at depth. Many of the synrift normal faults are interpreted as Paleozoic thrust faults reactivated during Mesozoic rifting. The Mesozoic basins are discussed further in Section 2.5.1.1.4.4.3 as well as the hypothesized Queen Anne basin shown as lying beneath the site (Figure 2.5.1 9).

Rift-stage (transitional) crust is extended continental crust intruded by mafic magmatic material during rifting. In the site region, this crustal type coincides with the basement hinge zone and postrift Baltimore Canyon Trough (Klitgord, 1995) (Figure 2.5.1-11). The basement hinge zone is defined where pre-Late Jurassic basement abruptly deepens seaward from about 1 to 2.5 mi (1.6 to 4 km) to more than 5 mi (8 km) . Overlying this lower crustal unit seaward of the basement hinge zone is the Jurassic volcanic wedge, representing a period of excess volcanism and is greater than 65 mi (105 km) wide and 1 to 5 mi (1.5 to 8 km) thick. The wedge is identified on seismic reflection lines as a prominent sequence of seaward-dipping reflectors. The East Coast magnetic anomaly (ECMA) coincides with the seaward edge of the wedge (Figure 2.5.1-17) (Section 2.5.1.1.4.3.2).

The last transitional crustal unit between continental and oceanic crust is marginal oceanic crust (Klitgord, 1995) (Figure 2.5.1-17). Marginal oceanic crust is located east of the ECMA where the Jurassic volcanic wedge merges with the landward edge of oceanic crust. Here, the transition from rifting to sea-floor spreading created a thicker than normal oceanic crust with possible magmatic underplating.

A postrift unconformity separates synrift from postrift deposits and represents the change in tectonic regime in the Middle Jurassic from continental rifting to the establishment of the passive margin ("drifting"). Sedimentary rocks below the unconformity are cut by numerous faults. In contrast, the rocks and strata above the unconformity accumulated within the environment of a broadly subsiding passive margin and are sparsely faulted. Sediments shed from the faulted blocks of the rifting phase and from the core of the Allegheny orogen accumulated on the coastal plain, continental shelf, slope, and rise above the postrift unconformity and contributed to subsidence of the cooling postrift crust by tectonic loading.

Postrift deformation is recorded in synrift basins and within postrift strata as normal faults seaward of the basement hinge zone and as contractional features landward of the basement hinge zone. Extensive normal faulting penetrates the postrift strata (and upper strata of the volcanic wedge) of the marginal basin overlying the volcanic wedge (Figure 2.5.1-17 and Figure 2.5.1-18). This set of faults is thought to have been caused by sediment loading on the outer edge of the margin due to differential compaction of the slope-rise deposits relative to adjacent carbonate platform deposits (Poag, 1991) (Klitgord, 1995). These faults are interpreted as margin-parallel structures that bound large mega-slump blocks and are not considered active tectonic features (Poag, 1991).

Schlische (2003) summarizes evidence for postrift shortening and positive basin inversion (defined as extension within basins followed by contraction) in several Atlantic margin basins, including the Newark, Taylorsville, and Richmond basins in the site region (Figure 2.5.1-9). Contractional postrift deformation is interpreted to record the change in stress regime from horizontal maximum extension during rifting to horizontal maximum compression during passive margin drifting. The hypothesis that the change in stress regime following rifting was recorded in reverse and strike slip faulting and folding was known prior to the 1986 EPRI study (e.g., (Sanders, 1963) (Swanson, 1982) (Wentworth, 1983)), but significant advances in the documentation and characterization of the rift to drift transition and postrift deformation has occurred since the mid-1980s (Withjack, 1998) (Schlische, 2003). Based on structural analysis and age control of basaltic dikes and faulting, much of the site region was under a state of northwest-southeast maximum compression by earliest Jurassic time (Withjack, 1998). This deformation regime may have persisted locally into the Cenozoic based on the recognized early Cenozoic contractional growth faulting associated with the northeast-striking Brandywine fault system (Jacobeen, 1972) (Wilson, 1990), Port Royal fault zone (Mixon, 1984) (Mixon, 2000) and Skinkers Neck anticline (Mixon, 1984) (Mixon, 2000) (Section 2.5.1.1.4.4.4). The present-day stress field of east-northeast to west-southwest maximum horizontal compression (Zoback, 1989a) is rotated from the hypothesized Jurassic and Cretaceous northwest-southeast orientation. The east-northeast to west-southwest maximum horizontal stress direction is consistent with resolved dextral transpressive slip locally documented on the northeast-striking Stafford fault system (Mixon, 2000), a recognized Tertiary tectonic feature (Section 2.5.1.1.4.4.4.1).

2.5.1.1.4.1.3 Cenozoic Passive Margin Flexural Tectonics

Tectonic processes along the Atlantic passive continental margin in the Cenozoic Era include vertical tectonics associated with lithospheric flexure. Vertical tectonics are dominated by: (1) cooling of the extended continental, transitional, and oceanic crust as the spreading center migrates eastward, and (2) the transfer of mass from the Appalachian core to the Coastal Plain and Continental Shelf, Slope, and Rise via erosion. Erosion and exhumation of the Allegheny crustal root of the Piedmont, Blue Ridge, Valley and Ridge, and Appalachian Plateau Provinces has been balanced by deposition on and loading of the Coastal Plain and offshore provinces by fluvial, fluvial-deltaic, and marine sediment transport. Margin-parallel variations in the amount of uplift and subsidence have created arches (e.g. South New Jersey and Norfolk Arches) and basins or embayments (e.g. Salisbury Embayment) along the Coastal Plain and Continental Shelf (Figure 2.5.1-11).

Flexural zones show both passive-margin-normal and passive-margin-parallel trends. Flexure normal to the passive margin is clearly recorded in the basement hinge zone (Figure 2.5.1-18). The vertical relief across the offshore basement hinge zone accounts for a change in postrift sediment thickness from 1 to 2.5 mi (1.6 to 4 km) to over 5 mi (8 km) and indicates lateral changes in tectonic loading (Klitgord, 1995). It has been proposed that the downwarping of the margin in the vicinity of the main depocenter of the Baltimore Canyon Trough led to the flexural uplift of the Coastal Plain units to the west (Watts, 1982). However, more recent studies show that sea-level variations since the Cretaceous are compatible with the present elevations of exposed Coastal Plain strata and thus do not support flexural uplift of the Coastal Plain (e.g., (Pazzaglia, 1993)).

A simple elastic model of Cenozoic flexural deformation across the Atlantic passive margin has been used to approximate the response of rifted continental crust to surface erosion of the Piedmont and deposition on the Coastal Plain and Continental Shelf (Pazzaglia, 1994) (Figure 2.5.1-11 and Figure 2.5.1-18). The boundary between areas of net Cenozoic erosion and deposition, the Fall Line, marks the flexural hinge between uplift and downwarping. Geologic correlation and longitudinal profiles of Miocene to Quaternary river terraces on the Piedmont with deltaic and marine equivalent strata on the Coastal Plain provide data for model validation (Pazzaglia, 1993). A one-dimensional elastic plate model replicates the form of the profiles and maintenance of the Fall Line with flexure driven by exhumation of the Piedmont and adjacent Appalachian provinces coupled with sediment loading in the Salisbury Embayment and Baltimore Canyon Trough (Pazzaglia, 1994). Model results suggest a long-term denudation rate of approximately 33 ft (10 m) per million years and about 115 to 426 ft (35 to 130 m) of upwarping of the Piedmont in the last 15 million years.

The flexural hinge zones (Fall Line and basement hinge zone) do not appear to be seismogenic. The spatial association between the Fall Line and observed Cenozoic faults such as the Stafford and Brandywine fault systems is commonly attributed to the fact that those faults are recognizable where Cenozoic cover is thin and there is greater exposure of bedrock compared to areas farther east toward the coast (e.g., (Wentworth, 1983)). It is suggested (Pazzaglia, 1994) that low rates of contractional deformation on or near the hinge zone documented on Cenozoic faults may be a second-order response to vertical flexure and horizontal compressive stresses. Neither the Fall Line nor basement hinge zone was considered a potential tectonic feature by EPRI (1986). They were considered zones where ground amplification could be affected. It is also suggested (Weems, 1998) that multiple fall lines (i.e., alignments of anomalously steep river gradients) located near or within the Fall Line may be of neo-tectonic origin. Subsequent studies performed during the North Anna ESP study demonstrates that the fall lines (Weems, 1998) are erosional features and not capable tectonic sources (NRC, 2005) (Section 2.5.1.1.4.4.5.1) Post-EPRI seismicity also shows no spatial patterns suggestive of seismicity aligned with either the basement hinge zone or Fall Line. Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2005) (Wheeler, 2006) also do not list these as potentially Quaternary active features. Accordingly, it is concluded that these features are not capable tectonic sources. Post-EPRI seismicity also shows no spatial patterns suggestive of seismicity aligned with either the basement hinge zone or Fall Line (Section 2.5.2).

Along-strike variations in the amount of epeirogenic movement along the Atlantic continental margin has resulted in a series of arches and embayments identified based on variations in thickness of Coastal Plain strata from Late Cretaceous through Pleistocene time. The Salisbury Embayment is a prominent, broad depocenter in the site region, and coincides with Chesapeake Bay and Delaware Bay (Figure 2.5.1-11). At the margins of the Salisbury Embayment are the South New Jersey Arch to the northeast and the Norfolk Arch to the south. Both arches are broad anticlinal warps reflected in the top of basement and overlying sediments. The processes that form and maintain the arches and embayments are poorly understood, and there has been little advancement in the thinking about these features since publication of the EPRI study report (EPRI, 1986). Poag (2004), however, use new basement data obtained from seismic reflection profiles and exploratory boreholes in the region of the main Chesapeake Bay impact crater to show that the Norfolk Arch is not as well expressed as originally interpreted by earlier authors (Brown, 1972) using limited data. Previous elevation differences cited as evidence for the basement arch appear to be due to subsidence differential between the impact crater and the adjacent deposits (Poag, 2004) (Section 2.5.1.1.4.4.4). Regardless, no published hypothesis was found suggesting causality between epeirogenic processes maintaining these specific arches and the embayment and potentially seismogenic structures, and there is no spatial association of seismicity with the basement arches. Thus, it is concluded that these features are not capable tectonic sources.

2.5.1.1.4.2 Tectonic Stress in the Mid-Continent Region

Expert teams that participated in the 1986 EPRI evaluation of intra-plate stress generally concluded that tectonic stress in the CEUS region is characterized by northeast-southwest-directed horizontal compression. In general, the expert teams concluded that the most likely source of tectonic stress in the mid-continent region was ridge-push force associated with the Mid-Atlantic ridge, transmitted to the interior of the North American plate by the elastic strength of the lithosphere. Other potential forces acting on the North American plate were judged to be less significant in contributing to the magnitude and orientation of the maximum compressive principal stress. Some of the expert teams noted that deviations from the regional northeast-southwest trend of principal stress may be present along the east coast of North America and in the New Madrid region. They assessed the quality of stress indicator data and discussed various hypotheses to account for what were interpreted as variations in the regional stress trajectories.

Since 1986, an international effort to collate and evaluate stress indicator data has resulted in publication of a new world stress map (Zoback, 1989a) (Zoback, 1989b). Data for this map are ranked in terms of quality, and plate-scale trends in the orientations of principal stresses are assessed qualitatively based on analysis of high-quality data (Zoback, 1992). Subsequent statistical analyses of stress indicators confirmed that the trajectory of the maximum compressive principal stress is uniform across broad continental regions at a high level of statistical confidence. In particular, the northeast-southwest orientation of principal stress in the CEUS inferred by the EPRI experts is statistically robust, and is consistent with the theoretical trend of compressive forces acting on the North American plate from the mid-Atlantic ridge (Coblentz and Richardson, 1995).

More recent assessments of lithospheric stress do not support inferences by some EPRI expert teams that the orientation of the principal stress may be locally perturbed in the New England area, along the east coast of the United States, or in the New Madrid region. A variety of data was summarized (Zoback, 1989a), including well-bore breakouts, results of hydraulic fracturing studies, and newly calculated focal mechanisms, which indicate that the New England and eastern seaboard regions of the U.S. are characterized by horizontal northeast-southwest to east-west compression. Similar trends are present in the expanded set of stress indicators for the New Madrid region. Zoback and Zoback (Zoback, 1989a) grouped all of these regions, along with a large area of eastern Canada, with the CEUS in an expanded "Mid-Plate" stress province characterized by northeast-southwest directed horizontal compression.

In addition to better documenting the orientation of stress, research conducted since 1986 has addressed quantitatively the relative contributions of various forces that may be acting on the North American plate to the total stress within the plate. Richardson and Reding (Richardson, 1991) performed numerical modeling of stress in the continental U.S. interior, and considered the contribution to total tectonic stress to be from three classes of forces:

- Horizontal stresses that arise from gravitational body forces acting on lateral variations in lithospheric density. These forces commonly are called buoyancy forces. Richardson and Reding emphasize that what is commonly called ridge-push force is an example of this class of force. Rather than a line-force that acts outwardly from the axis of a spreading ridge, ridge-push arises from the pressure exerted by positively buoyant, young oceanic lithosphere near the ridge against older, cooler, denser, less buoyant lithosphere in the deeper ocean basins (Turcotte, 2002). The force is an integrated effect over oceanic lithosphere ranging in age from about 0 to 100 million years (Dahlen, 1981). The ridge-push force is transmitted as stress to the interior of continents by the elastic strength of the lithosphere.
- Shear and compressive stresses transmitted across major plate boundaries (strike-slip faults and subduction zones).
- Shear tractions acting on the base of the lithosphere from relative flow of the underlying asthenospheric mantle.

Richardson and Reding (Richardson, 1991) concluded that the observed northeast-southwest trend of principal stress in the CEUS dominantly reflects ridge-push body forces. They estimated the magnitude of these forces to be about 2 to 3 X 1012 N/m (i.e., the total vertically integrated force acting on a column of lithosphere 1 m wide), which corresponds to average equivalent stresses of about 40 to 60 MPa distributed across a 30 mi (50 km) thick elastic plate. The fit of the model stress trajectories to data was improved by the addition of compressive stress (about 5 to 10 MPa) acting on the San Andreas Fault and Caribbean plate boundary structures. The fit of the modeled stresses to the data further suggested that shear stresses acting on these plate boundary structures is in the range of 5 to 10 MPa.

Richardson and Reding (Richardson, 1991) noted that the general northeast-southwest orientation of principal stress in the CEUS also could be reproduced in numerical models that assume a shear stress, or traction, acting on the base of the North American plate. Richardson and Reding (Richardson, 1991) and Zoback and Zoback (Zoback, 1989) do not favor this as a significant contributor to total stress in the mid-continent region. A basal traction predicts or requires that the horizontal compressive stress in the lithosphere increases by an order of magnitude moving east to west, from the eastern seaboard to the Great Plains. Zoback and Zoback (Zoback, 1989) noted that the state of stress in the southern Great Plains is characterized by north-northeast to south-southwest extension, which is contrary to this prediction. They further observed that the level of background seismic activity is generally higher in the eastern United States than in the Great Plains, which is not consistent with the prediction of the basal traction model that compressive stresses (and presumably rates of seismic activity) should be higher in the middle parts of the continent than along the eastern margin.

To summarize, analyses of regional tectonic stress in the CEUS since EPRI (1986) have not significantly altered the characterization of the northeast-southwest orientation of the maximum compressive principal stress. The orientation of a planar tectonic structure relative to the principal stress direction determines the magnitude of shear stress resolved onto the structure. Given that the current interpretation of the orientation of principal stress is similar to that adopted in EPRI (1986), a new evaluation of the seismic potential of tectonic features based on a favorable or unfavorable orientation to the stress field would yield similar results. Thus, there is no significant change in the understanding of the static stress in the CEUS since the publication of the EPRI source models in 1986, and there are no significant implications for existing characterizations of potential activity of tectonic structures.

2.5.1.1.4.3 Gravity and Magnetic Data and Features of the Site Region and Site Vicinity

Gravity and magnetic anomaly datasets of the site region have been published following the 1986 EPRI study. Significant datasets include regional maps of the gravity and magnetic fields in North America by the Geological Society of America (GSA), as part of the Society's DNAG project (Tanner, 1987) (Hinze, 1987). The DNAG datasets are widely available in digital form via the internet (Hittelman, 1994). A magnetic anomaly map of North America was published in 2002 that featured improved reprocessing of existing data and compilation of a new and more complete database (Bankey, 2002) (Figure 2.5.1-19).

These maps present the potential field data at 1:5,000,000-scale, and thus are useful for identifying and assessing gravity and magnetic anomalies with wavelengths on the order of tens of kilometers or greater (Bankey, 2000) (Hittelman, 1994). Regional gravity anomaly maps are based on Bouguer gravity anomalies onshore and free-air gravity anomalies offshore. The primary sources of magnetic data reviewed for this CCNPP Unit 3 study are from aeromagnetic surveys onshore and offshore (Bankey, 2002), and the DNAG datasets available digitally from the internet (Hittelman, 1994).

Most of the contributed gravity and magnetic data that went into the regional compilations were collected prior to the 1986 EPRI study; thus, most of the basic data were available for interpretation at local and regional scales. Large-scale compilations (1:2,500,000-scale) of the free-air anomalies offshore and Bouguer anomalies onshore were published in 1982 by the Society of Exploration Geophysicists (Lyons, 1982) (Sheridan, 1988). The DNAG magnetic anomaly maps were based on a prior analog map of magnetic anomalies of the U.S. published in the early 1980's (Zietz, 1982) (Behrendt, 1983) (Sheridan, 1988).

In addition, the DNAG Continent-Ocean transect program published a synthesis of gravity and magnetic data with seismic and geologic data (Klitgord, 1995). Transect E-3, which crosses the site region, is presented in Figure 2.5.1-15 and Figure 2.5.1-16. Much of the seismic and geophysical data through the Piedmont region was reanalyzed from a geophysical survey conducted along Interstate I-64 in Virginia that was published prior to release of the 1986 EPRI study (e.g., (Harris, 1982)).

In summary, the gravity and magnetic data published since 1986 do not reveal any new anomalies related to geologic structures that were not identified prior to the 1986 EPRI study. Rather, post-EPRI publications have refined the characteristics and tectonic interpretation of the anomalies. Discussion of the gravity and magnetic anomalies is presented in the following sections.

2.5.1.1.4.3.1 Gravity Data and Features

Gravity data compiled at 1:5,000,000-scale for the DNAG project provide documentation of previous observations that the gravity field in the site region is characterized by a long-wavelength, east-to-west gradient in the Bouguer gravity anomaly over the continental margin (Harris, 1982) (Hittelman, 1994) (Figure 2.5.1-20). The free-air gravity anomaly shows broad gravity lows over offshore oceanic crust near the continental margin and over the broad marginal embayments. Offshore marginal platforms are marked by shorter-wavelength, higher-amplitude gravity highs and lows. The present shelf edge is marked by a prominent free-air gravity anomaly that also corresponds to the continent-ocean boundary (Sheridan, 1988) (Klitgord, 1995).

Bouguer gravity values increase eastward from about -80 milligals (mgal) in the Valley and Ridge Province of western Virginia to about +10 mgal in the Coastal Plain Province, corresponding to an approximately 90 mgal regional anomaly across the Appalachian Orogen (Figure 2.5.1-16 and Figure 2.5.1-20). This regional gradient is called the "Piedmont gravity gradient" (Harris, 1982), and is interpreted to reflect the eastward thinning of the North American continental crust and the associated positive relief on the Moho discontinuity with proximity to the Atlantic margin.

The Piedmont gravity gradient is punctuated by several smaller positive anomalies with wavelengths ranging from about 15 to 50 mi (25 to 80 km) , and amplitudes of about 10 to 20 mgal. Most of these anomalies are associated with accreted Taconic terranes such as the Carolina/Chopawamsic terrane (Figure 2.5.1-16). Collectively, they form a gravity high superimposed on the regional Piedmont gradient that can be traced northeast-southwest on the 1:5,000,000-scale DNAG map relatively continuously along the trend of the Appalachian orogenic belt through North Carolina, Virginia, and Maryland (Figure 2.5.1-20). The continuity of this positive anomaly diminishes to the southwest in South Carolina, and the trend of the anomaly is deflected eastward in Maryland, Pennsylvania, and Delaware.

The short-wavelength anomalies and possible associations with upper crustal structure are illustrated by combining gravity profiles with seismic reflection data and geologic data (Harris, 1982) (Glover, 1995b). In some cases, short-wavelength positive anomalies are associated with antiformal culminations in Appalachian thrust sheets. For example, there is a positive anomaly associated with an anticline at the western edge of the Blue Ridge nappe along the Interstate I-64 transect (Harris, 1982) (Figure 2.5.1-16). The anomaly is presumably due to the presence of denser rocks transported from depth and thickened by antiformal folding in the hanging wall of the thrust.

The Salisbury geophysical anomaly (SGA) is a paired Bouguer gravity anomaly and magnetic high that is located along the west side of the Salisbury Embayment (Klitgord, 1995) (Figure 2.5.1-16, Figure 2.5.1-17, Figure 2.5.1-19, and Figure 2.5.1-20). The SGA is located about 10 mi (16 km) west of the CCNPP site (Figure 2.5.1-21). The anomaly is expressed most clearly as a magnetic lineation that separates a zone of short-wavelength, high-amplitude magnetic lineations to the west from a zone of low-amplitude, long-wavelength anomalies to the east. The gravity data show the SGA to form the western margin of a broad gravity low that extends seaward to the basement hinge zone. The anomaly takes the form of a north-northeast-trending gravity high having about 30 mgal relief (Johnson, 1973). The anomaly has also been named the Sussex-Curioman Bay trend (Levan, 1963) or the Sussex-Leonardtown anomaly (Daniels, 1985), and is believed to reflect an east-dipping mafic rock body associated with a suture zone buried beneath coastal plain sediments (Figure 2.5.1-16). The SGA is interpreted (Klitgord, 1995) to mark the likely location of the Taconic suture that separates the Goochland terrane on the west from a zone of island arc and oceanic metavolcanics formed in the Iapetus Ocean on the east. The SGA is shown (Horton, 1991) to be associated with the buried Sussex terrane is a probable mafic mélange that was interpreted by Lefort and Max (Lefort, 1989) to mark the Alleghenian "Chesapeake Bay suture" (Figure 2.5.1-15).

The offshore portions of the site region contain a prominent, long-wavelength free-air gravity anomaly associated with the transition from continental to oceanic crust (Sheridan, 1988) (Klitgord, 1995) (Figure 2.5.1-18). This anomaly is large (75 to 150 mgal peak to trough) and is 45 to 80 mi (72 to 129 km) wide. Variations in the amplitude and shape of the anomaly along the Atlantic margin are due to seafloor relief, horizontal density variations in the crust, and relief on the crust-mantle boundary (Sheridan, 1988) (Klitgord, 1995).

In summary, gravity data published since the mid-1980s confirm and provide additional documentation of previous observations of a gradual "piedmont gravity gradient" across the Blue Ridge and Piedmont Provinces of Virginia and a prominent gravity anomaly at the seaward margin of the continental shelf. Shorter-wavelength anomalies such as the SGA also are recognized in the data. All anomalies were known at the time of the 1986 EPRI study. The "piedmont gravity gradient" is interpreted to reflect eastward thinning of the North American crust and lithosphere. The free-air anomaly at the outer shelf edge is interpreted as reflecting the transition between continental and oceanic crust. Second-order features in the regional field, such as the Salisbury geophysical anomaly and the short discontinuous northeast-trending anomaly east of the site, primarily reflect density variations in the upper crust associated with the boundaries and geometries of Appalachian thrust sheets and accreted terranes.

2.5.1.1.4.3.2 Magnetic Data and Features

Magnetic data compiled for the 2002 Magnetic Anomaly Map of North America reveal numerous northeast-southwest-trending magnetic anomalies, generally parallel to the structural features of the Appalachian orogenic belt (Bankey, 2002) (Figure 2.5.1-19). Unlike the gravity field, the magnetic field is not characterized by a regional, long-wavelength gradient that spans the east-west extent of the site region. A magnetic profile along Interstate-64 published to accompany a seismic reflection profile (Harris, 1982) shows anomalies with wavelengths of about 6 to 30 mi (10 to 48 km). It has been concluded (Harris, 1982) that anomalies in the magnetic field primarily are associated with upper-crustal variations in magnetic susceptibility and, unlike the gravity data, do not provide information on crustal-scale features in the lithosphere.

Prominent north- to northeast-trending magnetic anomalies in the CCNPP site region include the interior New York-Alabama, Ocoee, and Clingman lineaments, the Coastal Plain Salisbury geophysical anomaly and near shore Brunswick magnetic anomaly, and the offshore East Coast magnetic anomaly (King, 1978) (Klitgord, 1988) (Klitgord, 1995) (Bankey, 2002) (Figure 2.5.1-19). The offshore Blake Spur magnetic anomaly is outside the site region.

King and Zietz (1978) identified a 1,000 mi (1,600 km) long lineament in aeromagnetic maps of the eastern U.S. that they referred to as the "New York-Alabama lineament" (NYAL) (Figure 2.5.1-19). The NYAL primarily is defined by a series of northeast-southwest-trending linear magnetic anomalies in the Valley and Ridge province of the Appalachian fold belt that systematically intersect and truncate other magnetic anomalies. The NYAL is located about 160 mi (257 km) northwest of the CCNPP site.

The Clingman lineament is an approximately 750 mi (1,200 km) long, northeast-trending aeromagnetic lineament that passes through parts of the Blue Ridge and eastern Valley and Ridge provinces from Alabama to Pennsylvania (Nelson, 1981). The Ocoee lineament splays southwest from the Clingman lineament at about latitude 36°N (Johnston, 1985a). The Clingman-Ocoee lineaments are sub-parallel to and located about 30 to 60 mi (48 to 97 km) east of the NYAL. These lineaments are located about 60 mi northwest of the CCNPP site.

King and Zietz (King, 1978) interpreted the NYAL to be a major strike-slip fault in the Precambrian basement beneath the thin-skinned fold-and-thrust structures of the Valley and Ridge province, and suggested that it may separate rocks on the northwest that acted as a mechanical buttress from the intensely deformed Appalachian fold belt to the southeast. Shumaker (Shumaker, 2000) interpreted the NYAL to be a right-lateral strike-slip fault that formed during an initial phase of Late Proterozoic continental rifting that eventually led to the opening of the Iapetus Ocean.

The Clingman lineament also is interpreted to arise from a source or sources in the Precambrian basement beneath the accreted and transported Appalachian terranes (Nelson, 1981). Johnston (Johnston, 1985a) observed that the "preponderance of southern Appalachian seismicity" occurs within the "Ocoee block", a Precambrian basement block bounded by the NYAL and Clingman-Ocoee lineaments (the Ocoee block was previously defined by (Johnston, 1985b)). Based on the orientations of nodal planes from focal mechanisms of small earthquakes, it was noted (Johnston, 1985) that most events within the Ocoee block occurred by strike-slip displacement on north-south and east-west striking faults, Johnston (Johnston, 1985a) did not favor the interpretation of seismicity occurring on a single, through-going northeast-southwest-trending structure parallel to the Ocoee block boundaries.

The Ocoee block lies within a zone defined by Wheeler (Wheeler, 1995) (Wheeler, 1996) as extended continental crust of the Late Proterozoic to Cambrian Iapetan terrane. Synthesizing geologic and geophysical data, Wheeler (Wheeler, 1995) mapped the northwest extent of the Iapetan normal faults in the subsurface below the Appalachian detachment, and proposed that earthquakes within the region defined by Johnston and Reinbold (Johnston, 1985b) as the Ocoee block may be the result of reactivation of Iapetan normal faults as reverse or strike-slip faults in the modern tectonic setting.

The East Coast magnetic anomaly (ECMA) is a prominent, linear, segmented magnetic high that extends the length of the Atlantic continental margin from the Carolinas to New England (Figure 2.5.1-19). The anomaly is about 65 mi (105 mi) wide and has an amplitude of about 500 nT. This anomaly approximately coincides with the seaward edge of the continental shelf, and has been considered to mark the transition from continental to oceanic crust. Klitgord et al. (1995) note that the anomaly is situated above the seaward edge of the thick Jurassic volcanic wedge and lower crustal zone of magmatic under plating along the boundary between rift-stage and marginal oceanic crust (Figure 2.5.1-17 and Figure 2.5.1-18). The ECMA is not directly associated with a fault or tectonic feature, and thus is not a potential seismic source.

The Brunswick magnetic anomaly (BMA) is located along the basement hinge zone offshore of the Carolinas, at the southern portion of the site region about 200 mi (322 km) from the CCNPP site (Figure 2.5.1-19). The lineament is narrower and has less amplitude than the ECMA (Klitgord, 1995). The BMA may continue northward along the hinge zone of the Baltimore Canyon Trough, but the magnetic field there is much lower in amplitude and the lineament is diffuse. The BMA is not directly related to a fault or other tectonic structure, and thus is not a potential seismic source.

The Blake Spur magnetic anomaly (BSMA) is located east of the site region above oceanic crust, about 290 mi (465 km) from the CCNPP site (Figure 2.5.1-19). The BSMA is a low-amplitude magnetic anomaly that lies subparallel to the East Coast magnetic anomaly (Klitgord et al., 1995). The BSMA probably formed during the Middle Jurassic as the midocean ridge spreading center shifted to the east. The BSMA coincides with a fault-bounded, west-side-down scarp in oceanic basement. Since its formation, the BSMA has been a passive feature in the Atlantic crust, and thus is not a potential seismic source.

The Salisbury geophysical anomaly (SGA), as mentioned above, is a paired Bouguer gravity and magnetic anomaly along the west side of the Salisbury embayment that is located about 10 mi (16 km) of the CCNPP site (Figure 2.5.1-21). The anomaly is expressed in the magnetic data as a lineament separating short-wavelength, high-amplitude magnetic lineations to the west from a zone of low-amplitude, long-wavelength anomalies to the east. The contrast in magnetic signature is related to the juxtaposition of terranes of contrasting affinity beneath coastal plain sediments, and in particular the mafic to ultramafic rocks and mélange termed the Sussex terrane by Horton et al. (1991) and believed to represent alternatively a Taconic (Glover, 1995b) or Alleghenian (Lefort, 1989) suture (Figure 2.5.1-15). Lower intensities to the west are associated with the Goochland terrane, which represents continental basement (Figure 2.5.1-16).

Discrete magnetic lows associated with the Richmond and Culpeper basins are discernible on the 2002 North America magnetic anomaly map (Bankey, 2002) (Figure 2.5.1-21). Basaltic and diabase dikes and sills are a component of the synrift fill of the exposed basins in the Piedmont and of the Taylorsville basin (Schlische, 2003) (Klitgord, 1995). The distinctive, elongate magnetic anomalies associated with these igneous bodies within the synrift basins of the Piedmont are also used beneath the Coastal Plain to delineate the Taylorsville, Queen Anne, and other synrift basins (e.g., (Benson, 1992)). The elongate magnetic anomalies are less prevalent in the magnetic field east of the Salisbury geophysical anomaly. Either the eastern rift basins do not contain as much volcanic material as the western set of rift basins or the depth to this volcanic material is considerably greater (Klitgord, 1995). Small, circular magnetic highs across the coastal plain have been interpreted as intrusive bodies (Horton, 1991) (Klitgord, 1995).

Approximately 5 to 7 mi (8 to 11 km) east of the CCNPP site is an unnamed short, discontinuous weak to moderate northeast-trending magnetic anomaly that aligns subparallel to the SGA (Figure 2.5.1 21). Similar features to the south have been interpreted as granitic intrusive anomalies, whereas Benson (1992) interprets the feature as being bound by a Mesozoic basin (Figure 2.5.1-9). A deep borehole (SM-DF-84, Figure 2.5.1-10) drilled near the southern margin of this feature encountered Jurassic (?) volcanic rocks (dated at 169 ± 8 million years old) related to Mesozoic rifting, or perhaps basic metavolcanic rocks accreted to North America as part of the Brunswick Terrane (Hansen, 1986).

A magnetic profile along an approximately west-northwest to east-southeast transect through central Pennsylvania (Glover, 1995b) (Figure 2.5.1-16) indicates that paired high and low magnetic anomalies are associated with the western margins of crustal units truncated by thrust faults. Many of these anomalies have very high amplitudes and short wavelengths. For example, there is a 400-600 nT anomaly associated with the western margin of the Blue Ridge thrust nappe. Similarly, along a continuing transect line through Virginia, Glover and Klitgord (Glover, 1995a) show a 1500-2000 nT anomaly associated with the western edge of the Potomac mélange. This transect crosses the Salisbury geophysical anomaly where it is expressed as an 600 nT anomaly (Figure 2.5.1-16). In summary, magnetic data published since the mid-1980's confirm and provide additional documentation of previous observations (i.e., pre-EPRI) across this region of eastern North America, and do not reveal any new anomalies related to geologic structures previously unknown to EPRI (EPRI, 1986).

2.5.1.1.4.4 Principal Tectonic Structures

Research since the EPRI study (EPRI, 1986) has advanced the understanding of the character and timing of the crustal architecture and tectonic history of the Atlantic continental margin. The research has explained the significance of many geophysical anomalies and has clarified the timing and kinematics of tectonic processes from the Late Precambrian through the Cenozoic. Since the EPRI study (EPRI, 1986) was completed, new Cenozoic tectonic features have been proposed and described in the site region, and previously described features have since been characterized in more detail. New features identified since the EPRI study (EPRI, 1986) in the CCNPP site region area include gentle folds and a hypothesized minor fault on the western shore of Chesapeake Bay directly south of the CCNPP site (Kidwell, 1997). Also, new geologic data collected since 1986 has clarified the geometry and location of the Port Royal fault zone and Skinkers Neck anticline, and tectonic features representing the southern continuation of the Brandywine fault system, all of which are discussed further in the following sections. Tectonic features suggested by poorly constrained data include an unnamed fault underlying the upper Chesapeake Bay inferred by Pazzaglia (Pazzaglia, 1993), a series of warps beneath the lower Patuxent River and Chesapeake Bay near the CCNPP site hypothesized by McCartan (McCartan, 1995), and a hypothesized Stafford fault system by Marple and Talwani (Marple, 2004b) that is significantly longer and more active than previously recognized (Mixon, 2000). An additional geologic feature discovered since EPRI (1986) in the site region is the Eocene Chesapeake Bay impact crater (Figure 2.5.1-5a and Figure 2.5.1-5b) (King, 1974) (Schruben, 1994). Based on the absence of published literature documenting Quaternary tectonic deformation and spatially associated with seismicity, we conclude that this feature is not a capable tectonic source (Section 2.5.1.1.4.4.4).

In the sections below, specific tectonic features and their evidence for activity published since the EPRI (1986) study are discussed. We find that no new information has been published since 1986 on any tectonic feature within the CCNPP site region that would cause a significant change in the EPRI seismic source model.

We divide principal tectonic structures within the 200 mi (322 km) CCNPP site region into five categories based on their age of formation or most recent reactivation. These categories include Late Proterozoic, Paleozoic, Mesozoic, Tertiary, and Quaternary. Late Proterozoic, Paleozoic, and Mesozoic structures are related to major plate tectonic events and generally are mapped regionally on the basis of geological and/or geophysical data. Late Proterozoic structures include normal faults active during post-Grenville orogeny rifting and formation of the Iapetan passive margin. Paleozoic structures include thrust and reverse faults active during Taconic, Acadian, Alleghenian, and other contractional orogenic events. Mesozoic structures include normal faults active during break-up of Pangaea and formation of the Atlantic passive margin.

Tertiary and Quaternary structures within the CCNPP site region are related to the tectonic environment of the Atlantic passive margin. This passive margin environment is characterized by southwest- to northeast-oriented, horizontal principal compressive stress, and vertical crustal motions. The vertical crustal motions associated with loading of the coastal plain and offshore sedimentary basins and erosion and exhumation of the Piedmont and westward provinces of the Appalachians. Commonly, these structures are localized, and represent reactivated portions of older bedrock structures. Zones of seismicity not clearly associated with a tectonic feature are discussed separately in Section 2.5.1.1.4.5.

2.5.1.1.4.4.1 Late Proterozoic Tectonic Structures

Extensional structures related to Late Proterozoic-Early Cambrian rifting of the former supercontinent Rhodinia and formation of the Iapetan Ocean basin are located along a northeast-trending belt between Alabama and Labrador, Canada, and along east-west-trending branches cratonward (Wheeler, 1995) (Johnston, 1994) (Figure 2.5.1-22). Major structures along this northeast-trending belt include the Reelfoot rift, the causative tectonic feature of the 1811-1812 New Madrid earthquake sequence. Within the 200 mi (322 km) site region, a discrete Late Proterozoic feature includes the New York-Alabama lineament (King, 1978) (Shumaker, 2000). The Rome Trough (Ervin and McGinnis, 1975) is located directly outside the 200-mile (322 km) site region. Extended crust of the Iapetan passive margin extends eastward beneath the Appalachian thrust front approximately to the eastern edge of Mesozoic extended crust within the eastern Piedmont physiographic province (Wheeler, 1996) (Figure 2.5.1-14). This marks the western boundary of major Paleozoic sutures that juxtapose Laurentian crust against exotic crust amalgamated during the Paleozoic orogenies (Wheeler, 1996) (Figure 2.5.1-15 and Figure 2.5.1-16). At its closest approach, the area of extended Iapetan crust is located about 70 mi (113 km) northwest of the CCNPP site (Figure 2.5.1-22).

The earthquake potential of Iapetan normal faults was recognized by the EPRI team members due to the association between the Reelfoot rift and the 1811 to 1812 New Madrid earthquake sequence (EPRI, 1986). Seismic zones in eastern North America spatially associated with Iapetan normal faults include the Giles County seismic zone of western Virginia, and the Charlevoix, Quebec seismic zone, both of which are located outside the CCNPP site region (Wheeler, 1995) (Figure 2.5.1-22). Because the Iapetan structures are buried beneath Paleozoic thrust sheets and/or strata, their dimensions are poorly known except in isolated, well studied cases.

Although published literature since the EPRI study (EPRI, 1986) has made major advances in showing the association between local seismic sources and Late Proterozoic structures (Wheeler, 1992) (Wheeler, 1995) and has highlighted the extent of extended Iapetan passive margin crust (Wheeler, 1995) (Wheeler, 1996), no new information has been published since 1986 on any Late Proterozoic feature within the CCNPP site region that would cause a significant change in the EPRI study (EPRI, 1986) seismic source model.

2.5.1.1.4.4.2 Paleozoic Tectonic Structures

The central and western portions of the CCNPP site region encompass portions of the Piedmont, Blue Ridge, Valley and Ridge, and Appalachian Plateau physiographic provinces (Figure 2.5.1-1). Structures within these provinces are associated with thrust sheets, shear zones, and sutures that formed during convergent and transpressional Appalachian orogenic events of the Paleozoic Era. Tectonic structures of this affinity exist beneath the sedimentary cover of the Coastal Plain and Continental Shelf Provinces. Paleozoic structures shown on Figure 2.5.1-22 include: 1) sutures juxtaposing allochthonous (tectonically transported) rocks against proto-North American crust, 2) regionally extensive Appalachian thrust faults and oblique-slip shear zones, and 3) a multitude of smaller structures that accommodated Paleozoic deformation within individual blocks or terranes (Figure 2.5.1-15, Figure 2.5.1-16, and Figure 2.5.1-17). The majority of these structures dip eastward and sole into one or more levels of low angle, basal Appalachian decollement (Figure 2.5.1-16). Below the decollement are rocks that form the North American basement complex (Grenville or Laurentian crust).

Researchers have observed that much of the sparse seismicity in eastern North America occurs within the North American basement below the basal decollement. Therefore, seismicity within the Appalachians may be unrelated to the abundant, shallow thrust sheets mapped at the surface (Wheeler, 1995). For example, seismicity in the Giles County seismic zone, located in the Valley and Ridge Province, is occurring at depths ranging from 3 to 16 mi (5 to 25 km) (Chapman, 1994), which is generally below the Appalachian thrust sheets and basal decollement (Bollinger, 1988).

2.5.1.1.4.4.2.1 Appalachian Structures

Paleozoic faults within 200 mi (322 km) of the CCNPP site and catalog seismicity are shown on Figure 2.5.1-22 and Figure 2.5.1-23 (see section 2.5.2 for a complete discussion on seismicity). Paleozoic faults with tectonostratigraphic units are shown on Figure 2.5.1-15, Figure 2.5.1-16, and Figure 2.5.1-17. Faults mapped within the Appalachian provinces (Piedmont, Blue Ridge, Valley and Ridge) are discussed in this section along with postulated Paleozoic faults in the Coastal Plain that are buried by Cenozoic strata. No new information has been published since 1986 on any Paleozoic fault in the site region that would cause a significant change in the EPRI study (EPRI, 1986) seismic source model. Paleozoic faults are discussed below from west to east across the CCNPP site region.

Major Paleozoic tectonic structures of the Appalachian Mountains within 200 mi (322 km) of the site include the Little North Mountain-Yellow Breeches fault zone, the Hylas shear zone, the Mountain Run–Pleasant Grove fault system, the Brookneal shear zone, and the Central Piedmont shear zone (including the Spotsylvania fault) (Figure 2.5.1-22). These structures bound lithotectonic units as defined in recent literature (Horton, 1991) (Glover, 1995b) (Hibbard, 2006) (Hibbard, 2007).

The northeast-striking Little North Mountain fault zone is located within the eastern Valley and Ridge Physiographic Province of western Virginia, eastern Maryland, and southern Pennsylvania (Figure 2.5.1 15 and Figure 2.5.1-22). The fault zone forms the tip of an upper level thrust sheet that attenuated Paleozoic shelf deposits of the Laurentian continental margin during the Alleghenian Orogeny (Hibbard, 2006). The east-dipping Little North Mountain thrust sheet soles into a decollement shown as a couple miles deep (Figure 2.5.1-16). This decollement represents an upper-level detachment above a deeper decollement about 5 mi (8 km) deep (Glover, 1995b) (Figure 2.5.1-16). The Little North Mountain fault and Yellow Breeches fault to the northeast mark the approximate location of the westernmost thrusts that daylight within the Valley and Ridge Province (Figure 2.5.1-22). Farther west, thrust ramps branching from the deeper decollement rarely break the surface and overlying fault-related folds control the morphology of the Valley and Ridge Province.

The Little North Mountain-Yellow Breeches fault zone is not considered a capable tectonic source. The decollement associated with the Little North Mountain thrust is within a couple miles of the surface, suggesting the fault probably does not penetrate to seismogenic depths. No seismicity is attributed to the Little North Mountain-Yellow Breeches fault zone and published literature does not indicate that it offsets late Cenozoic deposits or exhibits geomorphic expression indicative of Quaternary deformation. Therefore, this Paleozoic fault is not considered to be a capable tectonic source.

The Hylas shear zone, active between 330 and 220 million years ago, comprises a 1.5 mi (2.4 km) wide zone of ductile shear fabric and mylonites located 71 mi (115 km) southwest of the site (Bobyarchick, 1979). The Hylas shear zone also locally borders the Mesozoic Richmond basin and appears to have been reactivated during Mesozoic extension to accommodate growth of the basin (Figure 2.5.1-9). Based on review of published literature and historical seismicity, there is no reported geomorphic expression, historical seismicity, or Quaternary deformation along the Hylas shear zone, thus this feature is not considered to be a capable tectonic source.

The Mountain Run-Pleasant Grove fault system is located within the Piedmont Physiographic Province in Virginia and Maryland and may extend to near Newark, New Jersey (Hibbard et al., 1995) (Figure 2.5.1 16 and Figure 2.5.1-22). This fault system extends across the entire site region and juxtaposes multiple-tectonized, allochthonous rocks and terranes to the east against the passive margin rocks of North American affinity to the west. The fault zone exhibits mylonitic textures, indicative of the ductile conditions in which it formed during the Paleozoic Era. Locally the allochthonous rocks are the Potomac composite terrane (Horton et al., 1991), which consists of a stack of thrust sheets containing tectonic mélange deposits that include ophiolites, volcanic arc rocks, and turbidites. This east-dipping thrust probably shallows to a decollement a couple miles below ground surface, and is shown to be truncated by the Brookneal shear zone (Figure 2.5.1-16) (Glover, 1995b). In the site region, the Mesozoic Culpeper basin overlies the Mountain Run-Pleasant Grove fault system, suggesting that portions of the Paleozoic thrust fault system may have been reactivated as normal faults in the Triassic (Figure 2.5.1-9). In northern Virginia, about 70 mi (113 km) west of the site, the Everona fault was identified within Tertiary, and possibly early Quaternary, debris flow deposits (Pavlides, 1983) (Pavlides, 1986). Subsequent studies performed during the North Anna ESP (Dominion, 2004a) on the activity of the Everona-Mountain Run fault system indicate that this fault system is not a capable tectonic source (Section 2.5.1.1.4.4.5.2).

The Brookneal shear zone is located within the Piedmont in Virginia and probably extends beneath the Coastal Plain across Virginia and Maryland to within about 50 mi (80 km) of the site (Figure 2.5.1-15 and Figure 2.5.1-22). The dextral-reverse shear zone is the northern continuation of the Brevard zone, a major terrane boundary extending from Alabama to North Carolina (Hibbard, 2002). The Brookneal shear zone juxtaposes magmatic and volcaniclastic rocks of the Chopawamsic volcanic arc to the east against the Potomac mélange to the west. This east-dipping thrust possibly truncates the Mountain Run fault at about 2.5 mi (4 km) depth, then flattens to a decollement at about 4 to 5 mi (6 to 8 km) depth that dips gently eastward beneath the surface trace of the Spotsylvania fault (Figure 2.5.1-16) (Glover, 1995b). Southwest of the site region, the Mesozoic Danville basin locally coincides with the Brookneal shear zone, suggesting that portions of the Paleozoic fault may have been reactivated as normal faults in the Triassic Period. The Brookneal shear zone is not considered a capable tectonic source. No seismicity is attributed to it and published literature does not indicate that it offsets late Cenozoic deposits or exhibits geomorphic expression indicative of Quaternary deformation. Therefore, this Paleozoic fault is not considered to be capable tectonic source.

[The northeast-striking Central Piedmont shear zone - Spotsylvania fault has been mapped in the Virginia piedmont as far north as Fredericksburg and beneath the Coastal Plain in eastern Virginia and Maryland (Hibbard, 2006) (Horton, 1991) (Glover, 1995b) (Figure 2.5.1-15, Figure 2.5.1-16 and Figure 2.5.1-22). At its closest approach, the fault is about 40 mi (64 km) northwest of the site (Figure 2.5.1-15) The Spotsylvania fault is a dextral-reverse fault that is part of the Central Piedmont shear zone (Hibbard, 2006). The fault juxtaposes terranes of different affinity, placing continental rocks of the Goochland terrane to the east against volcanic arc rocks of the Chopawamsic terrane to the west. The east-dipping fault likely penetrates the crust at gentle to intermediate angles, and truncates the basal Appalachian decollement and higher decollement of the Brookneal shear zone (Figure 2.5.1-16) (Glover, 1995b).

The Spotsylvania fault is not considered a capable tectonic source. Specific studies of this feature by Dames and Moore (DM, 1977b) demonstrate that the Spotsylvania thrust fault exhibits negligible vertical deformation of a pre- to early-Cretaceous erosion surface and is not related to Tertiary faulting along the younger Stafford fault zone (Section 2.5.1.1.4.4.4). The fault was determined by the NRC (AEC) to be not capable within the definition of 10 CFR 100,

Appendix A (CFR, 2006). No subsequent evidence has been published since the Dames and Moore (DM, 1977b) study to indicate potential Quaternary activity on the fault.

2.5.1.1.4.4.2.2 Coastal Plain Structures

Major Paleozoic tectonic structures beneath the Coastal Plain in the 25 mi (40 km) CCNPP site vicinity include faults bounding the Sussex terrane west of the site and unnamed faults mapped seaward of the CCNPP site by Glover and Klitgord (Glover, 1995a) (Figure 2.5.1-15, Figure 2.5.1-16 and Figure 2.5.1-22). These fault zones, cited here as the western and eastern zones, are interpreted to dip steeply east, penetrate the crust, and juxtapose lithostratigraphic terranes.

The western fault zone coincides with the margins of the Sussex Terrane of Horton (Horton, 1991) (Figure 2.5.1-15 and Figure 2.5.1-16). The narrow Sussex Terrane and potential bounding faults are delimited in part by the Salisbury geophysical anomaly, a positive gravity and magnetic high described in Section 2.5.1.1.4.3. The eastern fault zone is shown to extend from coastal North Carolina to southern Delaware, trending north along the eastern part of southern Chesapeake Bay before branching into two splays that trend northeast across the Delmarva Peninsula (Figure 2.5.1-15 and Figure 2.5.1-22). The regional crustal cross section shows the fault zone as dipping east at moderate to steep angles (Figure 2.5.1-16).

No seismicity is attributed to the buried Paleozoic faults and published literature does not indicate that these faults offset late Cenozoic deposits or exhibit geomorphic expression indicative of Quaternary deformation. Therefore, the Paleozoic structures (faults bounding the Sussex terrane west of the site and unnamed faults mapped seaward of the CCNPP site by Glover and Klitgord (Glover, 1995a) in the site vicinity are not considered to be capable tectonic sources.

Other Paleozoic faults mapped by Hibbard (Hibbard, 2006) within the 200 mi (322 km) site region are smaller features that typically are associated with larger Paleozoic structures and accommodate internal deformation within the intervening structural blocks (Figure 2.5.1-22). No seismicity is attributed to these faults and published literature does not indicate that any of these faults offset late Cenozoic deposits or exhibit geomorphic expression indicative of Quaternary deformation. Therefore, these Paleozoic structures in the site region are not considered to be capable tectonic sources

2.5.1.1.4.4.3 Mesozoic Tectonic Structures

Mesozoic basins have long been considered potential sources for earthquakes along the eastern seaboard and were considered by most of the EPRI teams in their definition of seismic sources (EPRI, 1986). A series of elongate rift basins of early Mesozoic age are exposed in a belt extending from Nova Scotia to South Carolina and define the area of extended Mesozoic crust (Figure 2.5.1-9). These Mesozoic rift basins, also commonly referred to as Triassic basins, exhibit a high degree of parallelism with the surrounding structural grain of the Appalachian orogenic belt. The parallelism generally reflects reactivation of pre-existing Paleozoic structures (Ratcliffe, 1986). The rift basins formed during extension and thinning of the crust as Africa and North America rifted apart to form the modern Atlantic Ocean (Section 2.5.1.1.4.1.2).

Generally, the rift basins are asymmetric half-grabens with the primary rift-bounding faults on the western margin of the basin (Figure 2.5.1-9, Figure 2.5.1-17 and Figure 2.5.1-18) (Withjack, 1998). Within the 200 mi (322 km) CCNPP site region, rift basins with rift-bounding faults on the western margin include the exposed Danville, Richmond, Culpeper, Gettysburg, and Newark basins, and the buried Taylorsville, Norfolk, and other smaller basins (Figure 2.5.1-9). In most of the above-mentioned basins, the basin-bounding normal fault is located in close proximity to a Paleozoic thrust or reverse fault (e.g., the Culpeper basin and the Paleozoic Mountain Run fault zone; the Richmond basin and the Paleozoic Hylas shear zone) (Figure 2.5.1-9 and Figure 2.5.1-22). The rift-bounding normal faults are interpreted by some authors to be listric at depth and merge into Paleozoic low angle basal décollement (Manspeizer, 1989). Other authors interpret rift-bounding faults to penetrate deep into the crust following deep crustal fault zones (Figure 2.5.1-18).

The geometry and continuity of buried rift basins beneath the Coastal Plain and Continental Shelf is not clear, but the recognition and interpretation of these basins have expanded since the EPRI (1986) study. In addition to the identification of new basins since 1986, several alternative geometries have been proposed for the site region (Figure 2.5.1-9 and Figure 2.5.1-15) (Horton, 1991) (Benson, 1992) (Klitgord, 1995) (Withjack, 1998) (LeTourneau, 2003). Interpretations are constrained loosely based on sparse borehole, seismic, and aeromagnetic anomaly data (Benson, 1992). Some authors show the Queen Anne basin located beneath the CCNPP site (e.g., in Figure 2.5.1-9 (Benson, 1992) and in Figure 2.5.1-15 (Horton, 1991)). More recent compilations of rift basins do not show the CCNPP site overlying a Mesozoic basin (e.g., in Figure 2.5.1-9 (Withjack, 1998) and in Figure 2.5.1-15 (Glover, 1995b)).

Reactivation of faults bordering or within Triassic basins in the Cenozoic as reverse faults is recognized in several basins within the site region and is discussed in Section 2.5.1.1.4.1.2. (e.g., (Schlische, 2003)). For example, the buried Taylorsville basin coincides with numerous postrift contractional structures of Cretaceous and Tertiary age including the Brandywine, Port Royal, Skinkers Neck, and Hillville faults (Section 2.5.1.1.4.4.4).

Aside from the global finding of Johnston et al. (1994) that areas of Mesozoic extended crust are correlated with large magnitude earthquakes within stable continental regions (i.e., New Madrid seismic zone), there are no specific Mesozoic basin-bounding faults that have demonstrable associated seismic activity or evidence for recent fault activity (Figure 2.5.1-9 and Figure 2.5.1-23). The major postulated basins closest to the site (Taylorsville and Queen Anne) were considered during the 1980s to exist and several were incorporated into seismic sources by the different EPRI teams. Seismicity potentially associated with reactivation of faults bordering or beneath the Mesozoic basins is captured in the existing EPRI seismic source model. No new data have been developed to demonstrate that any of the Mesozoic basins are currently active, and Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) and Wheeler (Wheeler, 2006) do not recognize any basin-margin faults that have been reactivated during the Quaternary in the site region. No Mesozoic basin in the site region is associated with a known capable tectonic source, and no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4 Tertiary Tectonic Structures

Several faults were active during the Tertiary Period within the 200 mi (322 km) CCNPP site region (Figure 2.5.1-24). These faults have been recognized in the western part of the Coastal Plain Province where Tertiary strata crop out in river valleys and where the faults have bee investigated using seismic and borehole data. These faults include the relatively well characterized Stafford fault system in Virginia, the Brandywine fault system in Maryland, and the National Zoo/Rock Creek faults in Washington, D.C. Additional faults and fault-related folds defined by seismic and borehole data include the Port Royal fault zone and Skinkers Neck anticline in Virginia, and the Hillville fault in Maryland. Tertiary structures that have been proposed but are poorly constrained by data include east-facing monoclines along the western shore of Chesapeake Bay (McCartan, 1995) and a northeast-striking fault in the upper Chesapeake Bay (Pazzaglia, 1993). In addition, Kidwell (Kidwell, 1997) uses detailed stratigraphic analysis of the Calvert Cliffs area to postulate the existence of several broad folds developed in Miocene strata as well as a poorly constrained postulated fault. All of these structures are located within about 50 mi (80 km) of the site, and the proposed east-facing monoclines of McCartan (McCartan, 1995) are within a few miles of the CCNPP site. Within 25 mi (40 km) of the site, the only fault with documented Tertiary displacement is the Hillville fault (Hansen, 1978) (Hansen, 1986) (Figure 2.5.1-24).

Several faults associated with the Eocene Chesapeake Bay impact crater have been identified near the mouth of the Chesapeake Bay about 60 mi (97 km) south of the site (Powars, 1999) (Figure 2.5.1-5a). The impact crater formed on a paleo-continental shelf when the Eocene sea in this location was approximately 1,000 ft (305 m) deep. The Chesapeake Bay impact crater was discovered in 1993, and thus post-dates the EPRI study (EPRI, 1986). The 35-million year old Chesapeake Bay impact crater is a 56 mi (90 km) wide, complex peak-ring structure defined by a series of inner and outer ring faults, some of which penetrate the Proterozoic and Paleozoic crystalline basement rocks (Powars, 1999). These faults and others within the outer and inner ring include normal-faulted slump blocks and compaction faults that extend up-section into upper Miocene and possibly younger deposits. Published literature does not indicate that any faults related to the impact crater are seismogenic or offset Quaternary deposits.

Multiple, fault-bounded secondary craters of Eocene age also have been interpreted from multichannel seismic profiles previously collected by Texaco along the Potomac River and Chesapeake Bay 20 and 40 mi (32 and 64 km) north and northwest of the main Chesapeake Bay impact crater (Poag, 2004). The secondary impact craters have diameters ranging from 0.25 to 2.9 mi (0.4 to 4.7 km). Faults associated with the secondary craters occasionally penetrate Proterozoic and Paleozoic crystalline basement rocks (Poag, 2004). Primarily middle Miocene to Quaternary sediments thicken and sag into the primary and secondary craters. Faults associated with the impact crater are not considered capable tectonic sources and are not discussed further in this section.

Faults and folds mapped within the 200 mi (322 km) CCNPP site region that displace Tertiary Coastal Plain deposits are described below. These structures include the Stafford fault system, Brandywine fault system, National Zoo/Rock Creek faults, Port Royal fault zone, Skinkers Neck anticline, and the Hillville fault. Additional hypothesized Tertiary structures for which compelling geologic or geophysical evidence is lacking are then described. These structures include hypothesized east-facing monoclines along the western shore of Chesapeake Bay near the CCNPP site described by McCartan (McCartan, 1995), a hypothesized fault in the upper Chesapeake Bay mapped by Pazzaglia (Pazzaglia, 1993), and structures interpreted in Calvert Cliffs by Kidwell (Kidwell, 1997).

2.5.1.1.4.4.4.1 Stafford Fault of Mixon,et al.

The Stafford fault (#10 on Figure 2.5.1-29) approaches within 47 mi (76 km) southwest of the site (Figure 2.5.1-24). The 42 mi (68 km) long fault system strikes approximately N35°E (Newell, 1976). The fault system consists of several northeast-striking, northwest-dipping, high-angle reverse to reverse oblique faults including, from north to south, the Dumfries, Fall Hill, Brooke, Tank Creek, Hazel Run, and an unnamed fault (Mixon et al., 2000). Two additional northeast-striking, southeast-side-down faults, the Ladysmith and the Acadia faults, are included here as part of the Stafford fault system. These individual faults are 10 to 25 mi (16 to 40 km) long and are separated by 1.2 to 3 mi (2 to 5 km) wide en echelon, left step-overs. The left-stepping pattern and horizontal slickensides found on the Dumfries fault suggest a component of dextral shear on the fault system (Mixon, 2000).

Locally, the Stafford fault system coincides with the Fall Line and a northeast-trending portion of the Potomac River (Figure 2.5.1-24). Mixon and Newell (Mixon, 1977) suggest that the Fall Line and river deflection may be tectonically controlled. Detailed drilling, trenching, and mapping in the Fredericksburg region by Dames and Moore (DM, 1973) showed that the youngest identifiable fault movement on any of the four primary faults comprising the Stafford fault system was pre-middle Miocene in age.

Subsequent studies of the Stafford fault system better document the timing of displacement. Mesozoic and Tertiary movement is documented by displacement of Ordovician bedrock over lower Cretaceous strata along the Dumfries fault and abrupt thinning of the Paleocene Aquia Formation across multiple strands of the fault system (Mixon, 2000). Minor late Tertiary activity of the fault system is documented by an 11-nch displacement by the Fall Hill fault of a Pliocene terrace deposit along the Rappahannock River (Mixon, 1978) (Mixon, 2000) and an 18 in (46 cm) displacement by the Hazel Run fault of upland gravels of Miocene or Pliocene age (Mixon, 1978). Both offsets suggest southeast-side-down displacement (Mixon, 1978).

Recent geologic and geomorphic analysis of the Stafford fault system for the application of North Anna Early Site Permit (ESP) to the NRC provides additional constraints on the age of deformation (Dominion, 2004a). Geomorphic analyses (structure contour maps and topographic profiles) of upland surfaces capped by Neogene marine deposits and topographic profiles of Pliocene and Quaternary fluvial terraces of the Rappahannock River near Fredericksburg, Virginia, indicate that these surfaces are not visibly deformed across the Stafford fault system (Dominion, 2004a). In addition, field and aerial reconnaissance of these features during the North Anna ESP, and as part of this CCNPP Unit 3 study, indicate that there are no distinct scarps or anomalous breaks in topography on the terrace surfaces associated with the mapped fault traces. The NRC (2005) agreed with the findings of the subsequent study for the North Anna ESP, and stated: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period." Collectively, this information indicates that the Stafford fault system is not a capable tectonic source as defined in Appendix A of Regulatory Guide 1.165 (NRC, 1997).

Marple (Marple, 2004a) recently proposed a significantly longer Stafford fault system that extends from Fredericksburg, Virginia to New York City as part of a northeastern extension of the postulated East Coast fault system (ECFS), (Figure 2.5.1-29) (Section 2.5.1.1.4.4.5.15). The proposed northern extension of the Stafford fault system is based on: (1) aligned apparent right-lateral deflections of the Potomac (22 mi (35 km) deflection), Susquehanna (31 mi (50 km) deflection) and Delaware Rivers (65 mi (105 km) deflection) (collectively these are named the "river bend trend"), (2) upstream incision along the Fall Line directly west of the deflections, and (3) limited geophysical and geomorphic data. Marple and Talwani (Marple, 2004b) proposed that the expanded Stafford fault system of Marple (Marple, 2004a) was a northeast extension of the ECFS of Marple and Talwani (Marple, 2000). Marple and Talwani (Marple, 2004b) further speculate that the ECFS and the Stafford fault system were once a laterally continuous and through-going fault, but subsequently were decoupled to the northwest and southeast, respectively, during events associated with the Appalachian orogeny.

Data supporting the extended Stafford fault system of Marple (Marple, 2004a) is limited. Marple and Talwani (Marple, 2004b) suggest that poorly located historical earthquakes that occurred in the early 1870's and 1970's lie close to the southwestern bend in the Delaware River and concluded an association between historical seismicity and the postulated northern extension of the Stafford fault system. Review of seismicity data available both before and after the EPRI study (EPRI, 1986) indicates a poor correlation in detail between earthquake epicenters and the expanded Stafford fault system (Figure 2.5.1-24). Geophysical, borehole and trench data collected by McLaughlin (McLaughlin, 2002), near the Delaware River across the trace of the postulated expanded Stafford fault system of Marple (Marple, 2004a), provide direct evidence for the absence of Quaternary deformation. Collectively, there is little geologic and seismologic evidence to support this extension of the fault system beyond that mapped by Mixon (Mixon, 2000).

In summary, all significant information on timing of displacement for the Stafford fault system was available prior to 1986 and incorporated into the EPRI (1986) seismic source models. New significant information published since 1986 regarding the activity of the Stafford fault system includes the geomorphic and geologic analysis performed for the North Anna ESP that concluded the fault system was not active (Dominion, 2004a). Field and aerial reconnaissance performed for the North Anna ESP and this CCNPP COL application also did not reveal any geologic or geomorphic features indicative of potential Quaternary activity along the fault system. Therefore, on the basis of a review of existing geologic literature, the Stafford fault system is not considered a capable tectonic source, and there is no new information that would require a significant revision to the EPRI (1986) seismic source model.

2.5.1.1.4.4.4.2 Brandywine Fault System

The Brandywine fault system is located approximately 30 mi (48 km) west of the site and north of the Potomac River (Figure 2.5.1-24). The 12 to 30 mi (19 to 48 km) long Brandywine fault system consists of a series of en echelon northeast-trending, southeast-dipping reverse faults with east-side-up vertical displacement. Jacobeen (Jacobeen, 1972) and Dames and Moore (DM, 1973) first described the fault system from Vibroseis™ profiles and a compilation of borehole data as part of a study for a proposed nuclear power plant at Douglas Point along the Potomac River. The fault system is composed of the Cheltenham and Danville faults, which are 4 mi and 8 mi (6 to 13 km) long, respectively. These two faults are separated by a 0.6 to 1 mi (1 to 1.6 km) wide left step-over (Jacobeen, 1972). Later work by Wilson and Fleck (Wilson, 1990) interpret one continuous 20 to 30 mi (32 to 48 km) long fault that transitions into a west-dipping flexure to the south near the Potomac River. The mapped trace of the Brandywine fault system coincides with the western margin of the Taylorsville basin (Mixon, 1977) (Hansen, 1986) (Wilson, 1990). This observation lead Mixon and Newell (Mixon, 1977) to speculate the origin of the Brandywine fault system may be related to the reversal of a pre-existing zone of crustal weakness (i.e. Taylorsville Basin border fault).

The Brandywine fault system was active in the Early Mesozoic and reactivated during late Eocene and possibly middle Miocene time (Jacobeen, 1972) (Wilson, 1990). Basement rocks have a maximum vertical displacement of approximately 250 ft (76 m) across the fault (Jacobeen, 1972). Also, the Cretaceous Potomac Formation is 150 ft (46 m) thinner on the east (up-thrown) side of the fault indicating syndepositional activity of the fault. The faulting is interpreted to extend upward into the Eocene Nanjemoy Formation (70 ft (21 m) offset) (Wilson, 1990), and die out as a subtle flexure developed within the Miocene Calvert Formation (8 ft (2.4 km) flexure) (Jacobeen, 1972).

Wilson and Fleck (Wilson, 1990) speculate that the fault system continues northeast toward the previously mapped Upper Marlboro faults, near Marlboro, Maryland (Figure 2.5.1-24). Dryden (Dryden, 1932) reported several feet of reverse faulting in Pliocene Upland deposits in a railroad cut near Upper Marlboro, Maryland (Prowell, 1983). However, these faults are not observed beyond this exposure. Wheeler (Wheeler, 2006) suggests that the Upper Marlboro faults have a surficial origin (i.e., landsliding) based on the presence of very low dips and geometric relations inconsistent with tectonic faulting. Field reconnaissance conducted as part of this

CCNPP Unit 3 study used outcrop location descriptions from Prowell (Prowell, 1983) but failed to identify any relevant exposures associated with the faults of Dryden (Dryden, 1932). Wheeler's (Wheeler, 2006) assessment of the Upper Marlboro fault appears to be consistent with the outcrop described by Dryden (Dryden, 1932) as not being associated with the Brandywine fault system.

Geologic information indicates that the Brandywine fault system was last active during the Miocene. All geologic information on the timing of displacement on the Brandywine fault system was available and incorporated into the EPRI seismic source models in 1986. The post-EPRI study by Wilson and Fleck (Wilson, 1990) extended the fault north and south as an anticline, but offers no new information about the timing of the deformation. There is no pre-EPRI or post-EPRI seismicity associated with this fault system. This fault system is identified only in the subsurface and geologic mapping along the surface projection of the fault zone does not show a fault (DM, 1973) (McCartan, 1989a) (McCartan, 1989b). Field and aerial reconnaissance performed as part of this CCNPP Unit 3 study, coupled with interpretation of Light Detection and Ranging (LiDAR) data (see Section 2.5.3.1 for additional information regarding the general methodology), revealed no anomalous geomorphic features indicative of potential Quaternary activity. The Brandywine fault system, therefore, is not a capable tectonic source and there is no new information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4.3 Port Royal Fault Zone and Skinkers Neck Anticline

The Port Royal fault zone and Skinkers Neck anticline are located about 32 mi (51 km) west of the CCNPP site, south of the Potomac River (Figure 2.5.1-24). First described by Mixon and Powars (Mixon, 1984), these structures have been identified within the subsurface by: (1) contouring the top of the Paleocene Potomac Formation, (2) developing isopach maps of the Lower Eocene Nanjemoy Formation, and (3) interpreting seismic lines collected in northern Virginia (Milici, 1991) (Mixon, 1992) (Mixon, 2000). The fault and anticline are not exposed in surface outcrop. The Port Royal fault zone is located about 4 to 6 mi (6 to 10 km) east and strikes subparallel to the Skinkers Neck anticline and the Brandywine fault system. In our discussion, we consider the Skinkers Neck anticline to consist of a combined anticline and fault zone, following previous authors.

Mixon and Newell (Mixon, 1977) first hypothesized that a buried fault zone existed beneath Coastal Plain sediments and connected the Taylorsville basin in the north to the Richmond basin in the south along a fault zone coincident with the Brandywine fault zone of Jacobeen (Jacobeen, 1972). The inferred fault of Mixon and Newell (Mixon, 1977) coincides with a gravity gradient used to target exploration studies that led to the discovery of the Port Royal fault and Skinkers Neck anticline in 1984 (Mixon, 1984) (Mixon, 1992).

The Port Royal fault zone consists of a 32 mi (51 km) long, north to northeast-striking fault zone that delineates a shallow graben structure that trends parallel to a listric normal fault bounding the Taylorsville basin (Mixon, 2000) (Milici, 1991). In map view, the fault zone makes a short left-step to the Brandywine fault system (Figure 2.5.1-24). Along the northern part of the fault zone, near the town of Port Royal, Virginia, the fault is expressed in the subsurface as a 3 mi (5 km) wide zone of warping with a west-side-up sense of displacement. Water well and seismic reflection data show an apparent west-side-up vertical component for the southwestern part of the structure also (Mixon, 1992) (Mixon, 2000) (Milici, 1991).

The Skinkers Neck anticline is located directly west of the Port Royal fault zone and southwest of the mapped terminus of the Brandywine fault system (Figure 2.5.1-24). The north- to northeast-striking structure is 30-mi (48 km) long and 3 to 5 mi (5 to 8 km) wide, and is defined as an asymmetric, low-amplitude, north-plunging anticline with a west-bounding fault (Mixon, 2000). Locally, Mixon (Mixon, 2000) map the feature as two separate, closely-spaced anticlines. Along the west side of the structure, a fault zone strikes north-to-northeast and is interpreted as a fault-bounded, down-dropped block. The Skinkers Neck anticline is not mapped north of the Potomac River by Mixon (Mixon, 1992) (Mixon, 2000). However, McCartan (McCartan, 1989a) shows two folds north of the Potomac River, west of the Brandywine fault system, and along trend with the Skinkers Neck anticline as mapped by Mixon (Mixon, 2000).

The Port Royal fault zone and Skinkers Neck anticline likely are associated with Paleozoic structures that were reactivated in the Early Mesozoic, Paleocene, and possibly middle Miocene (Mixon, 1992) (Mixon, 2000) (McCartan, 1989c). Similar to the Brandywine fault system, these structures closely coincide with the Mesozoic Taylorsville basin (Mixon, 1992) (Milici, 1991). This apparent coincidence with a Mesozoic basin suggests that the Port Royal fault zone and the Skinkers Neck anticline represent possible pre-existing zones of crustal weakness. Post-Mesozoic deformation includes as much as 30 to 33 ft (9 to 10 km) of Paleocene offset, and less than 25 ft (7.6 m) of displacement across the basal Eocene Nanjemoy Formation. Deformation on the order of 5 to 10 ft (1.5 to 3 m) is interpreted to extend upward into the Middle Miocene Calvert and Choptank Formations (Mixon, 1992). The overlying Late Miocene Eastover Formation is undeformed across both the Port Royal fault zone and Skinkers Neck anticline, constraining the timing of most recent activity (Mixon, 1992) (Mixon, 2000).

Although the Port Royal fault zone and Skinkers Neck anticline where characterized after the EPRI study (EPRI, 1986), geological information available to the EPRI teams regarding the pre-Quaternary activity of the structures was available (Mixon, 1984). Both of these structures are mapped in the subsurface as offsetting Tertiary or older geologic units (Mixon, 2000). Field and aerial (inspection by plane) reconnaissance, coupled with interpretation of aerial photography (review and inspection of features preserved in aerial photos) and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this CCNPP Unit 3 study shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the fault zone (i.e., along the northern banks of the Potomac River and directly northeast of the fault zone). Also, there is no pre-EPRI or post-EPRI (EPRI, 1986) seismicity spatially associated with the Port Royal fault zone or the Skinkers Neck anticline. In summary, the Port Royal fault zone and Skinkers Neck anticline are not considered capable tectonic sources, there is no new information developed since 1986 that would require revision to the EPRI seismic source model regarding these features.

2.5.1.1.4.4.4.4 National Zoo Faults

The National Zoo faults in Washington D.C. approach to within 47 mi (76 km) of the site (Figure 2.5.1 24). The National Zoo faults are primarily low-angle to high-angle, northwest-striking, southwest-dipping thrust faults that occur within a 1.0 to 1.5 mi (1.6 to 2.4 km) long, north to northeast-trending fault zone (Prowell, 1983) (McCartan, 1990) (Fleming, 1994) (Froelich, 1975). The mapped surface traces of these faults range from 500 to 2000 ft (152 to 610 km) with up to 20 ft (6 m) of post-Cretaceous reverse displacement visible in outcrops at the National Zoo (Fleming, 1994). The faults were first identified by Darton (Darton, 1950) in exposures along Rock Creek in historic excavations between the National Zoo and Massachusetts Avenue in Washington D.C.

The National Zoo faults were active during the Early Mesozoic with probable reactivation during the Pliocene (Darton, 1950) (McCartan, 1990) (Fleming, 1994). This fault zone is coincident with the mapped trace of the Early Paleozoic Rock Creek shear zone, which led several researches to infer that the National Zoo faults are related to reversal of a pre-existing zone of crustal weakness (McCartan, 1990) (Fleming, 1994). Combined with the Rock Creek fault zone, the National Zoo faults could be up to 16 mi (26 km) long. Differential offset across basement and Potomac Group contacts also suggests Paleozoic fault reactivation (Fleming, 1994). The Cretaceous Potomac formation offsets are primarily less than 50 ft (15 m) and isopach maps show a thickening of Coastal Plain sediments east of these faults (Fleming, 1994) (Darton, 1950). The youngest two faults juxtapose basement rocks over Pliocene Upland gravels (Fleming, 1994) (McCartan, 1990). One exposure of these two faults is still preserved along Adams Mill road as a special monument (Prowell, 1983). Based on our field reconnaissance with USGS researchers, future additional investigations are planned by the USGS to further investigate the age of the gravels and lateral continuity of the National Zoo faults.

All information on timing of displacement of the National Zoo faults was available and incorporated into the EPRI seismic source models in 1986. Although later detailed mapping of these thrust faults with the Rock Creek shear zone was published after completion of the EPRI study (EPRI, 1986), Darton (Darton, 1950) and Prowell (Prowell, 1983) identified these faults as active during Cenozoic time. In addition, there is no pre-EPRI or post-EPRI seismicity spatially associated with this fault zone. Therefore, the conclusion is that the National Zoo faults are not a capable tectonic source. There also is no new published geologic information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.4.5 Hillville Fault Zone

The Hillville fault zone of Hansen (1978) approaches to within 5 mi (8 km) of the site in the subsurface (Figure 2.5.1-24, Figure 2.5.1-25, and Figure 2.5.1-26a). The 26 mi (42 km) long, northeast-striking fault zone is composed of steep southeast-dipping reverse faults that align with the east side of the north-to northeast-trending Sussex-Currioman Bay aeromagnetic anomaly (i.e. SGA, Figure 2.5.1-21). Based on seismic reflection data, collected about 9 mi (15 km) west-southwest of the site, the fault zone consists of a narrow zone of discontinuities that vertically separate basement by as much as 250 ft (76 m) (Hansen, 1978).

The Hillville fault zone delineates a possible Paleozoic suture zone reactivated in the Mesozoic and Early Tertiary. The fault zone is interpreted as a lithotectonic terrane boundary that separates basement rocks associated with Triassic rift basins on the west from low-grade metamorphic basement on the east (i.e., Sussex Terrane/Taconic suture of Glover and Klitgord, (Glover, 1995a) (Figure 2.5.1-16) (Hansen, 1986). The apparent juxtaposition of the Hillville fault zone with the Sussex-Currioman Bay aeromagnetic anomaly suggests that the south flank of the Salisbury Embayment may be a zone of crustal instability that was reactivated during the Mesozoic and Tertiary. Cretaceous activity is inferred by Hansen (Hansen, 1978) who extends the fault up into the Cretaceous Potomac Group. The resolution of the geophysical data does not allow an interpretation for the upward projection of the fault into younger overlying Coastal Plain deposits (Hansen, 1978). Hansen (Hansen, 1978), however, used stratigraphic correlations of Coastal Plain deposits from borehole data to speculate that the Hillville fault may have been active during the Early Paleocene.

There is no geologic data to suggest that the Hillville fault is a capable tectonic source. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this COL study shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the Hillville fault zone. A review of geologic cross sections (McCartan, 1989a) (McCartan, 1989b) (Glaser, 2003b) (Glaser, 2003c) show south-dipping Lower to Middle Miocene Calvert Formation and no faulting along projection with the Hillville fault zone. Furthermore Quaternary terraces mapped by McCartan (McCartan, 1989b) and Glaser (Glaser, 2003b) (Glaser, 2003c) bordering the Patuxent and

Potomac Rivers were evaluated for features suggestive of tectonic deformation by interpreting LiDAR data and aerial reconnaissance (Figure 2.5.1-25 and Figure 2.5.1-26a). No northeast-trending linear features coincident with the zone of faulting were observed where the surface projection of the fault intersects these Quaternary surfaces. Aerial reconnaissance of this fault zone also demonstrated the absence of linear features coincident or aligned with the fault zone. Lastly, interpretation of the detailed stratigraphic profiles collected along Calvert Cliffs and the western side of Chesapeake Bay provide geologic evidence for no expression of the fault where the projected fault would intersect the Miocene-aged deposits (Kidwell, 1997; see Section 2.5.3 for further explanation). Therefore, we conclude that the Hillville fault zone is not a capable tectonic source, and there is no new information developed since 1986 that would require a significant revision to the EPRI model.

2.5.1.1.4.4.4.6 Unnamed Fault beneath Northern Chesapeake Bay, Cecil County, Maryland

Pazzaglia (1993) proposed a fault in northern Chesapeake Bay that comes to within 70 mi (113 km) north of the site (Figure 2.5.1-24). On the basis of geologic data and assuming that the bay is structurally controlled, Pazzaglia (1993) infers a 14 mi (23 km) long, northeast-striking fault with a southwest-side up sense of displacement. Near the mouth of the Susquehanna River, in Maryland, the unnamed fault is interpreted to vertically separate Pleistocene Turkey Point gravels of the Quaternary Pennsauken Formation on the east at elevations higher than a similar gravel deposit mapped on the west side of the Chesapeake Bay. The amount of apparent vertical separation is unconstrained because the base of the gravel unit is not exposed west of the bay; however, estimates of the exposed section provide a minimum of 26 ft (8 m) of vertical separation of the Pleistocene Turkey Point gravels (Pazzaglia, 1993).

This fault is unconfirmed based on the lack of direct supporting evidence. First, the fault has not been observed as a local discontinuity on land. Second, the correlation of gravels is permissible based on the data, but has not been confirmed by detailed stratigraphic or chronologic studies. Geologic mapping of the area (Higgins, 1986) shows Miocene Upland gravels along the northeast mouth of the Susquehanna River where Pazzaglia (Pazzaglia, 1993) maps the Quaternary Pennsauken Formation.

There is no geologic data to suggest that this unnamed fault zone is a capable tectonic source. There is no pre-EPRI or post-EPRI seismicity spatially associated with this fault zone. Field and aerial reconnaissance conducted to support CCNPP Unit 3 shows that there are no geomorphic features indicative of potential Quaternary activity along the surface-projection of the unnamed fault; therefore, this fault is not a capable tectonic source.

2.5.1.1.4.4.4.7 Unnamed Monocline beneath Chesapeake Bay

McCartan (McCartan, 1995) show east-facing monoclinal structures bounding the western margin of Chesapeake Bay 1.8 and 10 mi (2.9 and 16 km) east and southeast, respectively, of the site (Figure 2.5.1-24). Also, McCartan (McCartan, 1995) interprets an east-facing monocline about 10 mi (16 km) west of the site. The three monoclinal structures are depicted on two cross sections as warping Lower Paleocene to Upper Miocene strata with approximately 60 to 300 ft (18 to 91 m) of relief. The monoclines exhibit a west-side up sense of structural relief that projects upward into the Miocene Choptank Formation (McCartan, 1995). The overlying Late Miocene St. Marys Formation is not shown as warped. Boreholes shown with the cross sections accompanying the McCartan (MaCartan, 1995) map provide the only direct control on cross section construction. The boreholes are widely spaced and do not appear to provide a constraint on the existence and location of the warps. No borehole data is available directly west of the cliffs and within the bay to substantiate the presence of the warp. No surface trace or surface projection of the warps is indicated on the accompanying geologic map. Based on text accompanying the map and cross sections, we infer that the cross sections imply two approximately north- to northeast-striking, west-side up structures, of presumed tectonic origin.

McCartan (McCartan, 1995) interpret the existence of the monocline based on three observations in the local landscape. Firstly, the north to northeast-trending western shore of Chesapeake Bay within Calvert County is somewhat linear and is suggestive of structural control (McCartan, 1995). Secondly, land elevation differences west and east of Chesapeake Bay are on the order of 90 ft (27 m), with the west side being significantly higher in elevation, more fluvially dissected, and composed of older material compared to the east side of Chesapeake Bay. On the west side of the bay, the landscape has surface elevations of 100 to 130 ft (30 to 40 m) msl and drainages are incised into the Pliocene Upland Deposits and Miocene-aged deposits of the St. Mary's, Choptank, and Calvert Formations. Along the eastern shoreline of the Delmarva Peninsula, surface elevations are less than 20 to 30 ft (6 to 9 m) msl and the surface exhibits minor incision and a more flat-lying topographic surface. These eastern shore deposits are mapped as Quaternary estuarine and deltaic deposits. Thirdly, variations in unit thickness within Tertiary deposits between Calvert Cliffs and Delmarva Peninsula are used to infer the presence of a warp. Based on these physiographic, geomorphic and geologic observations, McCartan (McCartan, 1995) infer the presence of a fold along the western shore of Chesapeake Bay (Figure 2.5.1-24).

Based on the paucity of geologic data constraining the cross sections of McCartan (McCartan, 1995), the existence of the monocline is speculative. The borehole data that constrain the location of the monocline are approximately 18 to 21 mi (29 to 34 km) apart and permit, but do not require the existence of a monocline. McCartan (McCartan, 1995) do not present additional data that are inconsistent with the interpretation of flat-lying, gently east-dipping Miocene strata shown in prior published cross sections north and south of this portion of Chesapeake Bay (Cleaves et al., 1968; Milici, et al., 1995) and within Charles and St. Mary's Counties, Maryland (McCartan, 1989a) (McCartan, 1989b) (DM, 1973). No geophysical data are presented as supporting evidence for this feature. In contrast, shallow, high-resolution geophysical data collected along the length of Chesapeake Bay to evaluate the ancient courses of the submerged and buried Susquehanna River provide limited evidence strongly indicating that Tertiary strata are flat lying and undeformed along the western shore of Chesapeake Bay (Colman, 1990) (Figure 2.5.1-27).

Alternatively, the change in physiographic elevation and geomorphic surfaces between the western and eastern shores of Chesapeake Bay can be explained by erosional processes directly related to the former course of the Susquehanna River, coupled with eustatic sea level fluctuations during the Quaternary (Colman, 1990) (Owens, 1979). Colman and Halka (Colman, 1989) also provide a submarine geologic map of Chesapeake Bay at and near the site which depicts Tertiary and Pleistocene deposits interpreted from high-resolution geophysical profiles. No folding or warping or faulting is depicted on the Colman and Halka (Colman, 1989) map which encompasses the warp of McCartan (McCartan, 1995). Colman (Colman, 1990) utilize the same geophysical data to track the former courses of the Susquehanna River between northern Chesapeake Bay and the southern Delmarva Peninsula. Paleo-river profiles developed from the geophysical surveys that imaged the depth and width of the paleochannels show that the Eastville (150 ka) and Exmore (200 to 400 ka) paleochannels show no distinct elevation changes within the region of the Hillville fault and McCartan (McCartan, 1995) features.

Field reconnaissance along much of the western shoreline shows that the north- to northeast-trending linear coastline could be controlled locally, in part, by a weak, poorly-developed, sub-vertical joint set oriented subparallel to the coast (Section 2.5.1.2.4). The observation that the west side of Chesapeake Bay is elevated and dissected, and that approximately 37 ka estuarine deposits are approximately 6 feet above sea level is compelling evidence for recent (late Quaternary) uplift. Similar elevated, dissected topography and approximately 37 ka estuarine deposits are observed over broad portions of the Coastal Plain along the eastern seaboard east and west of Chesapeake Bay. These surfaces of apparent anomalous elevations have recently been attributed to the presence of a glacial fore-bulge developed outboard of the Laurentide ice sheet (Scott, 2006).

There is no geologic data to suggest that the postulated monocline along the western margin of Chesapeake Bay of McCartan (McCartan, 1995), if present, is a capable tectonic source. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during this COL study, shows that there are no geomorphic features indicative of folding directly along the western shores of Chesapeake Bay. There is no pre-EPRI or post-EPRI seismicity spatially associated with this structure. These data indicate that the McCartan (McCartan, 1995) warps, if present, most likely do not deform Pliocene to Quaternary deposits, and thus are not capable tectonic sources that would require a revision to the EPRI (1986) seismic source model.

2.5.1.1.4.4.4.8 Unnamed Folds and Postulated Fault within Calvert Cliffs, Western Chesapeake Bay, Calvert County, Maryland

The Calvert Cliffs along the west side of Chesapeake Bay provide a 25 mile (40 km) long nearly continuous exposure of Miocene, Pliocene and Quaternary deposits (Figure 2.5.1-25). Kidwell (1988 and 1997) prepared over 300 comprehensive lithostratigraphic columns along a 25 mi (40 km) long stretch of Calvert Cliffs (Figure 2.5.1-28). Because of the orientation of the western shore of Chesapeake Bay, the cliffs intersect any previously potential structures (i.e., Hillville fault) trending northeast or subparallel to the overall structural trend of the Appalachians. The cliff exposures provide a 230 ft (70 m) thick section of Cenozoic deposits that span at least 10 million years of geologic time.

On the basis of the stratigraphic profiles, Kidwell (Kidwell, 1997) develops a chronostratigraphic sequence of the exposed Coastal Plain deposits and provides information on regional dip and lateral continuity. The Miocene Choptank Formation is subdivided into two units and is unconformably overlain by the St .Marys Formation. The St. Marys Formation is subdivided into three subunits each of which is bound by a disconformity. The youngest subunit is unconformably overlain by the Pliocene Brandywine Formation (i.e., Pliocene Upland gravels). The exposed Coastal Plain deposits strike northeast and dip south-southeast between 1 and 2 degrees. The southerly dip of the strata is disrupted occasionally by several low amplitude broad undulations in the Choptank Formation, and decrease in amplitude upward into the St Marys Formation (Figure 2.5.1-28). Kidwell (Kidwell, 1997) interprets the undulations as monoclines and asymmetrical anticlines. The undulations typically represent erosional contacts that have wavelengths on the order of 2.5 to 5 mi (4 to 8 km) and amplitudes of 10 to 11 ft (about 3 m). Any inferred folding of the overlying Pliocene and Quaternary fluvial strata is very poorly constrained or obscured because of highly undulatory unconformities within these younger sand and gravel deposits. For instance, the inferred folding of the overlying Pliocene and Quaternary channelized sedimentary deposits consist of intertidal sand and mud-flats, tidal channels and tidally-influenced rivers exhibit as much as 40 ft (12 m) of erosional elevation change (Figure 2.5.1-28).

Near Moran Landing, about 1.2 mi (1.9 km) south of the site, Kidwell (Kidwell, 1997) interprets an apparent 6 to 10 ft (2 to 3 m) elevation change in Miocene strata, and a 3 to 12 ft (1 to 3.6 m) elevation change in Pliocene and Quaternary(?) fluvial material (Figure 2.5.1-25 and

Figure 2.5.1-28). Kidwell (1997) infers the presence of a fault to explain the difference in elevation of strata across Moran Landing. The postulated fault is not shown on the Kidwell (Kidwell, 1997) section, or any published geologic map; however, the inferred location is approximately 1.2 mi (1.9 km) south of the CCNPP site. The hypothesized fault is not exposed in the cliff face and is based entirely on a change in elevation and bedding dip of Miocene stratigraphic boundaries projected across the fluvial valley of Moran Landing. Kidwell (Kidwell, 1997) postulates that the fault strikes northeast and exhibits a north-side down sense of separation across all the geologic units (Miocene through Quaternary). With regard to the apparent elevation changes for the Pliocene and Quaternary unconformities, these can be readily explained by channeling and highly irregular erosional surfaces (Figure 2.5.1-28).

LiDAR data was reviewed for the possible presence of northeast-striking lineaments in the region of Moran Landing and to the southeast along the Patuxent River. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (see Section 2.5.3.1 for additional information regarding the general methodology), conducted during the CCNPP Unit 3 investigation shows that there are no geomorphic features indicative of potential Quaternary activity developed in the Pliocene-Quaternary surfaces along a southeast projection from Chesapeake Bay across the Patuxent and Potomac Rivers (Figure 2.5.1-25). The features also do not coincide with magnetic and gravity anomalies, and thus are not rooted, and more likely are surficial in origin. There is no pre-EPRI or post-EPRI (1986) seismicity spatially associated with the Kidwell (Kidwell, 1997) features, nor are there direct geologic data to indicate that the features proposed by Kidwell (Kidwell, 1997) are capable tectonic sources (Section 2.5.3.2.3)

2.5.1.1.4.4.5 Quaternary Tectonic Features

In an effort to provide a comprehensive database of Quaternary tectonic features, Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005), and Wheeler (Wheeler, 2006) compiled geological information on Quaternary faults, liquefaction features, and possible tectonic features in the CEUS. Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2005) evaluated and classified these features into one of four categories (Classes A, B, C, and D; see Table 2.5.1-1 for definitions (Crone, 2000) (Wheeler, 2005)) based on strength of evidence for Quaternary activity.

Within a 200 mi (322 km) radius of the CCNPP site, Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) and Wheeler (Wheeler, 2006) identified 17 potential Quaternary features (Figure 2.5.1-29). Work performed as part of the CCNPP Unit 3 investigation, including literature review, interviews with experts, and geologic reconnaissance, did not identify any additional potential Quaternary tectonic features within the CCNPP site region, other than those previously mentioned (McCartan, 1995) (Kidwell, 1997). Within approximately 200 mi (322 km) of the site, Crone and Wheeler (Crone, 2000) found only one feature described in the literature that exhibited potential evidence for Quaternary activity (Figure 2.5.1-29). This feature (shown as number 12) is the paleo-liquefaction features within the Central Virginia seismic zone.

The following sections provide descriptions of 15 of the 17 potential Quaternary features identified by Crone and Wheeler (Crone, 2000), Wheeler (Wheeler, 2005) (Wheeler, 2006), and of the postulated East Coast fault system of Marple and Talwani (Marple, 2004). Note that the Central Virgina and Lancaster seismic zones are discussed in Section 2.5.1.1.4.5 and Section 2.5.2. Out of the 17 features evaluated for this CCNPP Unit 3 study, nearly all are classified as Class C features, with the exception of the Central Virginia seismic zone (Class A).

The features are labeled with the reference numbers utilized in Figure 2.5.1-29:

(1) Fall lines of Weems (1998) (Class C)
(2) Ramapo fault system (Class C)
(3) Kingston fault (Class C)
(4) New York Bight fault (offshore) (Class C)
(5) Cacoosing Valley earthquake (Class C)
(6) Lancaster seismic zone (Class C)
(7) New Castle County faults (Class C)
(8) Upper Marlboro faults (Class C)
(9) Everona-Mountain Run fault zone (Class C)
(10) Stafford fault of Mixon et al. (Class C)
(11) Lebanon Church fault (Class C)
(12) Central Virginia seismic zone (Class A)
(13) Hopewell fault (Class C)
(14) Old Hickory faults (Class C)
(15) Stanleytown-Villa Heights faults (Class C)
(16) (The Stafford fault system of Marple is included in (17), i.e. the East Coast fault system)
(17) East Coast fault system (Class C)

The Everona-Mountain Run fault zone and Stafford fault of Mixon (Mixon, 2000) also are discussed in detail in previous Section 2.5.1.1.4.4.2 and Section 2.5.1.1.4.4.4.1.

2.5.1.1.4.4.5.1 Fall Lines of Weems (1998)

In 1998, Weems defined seven fall lines across the Piedmont and Blue Ridge Provinces of North Carolina and Virginia (Figure 2.5.1-29). The eastern fall line is located approximately 47 mi (76 km) west of the CCNPP site. The fall lines, not to be confused with the Fall Line separating the Piedmont and Coastal Plain provinces, are based on the alignment of short stream segments with anomalously steep gradients. Weems (1998) explores possible ages and origins (rock hardness, climatic, and tectonic) of the fall lines and "based on limited available evidence favors a neo-tectonic origin" for these geomorphic features during the Quaternary. Weems (1998) interprets longitudinal profiles for major drainages flowing primarily southeast and northwest across the Piedmont and Blue Ridge Provinces to assess the presence and origin of the "fall zones".

A critical evaluation of Weems' (1998) study, as part of the North Anna ESP, demonstrates that there are inconsistencies and ambiguities in Weems' (1998) correlations and alignment of steep reaches of streams used to define continuous fall lines (Dominion, 2004b). The North Anna ESP study concludes that that the individual fall zones of Weems (1998) may not be as laterally continuous as previously interpreted. For instance, stratigraphic, structural and geomorphic relations across and adjacent to the Weems (1998) fall zones can be readily explained by differential erosion due to variable bedrock hardness rather than Quaternary tectonism (Dominion, 2004b). Furthermore, there is no geomorphic expression of recent tectonism, such as the presence of escarpments, along the trend of the fall lines between drainages where one would expect to find better preservation of tectonic geomorphic features. Similarly, Wheeler (2005) notes that the Weems (1998) fall zones are not reproducible and are subjective, thus tectonic faulting is not yet demonstrated as an origin, and the fall lines are designated as a

Class C feature. In the Safety Evaluation Report for the North Anna ESP site study, the NRC staff agrees with the assessment that the fall lines of Weems (1998) are nontectonic features (NRC, 2005). In summary, based on review of published literature, field reconnaissance, and geologic and geomorphic analysis performed previously for the North Anna ESP application, the fall lines of Weems (1998) are erosional features related to contrasting erosional resistances of adjacent rock types, and are not tectonic in origin, and thus are not capable tectonic sources.

2.5.1.1.4.4.5.2 Everona-Mountain Run Fault Zone

The Mountain Run fault zone is located along the eastern margin of the Culpeper Basin and lies approximately 71 mi (114 km) southwest of the site (Figure 2.5.1-16 and Figure 2.5.1-29). The 75 mi (121 m) long, northeast-striking fault zone is mapped from the eastern margin of the Triassic Culpeper Basin near the Rappahannock River southwestward to near Charlottesville, in the western Piedmont of Virginia (Pavlides, 1986). The fault zone consists of a broad zone of sheared rocks, mylonites, breccias, and phyllites of variable width.

The Mountain Run fault zone is interpreted to have formed initially as a thrust fault upon which back-arc basin rocks (mélange deposits) of the Mine Run Complex were accreted onto ancestral North America at the end of the Ordovician (Pavlides, 1989). This major suture separates the Blue Ridge and Piedmont terranes (Pavlides, 1983) (Figure 2.5.1-16). Subsequent reactivation of the fault during the Paleozoic and/or Mesozoic produced strike-slip and dip-slip movements. Horizontal slickensides found in borehole samples and at several places near the base of the Mountain Run scarp suggest strike-slip movement, whereas small-scale folds in the uplands near the scarp suggest an oblique dextral sense of slip (Pavlides, 2000). The timing of the reverse and strike-slip histories of the fault zone, and associated mylonitization and brecciation, is constrained to be pre-Early Jurassic, based on the presence of undeformed Early Jurassic diabase dikes that cut rocks of the Mountain Run fault zone (Pavlides, 2000).

The northeast-striking Mountain Run fault zone is moderately to well-expressed geomorphically (Pavlides, 2000). Two northwest-facing scarps occur along the fault zone, including: (1) the 1 mi (1.6 km) long Kelly's Ford scarp located directly northeast of the Rappahannock River and; (2) the 7 mi (11 km) long Mountain Run scarp located along the southeast margin of the linear Mountain Run drainage. Conspicuous bedrock scarps in the Piedmont, an area characterized by deep weathering and subdued topography, has led some experts to suggest that the fault has experienced a Late Cenozoic phase of movement (Pavlides, 2000) (Pavlides, 1983).

Near Everona, Virginia, a small reverse fault, found in an excavation, vertically displaces "probable Late Tertiary" gravels by 5 ft (1.5 m) (Pavlides, 1983). The fault strikes northeast, dips 20 degrees northwest, and based on kinematic indicators is an oblique strike-slip fault. More recently others have estimated that the offset colluvial gravels are Pleistocene age (Manspeizer et. al, 1989). The Everona fault is located about 0.5 mi (0.8 km) west of the Mountain Run fault zone. Due to the close proximity of these two faults and their shared similar orientation and sense of slip, the Everona and Mountain Run faults are considered to be part of the same fault zone, hence the Everona-Mountain Run fault zone (Crone, 2000). Crone and Wheeler (Crone, 2000) assessed that the faulting at Everona is likely to be of Quaternary age, but because the likelihood has not been tested by detailed paleo-seismological or other investigations, this feature was assigned to Class C.

Field and aerial reconnaissance, and geomorphic analysis of deposits and features associated with the fault zone, recently performed for the North Anna ESP provide new information on the absence of Quaternary faulting along the Everona-Mountain Run fault zone (Dominion, 2004a). In response to NRC comments for the North Anna ESP, geologic cross sections and topographic profiles were prepared along the Mountain Run fault zone to further evaluate the inferred tectonic geomorphology coincident with the fault zone. The results of the additional analysis were presented in the response to an NRC Request for Additional Information (RAI) (Dominion, 2004a) and are summarized below:

- There is no consistent expression of a scarp along the Mountain Run fault in the vicinity of the Rappahannock River. The northwest-facing Kelly's Ford scarp is similar to a northwest-facing scarp along the southeastern valley margin of Mountain Run; both scarps were formed by streams that preferentially undercut the southeastern valley walls, creating asymmetric valley profiles.
- There is no northwest-facing scarp associated with the Mountain Run fault zone between the Rappahannock and Rapidan Rivers. Undeformed late Neogene colluvial deposits bury the Mountain Run fault zone in this region, demonstrating the absence of Quaternary fault activity.
- The northwest-facing "Mountain Run" scarp southwest of the Rappahannock River alternates with a southeast-facing scarp on the opposite side of Mountain Run valley; both sets of scarps have formed by the stream impinging on the edge of the valley.

All of the information on timing of displacement of the Mountain Run fault zone and associated faults was available and incorporated into the EPRI seismic source models in 1986. Significant new information developed since 1986 includes the work performed for the North Ana ESP that shows the Mountain Run fault zone has not been active during the Quaternary. In addition, the NRC staff agrees that the scarps along the Mountain Run Fault zone were not produced by Cenozoic fault activity (NRC, 2005). Similarly, Crone and Wheeler (Crone, 2000) do not show the Mountain Run fault zone as a known Quaternary structure in their compilation of active tectonic features in the CEUS, having assigned it to Class C. Based on the findings of the previous studies performed for the North Anna ESP and approval by the Nuclear Regulatory Commission (NRC, 2005), it is concluded that the Everona-Mountain Run fault zone is not a capable tectonic source. No new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.5.3 Stafford Fault of Mixon, et al.

The Stafford fault (#10 on Figure 2.5.1-29) approaches within 47 mi southwest of the site (Figure 2.5.1-24). The Stafford fault (Mixon, 2000) is discussed in more detail in Section 2.5.1.1.4.4.4.1 (Stafford Fault System). The northern extension of the Stafford fault system as proposed by Marple (#16 on Figure 2.5.1-29) is discussed in Section 2.5.1.1.4.4.5.15. The 42 mile (68 km) long fault system strikes approximately N35°E and was identified and described first by Newell (Newell, 1976). The fault system consists of a series of five northeast-striking, northwest-dipping, high-angle reverse faults including, from north to south, the Dumfries, Fall Hill, Hazel Run, and Brooke faults, and an unnamed fault. The Brooke fault also includes the Tank Creek fault located northeast of the Brooke fault (Mixon, 2000).

No new significant information has been developed since 1986 regarding the activity of the Stafford fault system with the exception of the response to an NRC RAI for the North Anna ESP (Dominion, 2004a). Field reconnaissance performed for the CCNPP Unit 3 study also did not reveal any geologic or geomorphic features indicative of potential Quaternary activity along the fault system. In addition, near the site and along the portion of the Stafford fault mapped by Mixon et al. (2000) no seismicity is attributed to the Stafford fault. Similarly, Wheeler (Wheeler, 2005) does not show the Stafford fault system as a Quaternary structure in his compilation of active tectonic features in the CEUS. The NRC (NRC, 2005) agreed with the findings of the subsequent study for the North Anna ESP, and stated: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period." Based on a review of existing information for the Stafford fault system, including the response to the NRC RAI for the North Anna ESP, the Stafford fault system is not a capable tectonic source and there is no new information developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.4.5.4 Ramapo Fault System

The Ramapo fault system is located in northern New Jersey and southern New York State, approximately 130 mi (209 km) north-northeast of the CCNPP site (Figure 2.5.1-29). This fault system consists of northeast-striking, southeast-dipping, normal faults that bound the northwest side of the Mesozoic Newark basin that to the northeast become a single 40 mi (64 km) long northeast-striking fault (Ratcliffe, 1971) (Schlische, 1992) (Drake, 1996) (Figure 2.5.1-9). Bedrock mapping by Drake (Drake, 1996) shows primarily northwest-dipping Lower Jurassic and Upper Triassic Newark Supergroup rocks in the hanging wall and tightly folded and faulted Paleozoic basement rocks in the footwall of the fault. The Ramapo fault splays into several fault strands southwest of Bernardsville and merges with the Flemington Fault zone. This fault zone also splays into several northeast- to east-trending faults in Rockland and Westchester Counties, New York.

The Ramapo fault system has been considered a potentially active tectonic feature because the fault: (1) exhibits repeated reactivation during the Paleozoic, (2) bounds the Mesozoic Newark basin (i.e. the region is composed of extended crust), and (3) aligns with earthquake epicenters (Wheeler, 2006) (Aggarwal, 1978). In cross section and map view, the seismicity data and focal mechanisms illustrate a 60° to 65° southeast-dipping fault zone that projects upward to the mapped trace of the Ramapo fault. In addition, 14 focal mechanism solutions have orientations that are consistent with the present-day stress field and suggest reverse reactivation of the Ramapo fault. Collectively, these data led Aggarwal and Sykes (Aggarwal, 1978) to conclude that the Ramapo fault is likely active.

Many of the assumptions and conclusions made by Aggarwal and Sykes (Aggarwal, 1978) were later reevaluated with alternative interpretations suggesting the fault probably has not been active during the Quaternary. Subsequent fault activity studies included several types of geophysical and geologic techniques. First, a modified velocity model and a carefully re-evaluated earthquake catalog refined the location of the earthquakes previously inferred as aligned with the Ramapo fault, and demonstrated that approximately half of the reported earthquakes occur near the margins of the Newark Basin, far from the Ramapo fault, but still within the Ramapo fault system proper (Kafka, 1985) (Thurber, 1985) (Wheeler, 2006). In addition, a reassessment of the eastern U.S. stress field demonstrated that the present-day stress field is oriented east-southeast (Zoback, 1989a), which would be inconsistent with the previously inferred reverse reactivation of the fault. Kinematic analysis of fault zone samples collected from deep exploratory boreholes provides evidence that the latest style of deformation probably included extensional faulting during the Mesozoic (Ratcliffe, 1980) (Ratcliffe, 1982) (Burton, 1985) (Ratcliffe, 1990). The borehole data also confirm that the dip of the Ramapo fault is 10° to 15° shallower than inferred by Aggarwal and Sykes (Aggarwal, 1978).

In summary, several papers infer that evidence for Quaternary deformation exists near the Ramapo fault zone (Nelson, 1980) (Newman, 1983) (Newman, 1987) (Kafka, 1989); however, Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006) argue convincingly that none of the data used to infer seismic slip can be used to differentiate seismic from asesimic slip.

Additionally, trenches excavated across the up-dip projection of the fault zone revealed no evidence for Quaternary faulting (Stone, 1984) (Ratcliffe, 1990). Besides the presence of microseismicity within the vicinity of the Ramapo fault zone, there is no clear evidence of Quaternary tectonic faulting (Crone, 2000) (Wheeler, 2006), thus the Ramapo fault system is assigned a Class C designation by Crone and Wheeler (Crone, 2000). The Ramapo fault zone was a known structure for the EPRI study (EPRI, 1986). Based on the review of post-EPRI literature and seismicity, there is no new information developed since 1986 that would require a significant change to the EPRI seismic source model.

2.5.1.1.4.4.5.5 Kingston Fault

The Kingston fault is located in central New Jersey, approximately 175 mi (282 km) northeast of the CCNPP site (Figure 2.5.1-29). The Kingston fault is a 7 mi (11 km) long north to northeast-striking fault that offsets Mesozoic basement and is overlain by Coastal Plain sediments (Owens, 1998). Stanford (Stanford, 1995) use borehole and geophysical data to interpret a thickening of as much as 80 ft (24 m) of Pliocene Pennauken Formation across the surface projection of the Kingston fault. Stanford (Stanford, 1995) interprets the thickening of the Pennauken Formation gravel as a result of faulting rather than fluvial processes. Geologic cross sections prepared by Stanford (Stanford, 2002) do not show that the bedrock-Pennauken contact is vertically offset across the Kingston fault. Therefore, it seems reasonable to conclude that faulting of the Pennauken Formation is not required and that apparent thickening of the Pliocene gravels may represent a channel-fill from an ancient pre-Pliocene channel. Furthermore, Pleistocene glaciofluvial gravels that overlie the fault trace are not offset, thus indicating the fault is not a capable tectonic source (Stanford, 1995). Wheeler (Wheeler, 2006) reports that the available geologic evidence does not exclusively support a fault versus a fluvial origin for the apparent thickening of the Pennauken Formation. Wheeler (Wheeler, 2005) assigns the Kingston fault as a Class C feature based on a lack of evidence for Quaternary deformation. Given the absence of evidence for Quaternary faulting and the presence of undeformed Pleistocene glaciofluvial gravels overlying the fault trace, we conclude that the fault is not a capable tectonic feature.

2.5.1.1.4.4.5.6 New York Bight Fault

On the basis of seismic surveys, the New York Bight fault is characterized as an approximately 31 mile (50 km) long, north-northeast-striking fault, located offshore of Long Island, New York (Schwab, 1997a) (Schwab, 1997b) (Figure 2.5.1-29). The fault is located about 208 mi (335 km) northeast of the CCNPP site. Seismic reflection profiles indicate that the fault originated during the Cretaceous and continued intermittently with activity until at least the Eocene. The sense of displacement is northwest-side down and displaces bedrock as much as 280 ft (85 m), and Upper Cretaceous deposits about 150 ft (46 m) (PSEG, 2002). High-resolution seismic reflection profiles that intersect the surface projection of the fault indicate that middle and late Quaternary sediments are undeformed within a resolution of 3 ft (1 m) (Schwab, 1997a) (Schwab, 1997b). Only a few, poorly located earthquakes are spatially assoiciated within the vicinity of the New York Bight fault (Wheeler, 2006). Wheeler (Wheeler, 2006) defines the fault as a feature having insufficient evidence to demonstrate that faulting is Quaternary and assigns the New York Bight fault as a Class C feature. Based on the seismic reflection surveys of Schwab (Schwab, 1997a) (Schwab, 1997b) and the absence of Quaternary deformation, we conclude that the New York Bight fault is not a capable tectonic source.

2.5.1.1.4.4.5.7 Cacoosing Valley Earthquake Sequence

The 1993 to 1997 Cacoosing Valley earthquake sequence occurred along the eastern margin of the Lancaster seismic zone with the main shock occurring on January 16, 1994, near Reading, Pennsylvania, about 135 mi (217 km) north of the CCNPP site (Seeber, 1998) (Figure 2.5.1-29). This earthquake sequence also is discussed as part of the Lancaster seismic zone discussion (Section 2.5.1.1.4.5.2). The maximum magnitude earthquake associated with this sequence is an event of mbLg 4.6 (Seeber, 1998). Focal mechanisms associated with the main shock and aftershocks define a shallow subsurface rupture plane confined to the upper 1.5 mi (2.4 km) of the crust. It appears that the earthquakes occurred on a pre-existing structure striking N45°W in contrast to the typical north-trending alignment of microseismicity that delineates the Lancaster seismic zone. Seeber (Seeber, 1998) use the seismicity data, as well as the shallow depth of focal mechanisms, to demonstrate that the Cacoosing Valley earthquakes likely were caused by anthropogenic changes to a large rock quarry. Wheeler (Wheeler, 2006) defines the fault as a feature having insufficient evidence to demonstrate that faulting is Quaternary and assigns the Cacoosing Valley earthquake sequence as a Class C feature. Based on the findings of Seeber (Seeber, 1998), we interpret this earthquake sequence to be unrelated to a capable tectonic source.

2.5.1.1.4.4.5.8 New Castle County Faults

The New Castle faults are characterized as 3 to 4 mi (4.8 to 6.4 km) long buried north and northeast-striking faults that displace an unconformable contact between Precambrian to Paleozoic bedrock and overlying Cretaceous deposits. The faults are located in northern Delaware, near New Castle, about 97 mi (156 km) northeast of the CCNPP site (Figure 2.5.1-29). Spoljaric (Spoljaric, 1972) (Spoljaric, 1973) interprets the presence of the New Castle County faults using structural contours for the top of basement. On the basis of geophysical and borehole data, coupled with Vibroseis™ profiles, Spolijaric (Spolijaric, 1973) (Spolojaric, 1974) interprets a 1 mi (1.6 km) wide, N25°E-trending graben in basement rock. The graben is bounded by faults having displacements on the order of 32 to 98 ft (10 to 30 m) across the basement-Cretaceous boundary (Spoljaric, 1972). Also, there is a suggestion that the overlying Cretaceous deposits are tilted in a direction consistent with fault deformation; however, there is no direct evidence to indicate that these sediments are displaced. Sbar (Sbar, 1975) evaluates a 1973 M3.8 earthquake and its associated aftershocks, and note that the microseismicity defines a causal fault striking northeast and parallel to the northeast-striking graben of Spoljaric (Spoljaric, 1973). Subsequently, subsurface exploration by the Delaware Geological Survey (McLaughlin, 2002), that included acquisition of high resolution seismic reflection profiles, borehole transects, and paleoseismic trenching, provides evidence for the absence of Quaternary faulting on the New Castle faults. Wheeler (Wheeler, 2005) characterizes the New Castle County faults as a Class C feature. Based on McLaughlin (McLaughlin, 2002) there is strong evidence to suggest that the New Castle County faults as mapped by Spolijaric (Spolijaric, 1972) are not a capable tectonic source.

2.5.1.1.4.4.5.9 Upper Marlboro Faults

The Upper Marlboro faults are located in Prince George's County, Maryland, approximately 36 mi (58 km) northwest of the CCNPP site (Figure 2.5.1.-29). These faults were first shown by Dryden (Dryden, 1932) as a series of faults offsetting Coastal Plain sediments. The faults were apparently exposed in a road cut on Crain Highway at 3.3 mi (5.3 km) south of the railroad crossing in Upper Marlboro, Maryland (Prowell, 1983). Two faults displace Miocene and Eocene sediments and a third fault is shown offsetting a Pleistocene unit. These faults are not observed beyond this exposure. No geomorphic expression has been reported or was noticed during field reconnaissance for the CCNPP Unit 3 study. Based on a critical review of available literature, Wheeler (Wheeler, 2006) re-interprets the Upper Marlboro faults as likely related to surficial landsliding because of the very low dips and concavity of the fault planes. The Marlboro faults are classified by Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006), as a Class C feature based on a lack of evidence for Quaternary faulting. Given the absence of seismicity along the fault, lack of published literature documenting Quaternary faulting, coupled with the interpretation of Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006), we conclude that the Upper Marlboro faults are not a capable tectonic source.

2.5.1.1.4.4.5.10 Lebanon Church Fault

The Lebanon Church fault is a poorly-known northeast-striking reverse fault located in the Appalachian Mountains of Virginia, near Waynesboro, about 119 mi (192 km) southwest of the CCNPP site (Prowell, 1983) (Figure 2.5.1-29). The fault is exposed in a single road cut along U.S. Route 250 as a small reverse fault that offsets Miocene-Pliocene terrace gravels up to as about 5 ft (1.5 m) (Prowell, 1983). The terrace gravels overlie Precambrian metamorphic rocks of the Blue Ridge Province. An early author (Nelson, 1962) considered the gravels to be Pleistocene, whereas Prowell (1983) interprets the gravel to be Miocene to Pliocene. Wheeler (Wheeler, 2006) classifies the Lebanon Church fault as a Class C feature having insufficient evidence to demonstrate that faulting is Quaternary. As part of this CCNPP Unit 3 study, inquiries with representatives with the Virginia Geological Survey and United States Geological Survey indicate that there is no new additional geologic information on this fault. Based on literature review, discussion with representatives with Virginia Geological Survey, as well as the absence of seismicity spatially associated with the feature, we conclude that the Lebanon Church fault is not a capable tectonic source.

2.5.1.1.4.4.5.11 Hopewell Fault

The Hopewell fault is located in central Virginia, approximately 89 mi (143 km) southwest of the CCNPP site (Figure 2.5.1.-29). The Hopewell fault is a 30 mi (48 km) long, north-striking, steeply east-dipping reverse fault (Mixon, 1989) (Dischinger, 1987). The fault was originally named the Dutch Gap fault by Dischinger (Dischinger, 1987), and was renamed the Hopewell fault by Mixon (Mixon, 1989). The fault displaces a Paleocene-Cretaceous contact and is inferred to offset the Pliocene Yorktown Formation (Dischinger, 1987). Mixon (Mixon, 1989) extend the mapping of Dischinger (Dischinger, 1987), but include conflicting data regarding fault activity. For instance, a cross section presented by Mixon (Mixon, 1989) shows the Hopewell fault displacing undivided upper Tertiary and Quaternary units, whereas the geologic map used to produce the section depicts the fault buried beneath these units. A written communication from Newell (Wheeler, 2006) explains that the Hopewell fault was not observed offsetting Quaternary deposits and the representation of the fault in the Mixon (Mixon, 1989) cross section is an error. Thus, the Hopewell fault zone is assigned as a Class C feature because no evidence is available to demonstrate Quaternary surface deformation. Based on the written communication of Newell (Wheeler, 2006), an absence of published literature documenting Quaternary faulting, and an absence of seismicity spatially associated with the feature, we conclude that the Hopewell fault is not a capable tectonic source.

2.5.1.1.4.4.5.12 Old Hickory Faults

The Old Hickory faults are located near the Fall Line in southeastern Virginia, approximately 115 mi (185 km) south-southwest of the CCNPP site (Figure 2.5.1.-29). Based on mining exposures of the Old Hickory Heavy Mineral deposit, the Old Hickory faults consist of a series of five northwest-striking reverse faults that offset Paleozoic basement and Pliocene Coastal Plain sediments. The northwest-striking reverse faults juxtapose Paleozoic Eastern Slate Belt diorite over the Pliocene Yorktown Formation (Berquist, 1999). Strike lengths range between 330 to

490 ft (100 to 150 m) and are spaced about 164 ft (50 m) apart. Berquist and Bailey (Berquist, 1999) report up to 20 ft (6 m) of oblique dip-slip movement on individual faults, and suggest that the faults may be reactivated Mesozoic structures. There is no stratigraphic or geomorphic evidence of Quaternary or Holocene activity of the Old Hickory faults (Berquist, 1999). Crone and Wheeler (Crone, 2000) and Wheeler (Wheeler, 2006) conclude that "no Quaternary fault is documented" and assign a Class C designation to the Old Hickory faults. Based on the absence of published literature documenting the presence of Quaternary deformation, and the absence of seismicity spatially associated with this feature, we conclude that the Old Hickory faults are not a capable tectonic source.

2.5.1.1.4.4.5.13 Stanleytown-Villa Heights Faults

The postulated Stanleytown-Villa Heights faults are located in the Piedmont of southern Virginia, approximately 223 mi (359 km) southwest of the CCNPP site (Figure 2.5.1-29). The approximately 660 ft long (201 m long) faults juxtapose Quaternary alluvium against rocks of Cambrian age, and reflect an east-side-down sense of displacement (Crone, 2000). No other faults are mapped nearby (Crone, 2000). Geologic and geomorphic evidence suggests the "faults" are likely the result of landsliding. Crone and Wheeler (Crone, 2000) classify the Stanleytown-Villa Heights faults as a Class C feature based on lack of evidence for Quaternary faulting. Based on the absence of published literature documenting the presence of Quaternary faulting, and the absence of seismicity spatially associated with this feature, we conclude that the Stanleytown-Villa Heights faults are not a capable tectonic source.

2.5.1.1.4.4.5.14 East Coast Fault System

The postulated East Coast fault system (ECFS) of Marple and Talwani (2000) trends N34°E and is located approximately 70 mi (113 km) southwest of the site (Figure 2.5.1-29). The 370 mi (595 km) long fault system consists of three approximately 125 mi (201 km) long segments extending from the Charleston area in South Carolina northeastward to near the James River in Virginia (Figure 2.5.1-29). The three segments were initially referred to as the southern, central, and northern zones of river anomalies (ZRA-S, ZRA-C, ZRA-N) and are herein referred to as the southern, central and northern segments of the ECFS. The southern segment is located in South Carolina; the central segment is located primarily in North Carolina. The northern segment, buried beneath Coastal Plain deposits, extends from northeastern North Carolina to southeastern Virginia, about 70 mi (113 km) southwest of the CCNPP site. Marple and Talwani (Marple, 2000) map the northern terminus of the ECFS between the Blackwater River and James River, southeast of Richmond. Identification of the ECFS is based on the alignment of geomorphic features along Coastal Plain rivers, areas suggestive of uplift, and regions of local faulting. The right-stepping character of the three segments, coupled with the northeast orientation of the fault system relative to the present day stress field, suggests a right-lateral strike-slip motion for the postulated ECFS (Marple and Talwani, 2000).

The southern segment of the fault system, first identified by Marple and Talwani (1993) as an approximately 125 mi (201 km) long and 6 to 9 mi (10 to 14.5 km) wide zone of river anomalies, has been attributed to the presence of a buried fault zone. The southern end of this segment is associated with the Woodstock fault, a structure defined by fault-plane solutions of microearthquakes and thought to be the causative source of the 1886 Charleston earthquake (Marple, 2000). The southern segment is geomorphically the most well-defined segment of the fault system and is associated with micro-seismicity at its southern end. This segment was included as an alternative geometry to the areal source for the 1886 Charleston earthquake in the 2002 USGS hazard model (Section 2.5.2) for the National Seismic Hazard Mapping Project (Frankel, 2002).

Crone and Wheeler (Crone, 2000) do not include the central and northern segments of the ECFS in their compilation of potentially active Quaternary faults. The segments also were not presented in workshops or included in models for the Trial Implementation Project (TIP), a study that characterized seismic sources and ground motion attenuation models at two nuclear power plant sites in the southeastern United States (Savy, 2002). As a member of both the USGS and TIP workshops, Talwani did not propose the northern and central segments of the fault system for consideration as a potential source of seismic activity. There is no pre-EPRI or post-EPRI seismicity spatially associated with the northern and central segments of the fault system.

Recent geologic and geomorphic analysis of stream profiles across sections of the ECFS, and critical evaluation of Marple and Talwani (Marple, 2000) for the North Anna ESP, provides compelling evidence that the northern segment of the ECFS, which lies nearest to the CCNPP site, has a very low probability of existence (Dominion, 2004b). Wheeler (Wheeler, 2005) states that although the evidence for a southern section of the ECFS is good, there is less evidence supporting Quaternary tectonism along the more northerly sections of the ECFS, and designates the northern portion of the fault system as a Class C feature.

In the Safety Evaluation Report for the North Anna ESP site, the NRC staff agreed with the assessment of the northern segment of the East Coast Fault System (ECFS-N) presented by the North Anna applicant (NRC, 2005). Based on their independent review, the NRC staff concluded that:

- "Geologic, seismologic, and geomorphic evidence presented by Marple and Talwani is questionable."
- "The majority of the geologic data cited by Marple and Talwani in support of their postulated ECFS apply only to the central and southern segments."
- There are "no Cenozoic faults or structure contour maps indicating uplift along the ECFS-N."
- "The existence and recent activity of the northern segment of the ECFS is low."

Despite the statements above, the NRC concluded that the ECFS-N could still be a contributor to the seismic hazard at the North Anna site and should be included in the ground motion modeling to determine the Safe Shutdown Earthquake. The NRC agreed with the 10% probability of existence and activity proposed in the North Anna ESP application. The results of the revised ground motion calculations indicate that the ECFS-N does not contribute to the seismic hazard at the North Anna ESP site. The CCNPP site is approximately 70 mi (113 km) northeast of the ECFS-N, or 7 mi (11 km) further away than the North Anna site is from the ECFS-N. Based on the above discussion and the large distance between the site and the ECFS-N, this fault is not considered a contributing seismic source and need not be included in the seismic hazard calculations for the CCNPP site.

Marple and Talwani (Marple, 2004) suggest a northeast extension of the ECFS of Marple and Talwani (Marple, 2000), based on existing limited geologic, geophysical and geomorphic data. Marple and Talwani (Marple, 2004) postulate that the northern ECFS may step left (northwest) to the Stafford fault system near northern Virginia and southern Maryland (Figure 2.5.1-29) and thus extending the ECFS along the Stafford fault up to New York. As stated in Section 2.5.1.1.4.4.4.1, the NRC (NRC, 2005) agreed with an analysis of the Stafford fault performed as part of the North Anna ESP application and states: "Based on the evidence cited by the applicant, in particular the applicant's examination of the topography profiles that cross the fault system, the staff concludes that the applicant accurately characterized the Stafford fault system as being inactive during the Quaternary Period."

In summary, the ECFS in its entirety represents a new postulated tectonic feature that was not known to the EPRI Earth Science Teams in 1986. The 1986 EPRI models include areal sources to model the Charleston seismic source; therefore, the southern segment of the East Coast fault system is in essence covered by the different Charleston sources zone geometries. A review of the seismic sources that contribute 99% of the seismic hazard to the CCNPP shows that the Charleston source is not a contributor. The central and northern segments of the ECFS represent a new tectonic feature in the Coastal Plain that postdates the EPRI studies. The closest approach of the northern segment to the site is approximately 77 mi (124 km) as described above. Although the postulated ECFS represents a potentially new tectonic feature in the Coastal Plain of Virginia and North Carolina (Marple, 2000), current interpretations of the ECFS based on existing data indicate that the fault zone probably does not exist (especially the northern segment) and, if it does exists, has a very low probability of activity and does not contribute to hazard at the site.

2.5.1.1.4.5 Seismic Sources Defined by Regional Seismicity

Within 200 mi (322 km) of the CCNP site, two potential seismic sources are defined by a concentration of small to moderate earthquakes. These two seismic sources include the Central Virginia seismic zone in Virginia and the Lancaster seismic zone in southeast Pennsylvania, both of which are discussed below (Figure 2.5.1-29).

2.5.1.1.4.5.1 Central Virginia Seismic Zone

The Central Virginia seismic zone is an area of persistent, low level seismicity in the Piedmont Province (Figure 2.5.1-23 and Figure 2.5.1-29). The zone extends about 75 mi (121 km) in a north-south direction and about 90 mi (145 km) in an east-west direction from Richmond to Lynchburg and is coincident with the James River (Bollinger, 1985). The CCNPP site is located 47 to 62 mi (76 to 100 km) northeast of the northern boundary of the Central Virginia seismic zone. The largest historical earthquake to occur in the Central Virginia seismic zone was the body-wave magnitude (mb) 5.0 Goochland County event on December 23, 1875 (Bollinger, 1985). The maximum intensity estimated for this event was Modified Mercalli Intensity (MMI) VII in the epicentral region. More recently, an mb 4.5 earthquake (two closely-spaced events that when combined = Mw 4.1) occurred on December 9, 2003 within the Central Virginia seismic zone (Kim and Chapman, 2005). The December 9, 2003 earthquake occurred close to the Spotsylvania fault, but due to the uncertainty in the location of the epicenter (3.7 to 5 mi (6 to 8 km)), no attempt could be made to locate the epicenter with a specific fault or geologic lineament in the CVSZ (Kim, 2005).

Seismicity in the Central Virginia seismic zone ranges in depth from about 2 to 8 mi (3 to 13 km) (Wheeler, 1992). It is suggested (Coruh, 1988) that seismicity in the central and western parts of the zone may be associated with west-dipping reflectors that form the roof of a detached antiform, while seismicity in the eastern part of the zone near Richmond may be related to a near-vertical diabase dike swarm of Mesozoic age. However, given the depth distribution of 2 to 8 mi (3 to 13 km) (Wheeler, 1992) and broad spatial distribution, it is difficult to uniquely attribute the seismicity to any known geologic structure and it appears that the seismicity extends both above and below the Appalachian detachment.

No capable tectonic sources have been identified within the Central Virginia seismic zone, but two paleo-liquefaction sites have been identified within the seismic zone (Crone, 2000) (Obermier, 1998). The presence of these paleo-liquefaction features on the James and Rivanna Rivers shows that the Central Virginia seismic zone reflects both an area of paleo-seismicity as well as observed historical seismicity. Based on the absence of widespread paleo-liquefaction, however, it was concluded (Obermier, 1998) that an earthquake of magnitude 7 or larger has not occurred within the seismic zone in the last 2,000 to 3,000 years, or in the eastern portion of the seismic zone for the last 5,000 years. It was also conclude that the geologic record of one or more magnitude 6 or 7 earthquakes might be concealed between streams, but that such events could not have been abundant in the seismic zone. In addition, these isolated locations of paleo-liquefaction may have been produced by local shallow moderate magnitude earthquakes of M 5 to 6.

The paleo-liquefaction sites reflect pre-historical occurrences of seismicity within the Central Virginia seismic zone, and do not indicate the presence of a capable tectonic source. Recently, Wheeler (Wheeler, 2006) hypothesizes that there may be two causative faults for the small dikes of Obermier and McNulty (Obermier, 1998), and that earthquakes larger than those represented by historic seismicity are possible; whereas Marple and Talwani (Marple, 2004) interpret seismicity data to infer the presence of a hypothesized northwest-trending basement fault (Shenandoah fault) that coincides with the Norfolk fracture zone (Marple, 2004). However, no definitive causative fault or faults have been identified within the Central Virginia seismic zone (Wheeler, 2006).

The 1986 EPRI source model includes various source geometries and parameters to capture the seismicity of the Central Virginia seismic zone. Subsequent hazard studies have used maximum magnitude (Mmax) values that are within the range of maximum magnitudes used by the six EPRI models. Collectively, upper-bound maximum values of Mmax used by the EPRI teams range from mb 6.6 to 7.2 (Section 2.5.2.2). More recently, Bollinger (Bollonger, 1992) has estimated a Mmax of mb 6.4 for the Central Virginia seismic source. Chapman and Krimgold (Chapman, 1994) have used a Mmax of mb 7.25 for the Central Virginia seismic source and most other sources in their seismic hazard analysis of Virginia. This more recent estimate of Mmax is similar to the Mmax values used in the 1986 EPRI studies. Similarly, the distribution and rate of seismicity in the Central Virginia seismic source have not changed since the 1986 EPRI study (Section 2.5.2.2.8). Thus, there is no change to the source geometry or rate of seismicity. In 2005, the NRC agreed with the findings of the North Anna ESP application's assessment of the Central Virginia seismic zone (NRC, 2005). Therefore, the conclusion is that no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.4.5.2 Lancaster Seismic Zone

The Lancaster seismic zone, as defined by Armbruster and Seeber (Armbruster, 1987), of southeast Pennsylvania has been a persistent source of seismicity for at least two centuries. The seismic zone is about 80 mi (129 km) long and 80 mi (129 km) wide and spans a belt of allochthonous Appalachian crystalline rocks between the Great Valley and Martic Line about 111 mi (179 km) northwest of the CCNPP site (Figure 2.5.1-29). The Lancaster seismic zone crosses exposed Piedmont rocks that include thrust faults and folds associated with Paleozoic collisional orogenies. It also crosses the Newark-Gettysburg Triassic rift basin which consists of extensional faults associated with Mesozoic rifting. Most well-located epicenters in the Lancaster seismic zone lie directly outside the Gettysburg-Newark basin (Scharnberger, 2006). The epicenters of 11 events with magnitudes 3.04 to 4.61 rmb from 1889 to 1994 from the western part of Lancaster seismic zone define a north-south trend that intersects the juncture between the Gettysburg and Newark sub-basins. This juncture is a hinge around which the two sub-basins subsided, resulting in east-west oriented tensile stress. Numerous north-south trending fractures and diabase dikes are consistent with this hypothesis. It is likely that seismicity in at least the western part of the Lancaster seismic zone is due to present-day northeast-southwest compressional stress which is activating the Mesozoic fractures, with dikes perhaps serving as stress concentrators (Armbruster, 1987).

It also is probable that some recent earthquakes in the Lancaster seismic zone have been triggered by surface mining. For instance, the 16 January 1994 Cacoosing earthquake (mb 4.6) is the largest instrumented earthquake occurring in the Lancaster seismic zone (Section 2.5.1.1.4.4.5.7). This event was part of a shallow (depths generally less than 1.5 mi (2.4 km)) earthquake sequence linked to quarry activity (Seeber, 1998). The earthquake sequence that culminated in the January 16 event initiated after a quarry was shut down and the quarry began to fill with water. Seeber (Seeber, 1998) interprets the reverse-left lateral oblique earthquake sequence to be due to a decrease in normal stress caused by quarrying followed by an increase in pore fluid pressure (and decrease in effective normal stress) when the pumps were turned off and the water level increased.

Prior to the Cacoosing earthquake sequence, the 23 April 1984 Martic earthquake (mb 4.1) was the largest instrumented earthquake in the seismic zone and resembles pre-instrumental historical events dating back to the middle 18th century. The 1984 earthquake sequence appears centered at about 2.8 mi (4.5 km) in depth and may have ruptured a steeply east-dipping, north-to northeast-striking fault aligned subparallel to Jurassic dikes with a reverse-right lateral oblique movement, consistent with east-northeast horizontal maximum compression. These dikes are associated with many brittle faults and large planes of weakness suggesting that they too have an effect on the amount of seismicity in the Lancaster seismic zone. Most of the seismicity in the Lancaster seismic zone is occurring on secondary faults at high angles to the main structures of the Appalachians. The EPRI study (EPRI, 1986) source models do not identify the Lancaster seismic zone as a separate seismic source. However, the 5.3 to 7.2 Mb maximum magnitude distributions of EPRI source zones are significantly greater than any reported earthquake in this Lancaster seismic zone. Thus, the EPRI study (EPRI, 1986) models adequately characterized this region and no significant update is required.}

2.5.1.2 Site Geology

2.5.1.2.1 Site Area Physiography and Geomorphology

{The CCNPP site area is located within the Western Shore Uplands of the Atlantic Coastal Plain Physiographic Province and is bordered by the Chesapeake Bay to the east and the Patuxent River to the west (Figure 2.5.1-3 and Figure 2.5.1-6).

The site vicinity geologic map (Figure 2.5.1-26a and Figure 2.5.1-26b), compiled from the work of several investigators, indicates that the counties due east from the CCNPP site across Chesapeake Bay are underlain by Pleistocene to Recent sands. Most of the site vicinity is underlain by Tertiary Coastal Plain deposits. Quaternary to Recent alluvium beach deposits and terrace deposits are mapped along streams and estuaries. Quaternary terrace and Lowland deposits are shown in greater detail on the scale of the site area geologic map (Figure 2.5.1-30). Geologic cross sections in the site area indicate that the Tertiary Upland deposits are underlain by gently dipping Tertiary Coastal Plain deposits described in Section 2.5.1.2.2 (Figure 2.5.1-31).

The topography within 5 mi (8 km) of the site consists of gently rolling hills with elevations ranging from about sea level to nearly 130 ft (40 m) msl (Figure 2.5.1-3). The site is well-drained by short, ephemeral streams that form a principally dendritic drainage pattern with many streams oriented in a northwest-southeast direction (Figure 2.5.1-4). As shown on the site area and site topographic and geological maps, the ground surface above approximately 100 ft (30 m) msl is capped by the Upper Miocene-Pliocene Upland deposits (Figure 2.5.1-3, Figure 2.5.1-4, Figure 2.5.1-30, and Figure 2.5.1-31). These deposits occupy dissected upland areas of the Cove Point quadrangle in which the CCNPP site is located (Figure 2.5.1-30 and Figure 2.5.1-31) (Glaser, 2003a). The longest stream near the site is Johns Creek, which is approximately 3.5 mi (5.6 km) long before it drains into St. Leonard Creek (Figure 2.5.1-3 and Figure 2.5.1-32). The ephemeral stream channels near the CCNPP site are either tributary to Johns Creek or flow directly to the Chesapeake Bay. These stream channels maintain their dendritic pattern as they cut down into the underlying Choptank and St. Marys Formations (Figure 2.5.1-26a, Figure 2.5.1-30 and Figure 2.5.1-31).

The Chesapeake Bay shoreline forms the eastern boundary of the CCNPP site and generally consists of steep cliffs with narrow beach at their base. The cliffs reach elevations of about 100 ft (30 m) msl along the eastern portion of the site's shoreline. Narrow beaches whose width depends upon tidal fluctuations generally occur at the base of the cliffs. Field observations indicate that these steep slopes fail along nearly vertical irregular surfaces. The slope failure appears to be caused by shoreline erosion along the base of the cliffs. Shoreline processes and slope failure along Chesapeake Bay are discussed in Section 2.4.9. Approximately 2500 ft (762 m) of the shoreline from the existing CCNPP Units 1 and 2 intake structure southward to the existing barge jetty is stabilized against shoreline erosion (Figure 2.4.9-2). The CCNPP Unit 3 will be constructed at a final grade elevation of approximately 85 ft (26 km) msl and will be set back approximately 1,000 ft (305 m) from the Chesapeake Bay shoreline.

As described in Section 2.5.1.1.1, the Chesapeake Bay was formed toward the end of the Wisconsinan glacial stage, which marked the end of the Pleistocene epoch. As the glaciers retreated, the huge volumes of melting ice fed the ancestral Susquehanna and Potomac Rivers, which eroded older Coastal Plain deposits forming a broad river valley. The rising sea level covered the Continental Shelf and reached the mouth of the Bay about 10,000 years ago. Sea level continued to rise, eventually submerging the area now known as the Susquehanna River Valley prior to sea level dropping to the current elevation. The Bay assumed its present dimensions about 3000 years ago (Section 2.4.9).}

2.5.1.2.2 Site Area Geologic History

{The site area geologic history prior to the early Cretaceous is inferred from scattered borehole data, geophysical surveys and a synthesis of published information. Sparse geophysical and borehole data indicate that the basement rock beneath the site may consist of exotic crystalline magmatic arc material (Glover, 1995b). Although the basement has not been penetrated directly beneath the site with drill holes, regional geologic cross sections developed from geophysical, gravity and aeromagnetic, as well as limited deep borehole stratigraphic data beyond the site area, suggest Precambrian and Paleozoic crystalline rocks are most likely present at a depth of about 2,600 ft (792 m) beneath the site (Section 2.5.1.2.3 and Section 2.5.1.2.4). Tectonic models discussed in Section 2.5.1.2.4 hypothesize that the crystalline basement was accreted to the pre-Taconic North American margin during the Paleozoic along a suture that lies about 10 mi (16 km) west of the site (Figure 2.5.1-16 and Figure 2.5.1-22). Therefore, the crystalline basement beneath the Coastal Plain sediments in the site area might consist of an accreted nappe-like block of Carolina-Chopawamsic magmatic arc terrane with windows of Laurentian Grenville basement (Figure 2.5.1-15 and Figure 2.5.1-16) (Klitgord, 1995).

As discussed in Section 2.5.1.1.2 and Section 2.5.1.2.4, Mesozoic rift basins are exposed in the Piedmont Physiographic Province and are buried beneath Coastal Plain sediments. The Queen Anne Basin was originally postulated by Hansen (1988) and was considered to underlie the site (Horton, 1991). However, this interpretation does not appear to be supported by most of the borehole data and current interpretations (Section 2.5.1.2.4).

During the early Cretaceous, sands, clays, sandy clays, and arkosic sands of the Arundel/Patuxent Formations (undivided) were deposited on the crystalline basement in a continental and fluviatile environment. Individual beds of sand or silt grade rapidly into sediments with different compositions or gradations, both vertically and horizontally, which suggests they were deposited in alluvial fan or deltaic environments. Clay layers containing carbonized logs, stumps and other plant remains indicate the existence of quiet-water, swamp environments between irregularly distributed stream channels. Thicker clays near the top of this unit in St Mary's County are interpreted to indicate longer periods of interfluvial quiet water deposition (Hansen, 1984).

The overlying beds of the Patapsco Formation are similar to the deposits in the Arundel/Patuxent (undivided) formations and consist chiefly of materials derived from the eroded crystalline rocks of the exposed Piedmont to the west and reworked Lower Cretaceous sediments. These sediments were deposited in deltaic and estuarine environments with relatively low relief. The Upper Cretaceous Raritan Formation appears to be missing from the site area due either to non-deposition or erosion on the northern flank of the structurally positive Norfolk Arch.

The Magothy Formation represents deposits from streams flowing from the Piedmont and depositing sediments in the coastal margins of the Upper Cretaceous sea. Subsequent uplift and tilting of the Coastal Plain sediments mark the end of continental deposition and the beginning of a marine transgression of the region. This contact is a regional unconformity marked in places by a basal layer of phosphatic clasts in the overlying Brightseat Formation.

During the Early Paleocene Epoch, the Brightseat Formation marks a marine advance in the Salisbury embayment (Ward, 2004). Uplift or sea level retreat is indicated by the burrowed contact (unconformity) of the Brightseat Formation with the overlying Aquia Formation. The marine Aquia Formation which is noted for its high glauconite content and shell beds was deposited in a shoaling marine environment indicated by a generally coarsening upward lithology (Hansen, 1996). A mix of light-colored quartz grains and greenish to blackish glauconite grains and iron staining indicated the change to a sandbank facies in the upper Aquia formation (Hansen, 1996). A marine transgression during the Late Paleocene/Early Eocene into the central portion of the Salisbury Embayment deposited the Marlboro Clay (Ward, 2004). During the Early Eocene, a moderately extensive marine transgression deposited the Potopaco Member of the Nanjemoy Formation. A subsequent transgression deposited the Woodstock Member of the Nanjemoy Formation (Ward, 2004). The most extensive marine transgression during the middle Eocene resulted in the deposition of the Piney Point Formation (Ward, 2004). The site area may have been emergent during the Oligocene as the Late Oligocene Old Church Formation indicates sea level rise and submergence to the north and south of the site area (Ward, 2004). A brief regression was followed by nearly continuous sedimentation in the Salisbury Embayment punctuated by short breaks, resulting in a series of thin, unconformity-bounded beds (Ward, 2004). A series of marine transgressions into the Salisbury Embayment during the Miocene produced the Calvert, Choptank and St. Marys Formations. Pliocene and Quaternary geologic history is discussed in Section 2.5.1.2.1.}

2.5.1.2.3 Site Area Stratigraphy

{Site specific information on the stratigraphy underlying the CCNPP site is limited by the total depths of the various borings advanced by site investigators over the years. Only a few scattered borings have been advanced below the Aquia Formation (Hansen, 1986). The deepest boring known to have been advanced at the site is CA-Ed 22 which was drilled to a total depth of 789 ft (240 m) and completed as a water supply well in 1968 (Hansen, 1996). This boring penetrates the full Tertiary stratigraphic section and intersects the contact between the Tertiary and the Cretaceous section at the base of the Aquia Formation. The closest boring which advances to pre-Cretaceous bedrock is approximately 13 mi (21 km) south of the site at

Lexington Park in St. Mary's County, (Figure 2.5.1-10) (Hansen, 1986). This boring cored a diabase dike in the pre-Cretaceous basement (Section 2.5.1.1.3). The few borings that have reached basement rock in the site area are widely scattered (Figure 2.5.1-10) but the majority indicates that the basement rock beneath the site is likely to be similar to the schists and gneisses found in the Piedmont Physiographic Province approximately 50 mi (80 km) to the west (Figure 2.5.1-1). Alternatively, this crystalline basement might have been accreted to the exposed Piedmont as a result of continental collision during a Paleozoic orogeny (Section 2.5.1.1.1.4 and Section 2.5.1.2.2). Figure 2.5.1-33 shows the locations of the various borings at the site and identifies those completed as either water supply wells or observation wells. Many of these borings were drilled to 200 ft (61 m) in total depth; six were advanced to a total depth of 400 ft (122 m). Figure 2.5.1-34 is a site-specific stratigraphic column based on correlations by Hansen (Hansen, 1996), Achmad and Hansen (Achmad, 1997) and Ward and Powars (Ward, 2004).

The CCNPP site is located on Coastal Plain sediments ranging in age from Lower Cretaceous to Recent, which, in turn, were deposited on the pre-Cretaceous basement rock. The Cretaceous section shown on the site stratigraphic column is projected to the site from proximal borings which intersect the pre-Cretaceous basement (Figure 2.5.1-12).

Coastal Plain sediments were deposited in a broad basement depression known as the Salisbury Embayment extending from eastern Virginia to southern New Jersey (Figure 2.5.1-11) (Ward, 2004). These sediments were deposited during periods of marine transgression/ regression and exhibit lateral and vertical variation in both lithology and texture.

2.5.1.2.3.1 Lower Cretaceous Potomac Group and pre-Potomac sediments

As discussed in Section 2.5.1.1.3, Hansen and Wilson (Hansen, 1984) assign the lowermost 30 ft (9 m) of the Lexington Park well (SM-Df 84), 13 mi (21 km) south of the CCNPP site (Figure 2.5.1-10) (Hansen, 1986), to the Waste Gate formation. These sediments are described as gray silts and clays, interbedded with fine to medium silty fine to medium sands. Although these sediments might correlate with the Waste Gate Formation identified in a well in Crisfield, Maryland (Do-CE 88), east of the Chesapeake Bay (Figure 2.5.1-10), there is no direct evidence indicating whether this unit occurs beneath the CCNPP site.

The Potomac Group is comprised of a sequence of interbedded sands and silty to fine sandy clays. Because this formation was not encountered by any borings drilled at the CCNPP site, the description of these units is based on published data (Hansen, 1984) (Achmad, 1997). Regionally, the Potomac Group consists of, from oldest to youngest, the Patuxent Formation, the Arundel Formation and the Patapsco Formation. These units are considered continental in origin and are in unconformable contact with each other.

The Lower Cretaceous Patuxent Formation consists of a sequence of variegated sands and clays which form a major aquifer in the Baltimore area, approximately 50 mi (80 km) up-dip from the site, but which have not been tested in the vicinity of the site. The nearest well intercepting the Patuxent is approximately 13 mi (21 km) south of the site and here the formation contains much less sand than is found in the upper part of the Potomac Group. The Patuxent is approximately 600 to 700 ft (182 m to 213 m) thick and is overlain by the Arundel/ Patapsco formations (undivided)

In the Baltimore area, the Arundel Formation consists of clays which are brick red near the Fall Line. Further down-dip toward the southeast, the color changes to gray and this unit is difficult to separate in the subsurface from those clays present in the underlying Patuxent and overlying Patapsco formations. Consequently, the Arundel and the Patuxent are often undivided (Hansen, 1984) in the literature and referred as the Arundel/Patuxent formations (undivided).

Hansen and Wilson (Hansen, 1984) describe the upper portion of the Arundel/Patuxent formations (undivided) as variegated silty clay with thin very fine sand and silt interbeds that may be as thick as 150 to 200 ft (46 to 61 m) beneath the CCNPP site (Figure 2.5.1 12). The Arundel Formation is not recognized in southern Maryland (Hansen, 1996).

2.5.1.2.3.2 Upper Cretaceous Formations

The Patapsco formation is the uppermost unit in the Potomac Group and consists of gray, brown and red variegated silts and clays interbedded with lenticular, cross-bedded clayey sands and minor gravels. This formation is a major aquifer near the Fall Line in the Baltimore area, but the Patapsco is untested near the CCNPP site. The thickness of the Patapsco Formation based on regional correlations is 1,000 to 1,100 ft thick beneath the CCNPP site.

The Mattaponi (?) formation described as overlying the Potomac group in Hansen and Wilson (Hansen, 1984) is no longer recognized by the Maryland Geological Survey. The section formerly assigned to the Mattaponi (?) has been included within the Patapsco Formation.

The Magothy Formation unconformably overlies the Patapsco Formation beneath the site. The Magothy is comprised chiefly of pebbly, medium coarse sand, although there are clayey portions in the upper part (Achmad, 1997). This formation is much thinner at the site than further north in Calvert County and pinches out within a few mi to the south (Achmad, 1997). The Monmouth and Matawan formations have not been differentiated from the Magothy Formation in the site area.

2.5.1.2.3.3 Tertiary Formations

The earliest Tertiary sediments beneath the site are assigned to the Lower Paleocene Brightseat Formation, a thin dark gray sandy clay identified in the deepest boring (CA-Ed 22) at the site as the Lower Confining Unit (Figure 2.5.1-12). The Brightseat Formation is identified in the gamma log as a higher than normal gamma response below the Aquia sand. According to Ward and Powars (Ward, 2004) the Brightseat Formation marks a marine advance in the Salisbury Embayment and occurs principally in the northeastern portion of the Embayment. This stratigraphic unit was reached by the water supply well CA-Ed 22 in 1968 (Figure 2.5.1-12). Achmad and Hansen (Achmad, 1997) describe the Brightseat Formation as approximately 10 ft (3 m) thick consisting mainly of very fine sand and clay with a bioturbated fabric. The absence of a bioturbated contact with the underlying beds suggests an unconformable contact.

The Aquia Formation unconformably overlies the Brightseat Formation and consists of clayey, silty, very shelly glauconitic sand (Ward, 2004). Microfossil study has placed the Aquia in the upper Paleocene. In the type section, the Aquia Formation is divided into two members, the Piscataway Creek and the Paspotansa, but at the CCNPP site, these members are not differentiated. Achmad and Hansen (Achmed, 1997) describe the Aquia Formation as approximately 150 ft (46 m) thick. The sand becomes fine-grained in the lower 50 ft (15 m) of the formation.

The Marlboro clay is a silvery-gray to pale-red plastic clay interbedded with yellowish-gray to reddish silt occurring at the base of the Nanjemoy Formation (Ward, 2004). Achmad and Hansen (1997) describe approximately 10 ft (3 m) of clay with thin, indistinct laminae of differing colored silt. Its contact with the underlying Aquia Formation is somewhat gradational while the contact between the Marlboro and the overlying Nanjemoy appears to be sharp indicating that the Nanjemoy unconformably overlies the Marlboro. Microfossil studies indicate the presence of a mixture of very late Paleocene and very early Eocene flora. Based on geophysical logs from CA-Ed 22, the Marlboro clay appears to be approximately 15 ft (4.6 m) thick beneath the CCNPP site (Figure 2.5.1-12).

At the CCNPP site, the Nanjemoy Formation is divided into the Potapaco and Woodstock members between the overlying Piney Point Formation and the underlying Marlboro clay. The Nanjemoy Formation is described as olive black, very fine grained, well-sorted silty glauconitic sands (Ward, 2004). Based on electric log data, the thickness of the Nanjemoy Formation beneath the CCNPP site is approximately 180 ft (55 m). About 80 ft (24 m) of this unit was penetrated by CCNPP Unit 3 borings, B-301 and B-401 (Figure 2.5.1-35 and Figure 2.5.1-36), drilled during the subsurface investigation.

The Piney Point Formation is a thin glauconitic sand and clay unit unconformably overlying the Nanjemoy formation. According to Achmad and Hansen (Achmad, 1997), the Piney Point is approximately 20 ft (6 m) thick at the CCNPP site and extends from about the middle of Calvert County, north of the CCNPP site, toward the south to beyond the Potomac River; increasing in thickness to approximately 130 ft (40 m) at Point Lookout at the confluence of the Potomac River and Chesapeake Bay. Formerly considered late Eocene in age, the Piney Point is assigned to the middle Eocene (Achmad, 1997) (Ward, 2004). The unit has a distinctive natural gamma signature associated with the presence of glauconite and is a useful marker bed.

This distinctive natural gamma signature is present in boring B-301 at a depth of 302 ft (92 m) (205 ft (62 m) msl). This interval is described as dark greenish gray, dense clayey sand grading to very dense silty sands in their bottom 25 ft (8 m). Boring B-401 encountered the Piney Point Formation at a depth of 278 ft (85 m) (-181 ft (-55 m) msl).

According to Hansen (Hansen, 1996), the top of the Piney Point Formation occurs at an approximate elevation of -200 ft (-61 m) msl in the CCNPP site area (Figure 2.5.1-13). The absence of late Eocene and early Miocene sediments indicate the absence of deposition or erosion for millions of years. A structure contour map of the top of the Piney Point Formation shows an erosion surface that dips gently toward the southeast (Figure 2.5.1 13).

The Chesapeake Group at the CCNPP site is divided into three marine formations which are, from oldest to youngest, the Calvert Formation, the Choptank Formation and the St. Marys Formation. These units are difficult to distinguish in the subsurface due to similar sediment types and are undivided at the CCNPP site (Glaser, 2003c). Achmad and Hansen (Achmad, 1997) indicate that the Chesapeake Group is approximately 245 ft (75 m) thick beneath the CCNPP site, based on boring CA-Ed 22 data. Kidwell (Kidwell, 1997) states that the stratigraphic relations within this group are highly complex. Based on cross sections presented in Kidwell (Kidwell, 1997), the contact between the St. Marys Formation and the underlying Choptank is estimated to be approximately 22 ft (7 m) deep in boring B-301 and at 10 ft (3 m) deep in B-401. The thickness of the Chesapeake Group (undifferentiated) is 280 ft in boring B-301 and 268 ft in B-401. The difference in these thicknesses and that in CA-Ed 22 is attributed to the geophysical log of the latter boring not continuing to the top of the boring and/or difference in the chosen top of the St. Marys Formation.

Although the formational contacts within the Chesapeake Group are difficult to impossible to identify, there are several strata which are encountered in most of the CCNPP Unit 3 investigation borings. The most persistent of these is the calcite-cemented sand shown in Figure 2.5.1-40 and probably is one of the units Kidwell (Kidwell, 1997) interprets as the Choptank Formation.

About 20 ft below the base of this cemented sand unit as a second, but much thinner cemented sand which is identified primarily by "N" values (the sum of the blow counts for the intervals 6 to 12 in (15 to 30 cm) and 12 to 18 in (30 to 46 cm) sample intervals in a standard SPT) higher than those immediately above and below.

The base of the Chesapeake Group (Piney Point Formation) is clearly identified in the geophysical log (Figure 2.5.1-35 and Figure 2.5.1-36) by the characteristic gamma curve response. Based on the boring log, this gamma curve response appears to be related to calcite-cemented sand.

The surficial deposits consist of two informal stratigraphic units: the Pliocene-age Upland deposits and Pleistocene to Holocene Lowland deposits. The Upland deposits consist of two units deposited in a fluviatile environment. The Upland deposits are areally more extensive in St. Mary's County than in Calvert County (Glaser, 1971). The outcrop distribution has a dendritic pattern and since it caps the higher interfluvial divides, this unit is interpreted as a highly dissected sediment sheet whose base slopes toward the southwest (Glaser, 1971) (Hansen, 1996). This erosion might have occurred due to differential uplift during the Pliocene or down cutting in response to lower base levels when sea level was lower during periods of Pleistocene glaciation.

2.5.1.2.3.4 Quarternary Formations

The Lowland deposits are considered to consist of three lithologic units. The basal unit is estimated to be 10 to 20 ft (3 to 6 m) thick and is often described as cobbly sand and gravel. This unit may represent high energy stream deposits in an alluvial environment near the base of eroding highlands to the west. The basal unit is overlain by as much as 90 ft (27 m) of bluish gray to dark brown clay that may be silty or sandy (Glaser, 1971) The uppermost of the three units consists of 10 to 30 ft (3 to 9 m) of pale gray, fairly well sorted, medium to coarse sand (Glaser, 1971). The Lowland deposits were laid down in fluvial to estuarine environments (Hansen, 1996) and are generally found along the Patuxent and Potomac River valleys and the Chesapeake Bay. These deposits occur in only a few places along the east shore of Chesapeake Bay.

Sands overlying the Chesapeake Group at the CCNPP site are mapped by Glaser (2003c) as Upland Deposits. Within the CCNPP Unit 3 power block these sands range in thickness from a feather edge in borings on the southern edge, to more than 50 ft in B-405.

Boring B-301 intersected 22 ft (7 m) of silty sand above the contact with the Chesapeake Group, while B-401 has 10 ft (3 m) of silty sand (Figure 2.5.1-35 and Figure 2.5.1-36). The sand in both borings grades into a coarser sand unit just above the contact. These sands are attributed to the Upland deposits previously mapped (Glaser, 2003c).

Terrace deposits in the CCNPP site area (Figure 2.5.1-30 and Figure 2.5.1-32) consist of interbedded light gray to gray silty sands and clay with occasional reddish brown pockets and are approximately 50 ft (15 m) thick. These units are Pliocene to Holocene in age.

Holocene deposits, mapped as Qal on the site Geologic Map, includes heterogeneous sediments underlying floodplains and beach sands composed of loose sand.}

2.5.1.2.4 Site Area Structural Geology

{The local structural geology of the CCNPP site described in this section is based primarily on a summary of published geologic mapping (Cleaves, 1968) (Glaser, 1994) (McCartan, 1995) (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c), aeromagnetic and gravity surveys (Hansen, 1978) (Hittelman, 1994) (Milici, 1995) (Bankey, 2002), detailed lithostratigraphic profiles along Calvert Cliffs (Kidwell, 1988) (Kidwell, 1997), results of earlier investigations performed at the CCNPP site (BGE, 1968) (CEG, 2005), as well as CCNPP site reconnaissance and subsurface exploration performed for this CCNPP Unit 3 study. Sparse geophysical and borehole data indicate that the basement likely consists of exotic crystalline magmatic arc material (Hansen, 1986) (Glover, 1995b). Although the basement beneath the site has not been penetrated with drill holes, regional geologic cross sections developed from geophysical, gravity and aeromagnetic, as well as limited deep borehole data from outside of the CCNPP site area, suggest that Precambrian and Paleozoic crystalline rocks and, less likely, Mesozoic rift-basin deposits are present at about 2,500 ft (762 m) msl (Section 2.5.1.2.2).

Tectonic models hypothesize that the crystalline basement underlying the site was accreted to a pre-Taconic North American margin in the Paleozoic along a suture that lies about 10 mi (16 km) west of the site (Figure 2.5.1-16 and Figure 2.5.1-22). The plate-scale suture is defined by a distinct north-northeast-trending magnetic anomaly that dips easterly between 35 and 45 degrees and lies about 7.5 to 9 mi (12 to 14.5 m) beneath the CCNPP site (Glover, 1995b) (Figure 2.5.1-16). Directly west of the suture lies the north to northeast-trending Taylorsville Basin and to the east, the postulated Queen Anne Mesozoic rift basin (Figure 2.5.1-9). These Mesozoic basins are delineated from geophysical data and a limited number of deep boreholes that penetrate the crust, and generally are considered approximately located where buried beneath the Coastal Plain (Jacobeen, 1972) (Hansen, 1986) (Benson, 1992) (LeTourneau, 2003). Most authors interpret Mesozoic basins directly west or east of the site; however, because the available geologic information used to constrain the basin locations is sparse, some depict the CCNPP site area to be underlain by a Mesozoic basin (Benson, 1992) (Figure 2.5.1-9). However, on the basis of a review of existing published geologic literature, site-specific data, and field reconnaissance, suggests there is no known basin-related fault or geologic evidence of basin-related faulting in the basement directly beneath the CCNPP site area.

Recent 1:24,000-scale mapping (Glaser, 2003b) (Glaser, 2003c) for Calvert County and St. Mary's County shows the stratigraphy at the CCNPP site area consisting of nearly flat-lying Cenozoic Coastal Plain sediments that have accumulated within the west-central part of the Salisbury Embayment (Figure 2.5.1-30 and Figure 2.5.1-31). The Salisbury Embayment is defined as a regional depocenter that has undergone slow crustal and regional downwarping as a result of sediment overburden during the Early Cretaceous and much of the Tertiary. The Coastal Plain deposits within this region of the Salisbury Embayment generally strike northeast-southwest and have a gentle dip to the southeast at angles close to or less than one to two degrees (Figure 2.5.1-30 and Figure 2.5.1-31). The gentle southerly dip of the sediments result in a surface outcrop pattern in which the strata become successively younger in a southeast direction across the embayment. The gentle-dipping to flat-lying Miocene Coastal Plain deposits are exposed in the steep cliffs along the western shoreline of Chesapeake Bay and provide excellent exposures to assess the presence or absence of tectonic-related structures.

Local geologic cross sections of the site area depict unfaulted, southeast-dipping Eocene-Miocene Coastal Plain sediments in an unconformable contact with overlying Pliocene Upland deposits (Glaser, 1994) (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c) (Figure 2.5.1 12, Figure 2.5.1-30, and Figure 2.5.1-31). No faults or folds are depicted on these geologic cross sections. A review of an Early Site Review report (BGE, 1977), i.e. Perryman site, and a review of the Preliminary Safety Analysis Report for the Douglas Point site (Potomac Electric Power Company, 1973), located along the eastern shore of the Potomac River about 45 mi (72 km) west-southwest of the CCNPP site, also reported no faults or folds within a 5 mi (8 km) radius of the CCNPP site. The Updated Final Safety Analysis Report for the Hope Creek site, located in New Jersey along the northern shore of Delaware Bay, also was reviewed for tectonic features previously identified within 5 mi (8 km) of the CCNPP site, yet none were identified (PSEG, 2002). Review of a seismic source characterization study (URS, 2000) for a liquidefied natural gas plant at Cove Point, about 3 mi (5 km) southeast of the site, also identified no faults or folds projecting toward or underlying the CCNPP site area.

On the basis of literature review, and aerial and field reconnaissance, the only potential structural features at and within the CCNPP site area consist of a hypothetical buried northeast-trending fault (Hansen, 1986), two inferred east-facing monoclines developed within Mesozoic and Tertiary deposits along the western shore of Chesapeake Bay (McCartan, 1995), and multiple subtle folds or inflections in Miocene strata and a postulated fault directly south of the site (Kidwell, 1997) (Figure 2.5.1-24). The Hillville fault of Hansen and Edwards (Hansen, 1986) and inferred fold of McCartan (McCartan, 1995) and Kidwell (Kidwell, 1997) are described in Sections 2.5.1.1.4.4.4 and Section 2.5.3. As previously discussed in Section 2.5.1.1.4.4.4, none of these features are considered capable tectonic sources, as defined in RG 1.165, Appendix A. Each of these features is discussed briefly below. Only the Hillville fault has been mapped within or directly at the 5 mi (8 km) radius of the CCNPP site area (Figure 2.5.1-26a, Figure 2.5.1-26b, and Figure 2.5.1-30).

Hillville fault of Hansen and Edwards (Hansen, 1986): The 26 mile long Hillville fault approaches to within 5 mi (8 km) of the CCNPP site (Figure 2.5.1-30). The fault consists of a northeast-striking zone of steep southeast-dipping reverse faults that coincide with the Sussex-Currioman Bay aeromagnetic anomaly. The style and location of faulting are based on seismic reflection data collected about 9 mi (14 km) west-southwest of the site. A seismic line imaged a narrow zone of discontinuities that vertically separate basement by as much as 250 ft (76 m) (Hansen, 1978). Hansen and Edwards (Hansen, 1986) interpret this offset as part of a larger lithotectonic terrane boundary that separates basement rocks associated with Triassic rift basins on the west and low-grade metamorphic basement on the east. The Hillville fault may represent a Paleozoic suture zone that was reactivated in the Mesozoic and Early Tertiary. Based on stratigraphic correlation between boreholes within Tertiary Coastal Plain deposits, Hansen and Edwards (Hansen, 1986) speculate that the Hillville fault was last active in the Early Paleocene. There is no pre-EPRI and post-EPRI (1986) seismicity spatially associated with this feature (Figure 2.5.1-24) nor is there any geomorphic evidence of Quaternary deformation. The Hillville fault is not considered a capable tectonic source.

In addition, two speculative and poorly constrained east-facing monoclines along the western margin of Chesapeake Bay are mapped within the 5 mi (8 km) radius of the CCNPP site area. East-facing monoclines (McCartan, 1995): The unnamed monoclines are not depicted on any geologic maps of the area, including those by the authors, but they are shown on geologic cross sections that trend northwest-southeast across the existing site and south of the CCNPP site near the Patuxent River (McCartan, 1995) (Figure 2.5.1-24). East-facing monoclines are inferred beneath Chesapeake Bay at about 2 and 10 mi (3.2 to 16 km) east and southeast, respectively, from the CCNPP site. Along a northerly trench, the two monoclines delineate a continuous north-trending, east-facing monocline. As mapped in cross section and inferred in plan view, the monoclines trend approximately north along the western shore of Chesapeake Bay. The monoclines exhibit a west-side up sense of structural relief that projects into the Miocene Choptank Formation (McCartan, 1995). The overlying Late Miocene St. Marys Formation is not shown as warped. Although no published geologic data are available to substantiate the existence of the monoclines, McCartan (McCartan, 1995) believes the distinct elevation change across Chesapeake Bay and the apparent linear nature of Calvert Cliffs are tectonically controlled. CCNPP site and aerial reconnaissance, coupled with literature review, for the CCNPP Unit 3 study strongly support a non-tectonic origin for the physiographic differences across the Chesapeake Bay (Section 2.5.1.1.4.4.4). There is no pre-EPRI or post-EPRI (1986) seismicity spatially associated with this feature, nor is there geologic data to suggest that the monocline proposed by McCartan (McCartan, 1995) is a capable tectonic source.

Multiple subtle folds or inflections developed in Miocene Coastal Plain strata including a postulated fault are mapped in the cliff exposures along the west side of Chesapeake Bay. Kidwell's (Kidwell, 1997) postulated folds and fault: Kidwell (Kidwell, 1988) (Kidwell, 1997) prepared over 300 lithostratigraphic columns along a 25 mi (40 km) long stretch of Calvert Cliffs that intersect much of the CCNPP site (Figure 2.5.1-28). When these stratigraphic columns are compiled into a cross section, they collectively provide a 25 mi (40 km) long nearly continuous exposure of Miocene, Pliocene and Quaternary deposits. Kidwell's (Kidwell, 1997) stratigraphic analysis indicates that the Miocene Coastal Plain deposits strike northeast and dip very shallow between 1 and 2 degrees to the south-southeast, which is consistent with the findings of others (McCartan, 1995) (Glaser 2003b) (Glaser, 2003c). The regional southeast-dipping strata are disrupted occasionally by several low amplitude broad undulations developed within Miocene Coastal Plain deposits (Figure 2.5.1-28). The stratigraphic undulations are interpreted as monoclines and asymmetrical anticlines by Kidwell (Kidwell, 1997). In general, the undulatory stratigraphic contacts coincide with basal unconformities having wavelengths of 2.5 to 5 mi (4 to 8 km) and amplitudes of 10 to 11 ft (approximately 3 meters). Based on prominent stratigraphic truncations, the inferred warping decreases upsection into the overlying upper Miocene St. Marys Formation. Any inferred folding of the overlying Pliocene and Quaternary fluvial deposits is poorly constrained and can be readily explained by highly variable undulating unconformities.

Near Moran Landing, about 1.2 mi (1.9 km) south of the site, Kidwell (Kidwell, 1997) interprets an apparent 6 to 10 ft (2 to 3 m) elevation change in Miocene strata, and a 3 to 12 (0.9 to 3.7 m) ft elevation change in Pliocene and Quaternary (?) fluvial material (Figure 2.5.1-24 and Figure 2.5.1-28). Kidwell (Kidwell, 1997) infers the presence of a fault to explain the difference in elevation of strata across Moran Landing. The postulated fault is not shown on the Kidwell (Kidwell, 1997) section, or any published geologic map, however the inferred location is approximately 1.2 mi (1.9 m) south of the CCNPP site. The hypothesized fault is not exposed in the cliff face, but Kidwell (Kidwell, 1997) postulates the presence of a fault, and is based entirely on a change in elevation and bedding dip of Miocene stratigraphic boundaries projected across the fluvial valley of Moran Landing. Kidwell (Kidwell, 1997) postulates that the fault strikes northeast and exhibits a north-side down sense of separation across all the geologic units (Miocene through Quaternary). With regard to the apparent elevation changes for the Pliocene and Quaternary unconformities, these can be readily explained by channeling and highly irregular erosional surfaces. Field and aerial reconnaissance, coupled with interpretation of aerial photography and LiDAR data (Section 2.5.3.1 for additional information regarding the general methodology) conducted as part of this CCNPP Unit 3 study revealed no features suggestive of tectonic deformation developed in the surrounding Pliocene and Quaternary surfaces.

There is no pre-EPRI or post-EPRI study (EPRI, 1986) seismicity spatially associated with the Kidwell (Kidwell, 1997) features, the hypothetical features are not aligned or associated with gravity and magnetic anomalies, nor is there data to indicate that the features proposed by Kidwell (Kidwell, 1997) are capable tectonic sources.

The most detailed subsurface exploration of the site was performed by Dames & Moore as part of the original PSAR (BGE, 1968) for the existing CCNPP foundation and supporting structures. The PSAR study included drilling as many as 85 geotechnical boreholes, collecting downhole geophysical data, and acquiring seismic refraction data across the site. Dames and Moore (BGE, 1968) developed geologic cross sections extending from Highway 2/4 northwest of the site to Camp Conoy on the southeast which provide valuable subsurface information on the lateral continuity of Miocene Coastal Plain sediments and Pliocene Upland deposits (Figure 2.5.1-30 and Figure 2.5.1-32). Cross sections C-C' and D-D' pre-date site development and intersect the existing and proposed CCNPP site for structures trending north-northeast, parallel to the regional structural grain. These sections depict a nearly flat-lying, undeformed geologic contact between the Middle Miocene Piney Point Formation and the overlying Middle Miocene Calvert Formation at about –200 ft (-61 m) msl (Figure 2.5.1-39 and Figure 2.5.1-40).

Geologic sections developed from geotechnical borehole data collected as part of the CCNPP Unit 3 study also provide additional detailed sedimentological and structural relations for the upper approximately 400 ft (122 m) of strata directly beneath the footprint of the site. Similar to the previous cross sections prepared for the site, new geologic borehole data support the interpretation of flat-lying and unfaulted Miocene and Pliocene stratigraphy at the CCNPP site (Figure 2.5.1-37 and Figure 2.5.1-41). A cross section prepared oblique to previously mapped northeast-trending structures (i.e., Hillville fault), inferred folds (McCartan, 1995) (Kidwell, 1997), and the fault of Kidwell (Kidwell, 1997) shows nearly flat-lying Miocene and Pliocene stratigraphy directly below the CCNPP site. Multiple key stratigraphic markers provide evidence for the absence of Miocene-Pliocene faulting and folding beneath the site. Minor perturbations are present across the Miocene-Pliocene stratigraphic boundary, as well as other Miocene-related boundaries, however these minor elevation changes are most likely related to the irregular nature of the fluvial unconformities and are not tectonic-related.

Numerous investigations of the Calvert Cliffs coastline over many decades by government researchers, stratigraphers, and by consultants for Baltimore Gas and Electric, as well as investigations for the CCNPP Unit 3, have reported no visible signs of tectonic deformation within the exposed Miocene deposits near the site, with the only exception being that of Kidwell (Kidwell, 1997) (Figure 2.5.1-42). Collectively, the majority of published and unpublished geologic cross sections compiled for much of the site area and site, coupled with regional sections (Achmad, 1997) (Glaser, 2003b) (Glaser, 2003c) and site and aerial reconnaissance, indicate the absence of Pliocene and younger faulting and folding. A review and interpretation of aerial photography, digital elevation models, and LiDAR data of the CCNPP site area, coupled with aerial reconnaissance, identified few discontinuous north to northeast-striking lineaments. None of these lineaments were interpreted as fault-related, nor coincident with the Hillville fault or the other previously inferred Miocene-Pliocene structures mapped by McCartan (McCartan, 1995) and Kidwell (Kidwell, 1997) (Section 2.5.3). A review of regional geologic sections and interpretation of LiDAR data suggest that the features postulated by Kidwell (Kidwell, 1997), if present, are not moderate or prominent structures, and do not deform Pliocene and Quaternary strata. In summary, on the basis of regional and site geologic and geomorphic data, there are no known faults within the site area, with the exception of the poorly constrained Hillville fault that lies along the northwestern perimeter of the 5 mi (8 km) radius of the site (Hansen, 1986).}

2.5.1.2.5 Site Area Geologic Hazard Evaluation

{No geologic hazards have been identified within the CCNPP site area. No geologic units at the site are subject to dissolution. No deformation zones were encountered in the exploration or excavation for CCNPP Units 1 and 2 and none have been encountered in the site investigation for CCNPP Unit 3. Because the CCNPP Unit 3 plant site is located at an elevation of approximately 85 ft (26 m) msl and approximately 1,000 ft (305 m) from the Chesapeake Bay shoreline, it is unlikely that shoreline erosion or flooding will impact the CCNPP site.}

2.5.1.2.6 Site Engineering Geology Evaluation

2.5.1.2.6.1 Engineering Soil Properties and Behavior of Foundation Materials

Engineering soil properties, including index properties, static and dynamic strength, and compressibility are discussed in Section 2.5.4. Variability and distribution of properties for the foundation bearing soils will be evaluated and mapped as the excavation is completed.

Settlement monitoring will based on analyses performed for the final design.

2.5.1.2.6.2 Zones of Alteration, Weathering, and Structural Weakness

{No unusual weathering profiles have been encountered during the site investigation. No dissolution is expected to affect foundations. Any noted desiccation, weathering zones, joints or fractures will be mapped during excavation and evaluated.}

2.5.1.2.6.3 Deformational Zones

{No deformation zones were encountered in the exploration or excavation for CCNPP Units 1 and 2 and none have been encountered in the site investigation for CCNPP Unit 3. Excavation mapping is required during construction and any noted deformational zones will be evaluated. No capable tectonic sources as defined by Regulatory Guide 1.165 (NRC, 1997) exist in the CCNPP site region.}

2.5.1.2.6.4 Prior Earthquake Effects

{Outcrops are rare within the CCNPP site area. Studies of the CCNPP Unit 1 and 2 excavation, available outcrops, and extensive exposures along the western shore of Chesapeake Bay have not indicated any evidence for earthquake activity that affected the Miocene deposits. There is no evidence of earthquake-induced liquefaction in the State of Maryland (Crone, 2000) (Wheeler, 2005).}

2.5.1.2.6.5 Effects of Human Activities

{No mining operations, excessive extraction or injection of groundwater or impoundment of water has occurred within the site area that can affect geologic conditions.}

2.5.1.2.6.6 Site Groundwater Conditions

A detailed discussion of groundwater conditions is provided in Section 2.4.12.

2.5.1.3 References

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Table 2.5.1-1 Definitions of Classes Used in the Compilation of Quaternary Faults, Liquefaction Features, and Deformation in the Central and Eastern United States (Page 1 of 1)

Class Category	Definition
Class A	Geologic evidence demonstrates the existence of a Quaternary fault of tectonic origin, whether the fault is exposed for mapping or inferred from liquefaction to other deformational features.
Class B	Geologic evidence demonstrates the existence of a fault or suggests Quaternary deformation, but either (1) the fault might not extend deeply enough to be a potential source of significant earthquakes, or (2) the currently available geologic evidence is too strong to confidently assign the feature to Class C but not strong enough to assign it to Class A.
Class C	Geologic evidence is insufficient to demonstrate (1) the existence of tectonic fault, or (2) Quaternary slip or deformation associated with the feature.
Class D	Geologic evidence demonstrates that the feature is not a tectonic fault or feature; this category includes features such as demonstrated joints or joint zones, landslides, erosional or fluvial scarps, or landforms resembling fault scarps, but of demonstrable non-tectonic origin.

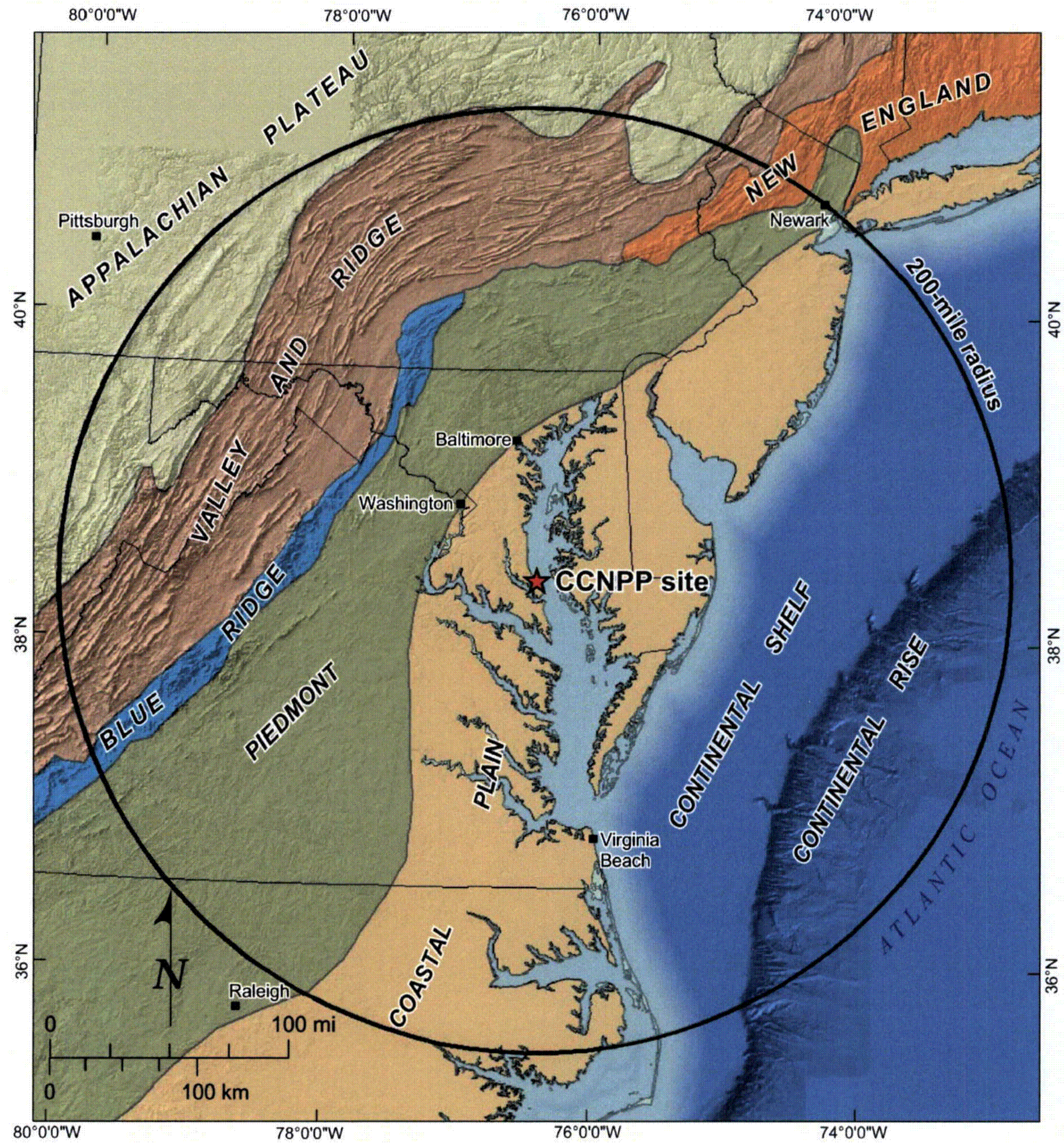
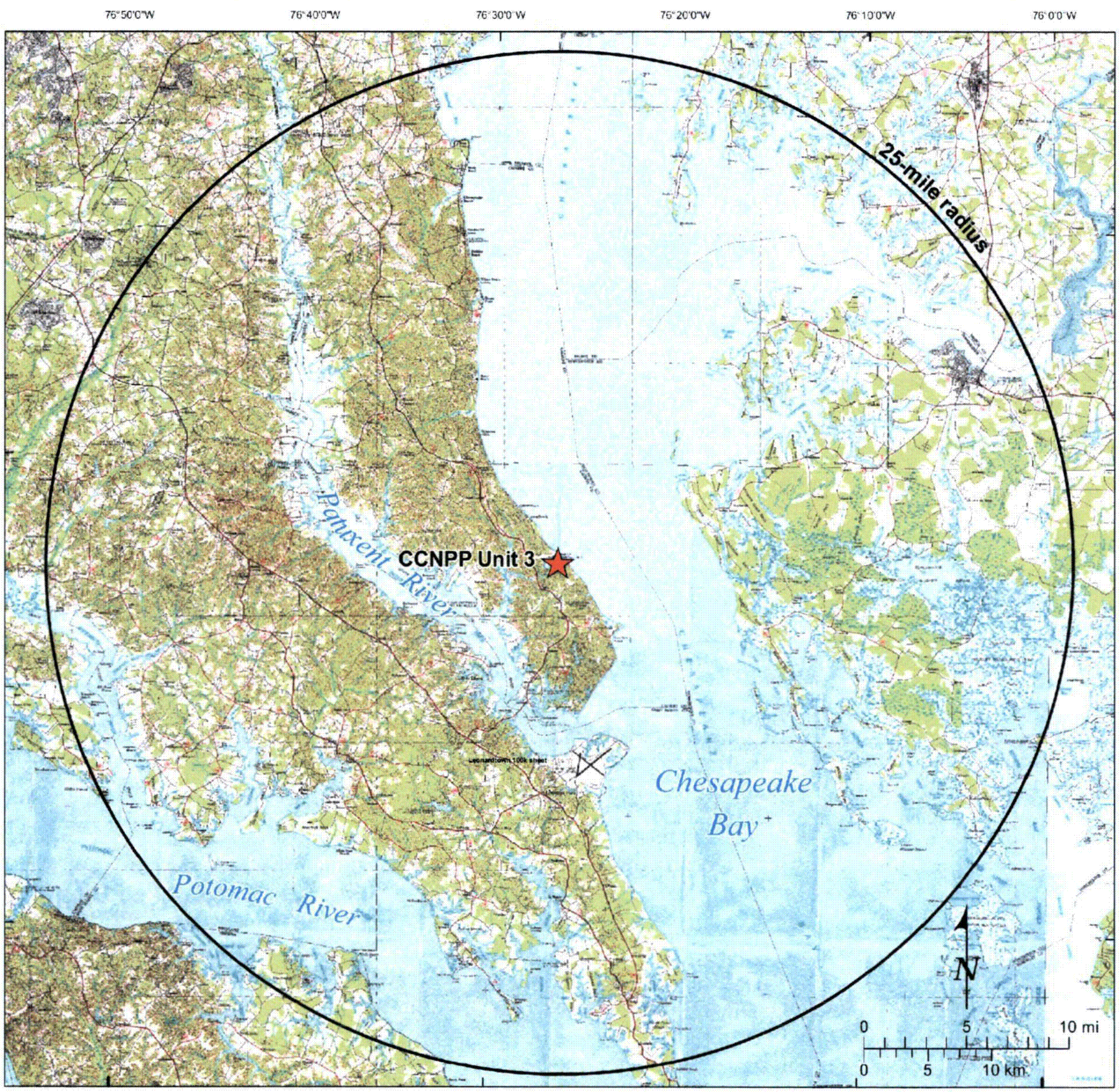


FIGURE 2.5.1-1 Rev. 0

MAP OF PHYSIOGRAPHIC PROVINCE (FENNEMAN AND JOHNSON, 1946)

CCNPP UNIT 3 FSAR



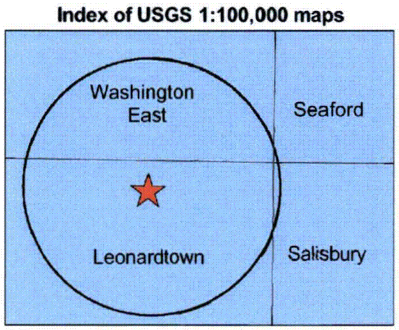
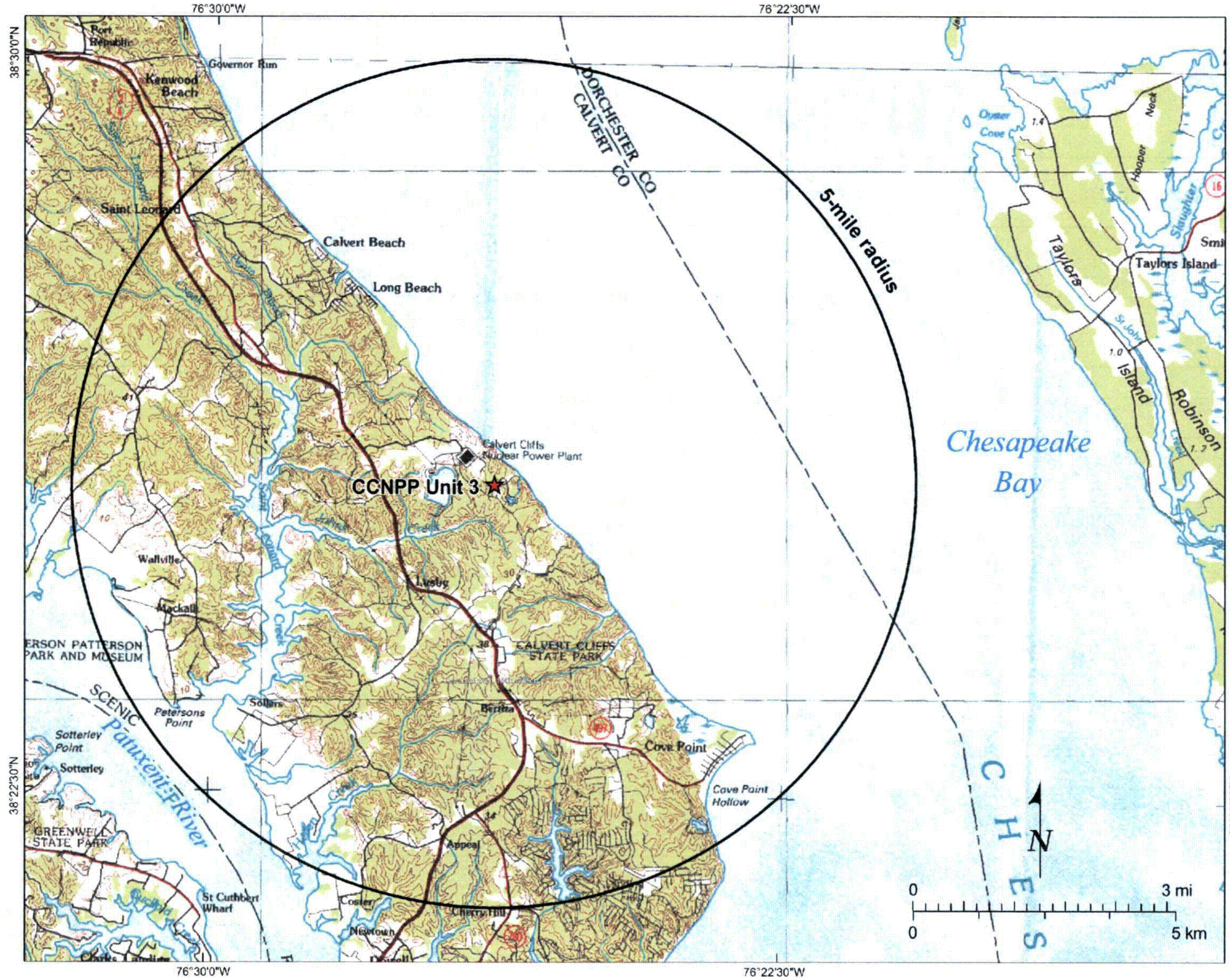


FIGURE 2.5.1-2 **Rev. 0**

SITE VICINITY TOPOGRAPHIC MAP
25-MILE (40-Km) RADIUS

CCNPP UNIT 3 FSAR

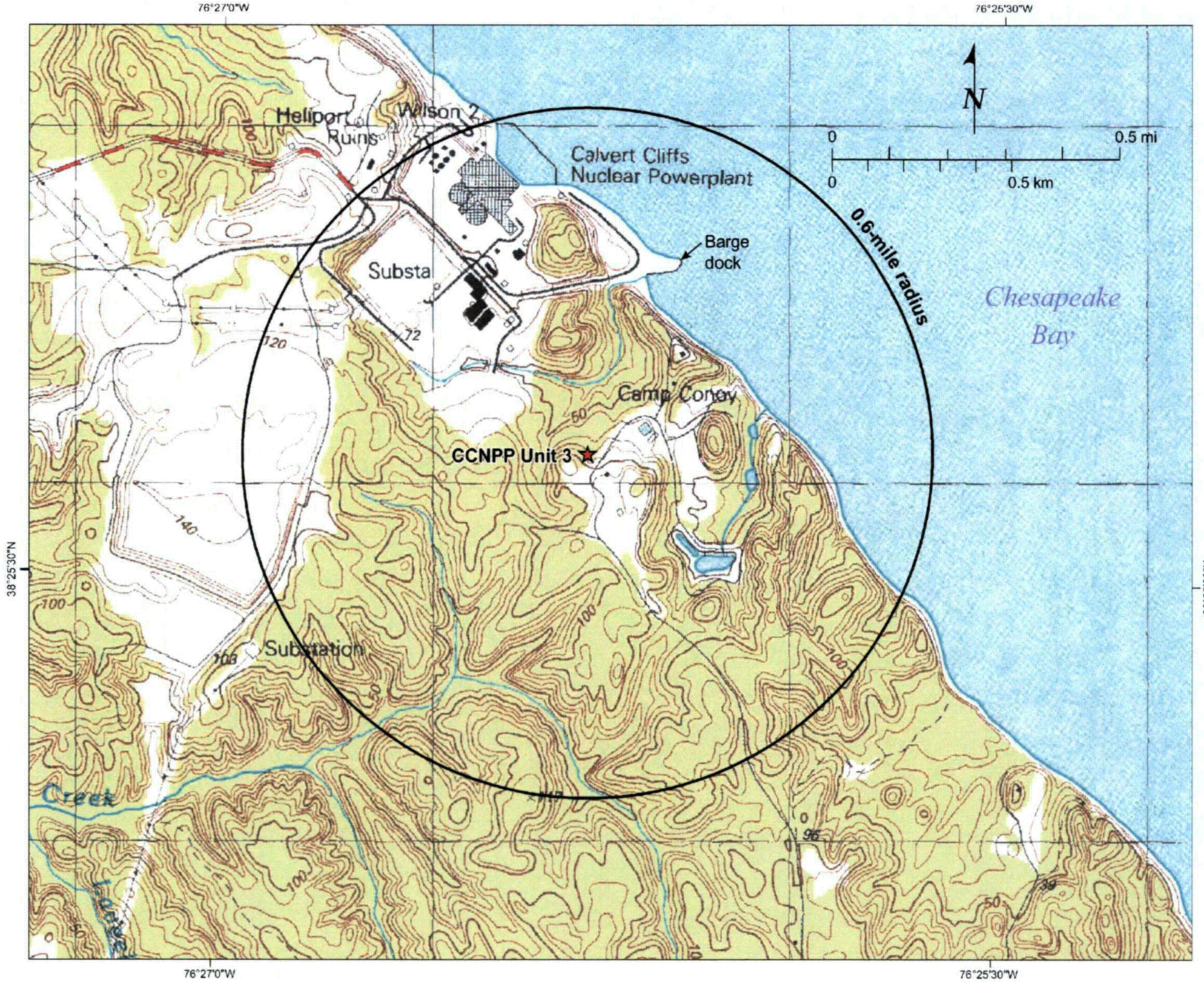


Base map: Leonardtown 30' x 60' U.S. Geological Survey Topographic Map

FIGURE 2.5.1-3 **Rev. 0**

SITE AREA TOPOGRAPHIC MAP
5-MILE (8-Km) RADIUS

CCNPP UNIT 3 FSAR



Base map: 1987 USGS 7.5-minute Cove Point quadrangle
North American datum 1983, 10-foot contour interval

FIGURE 2.5.1-4	**Rev. 0**

SITE TOPOGRAPHIC MAP
0.6–MILE (1–Km) RADIUS

CCNPP UNIT 3 FSAR

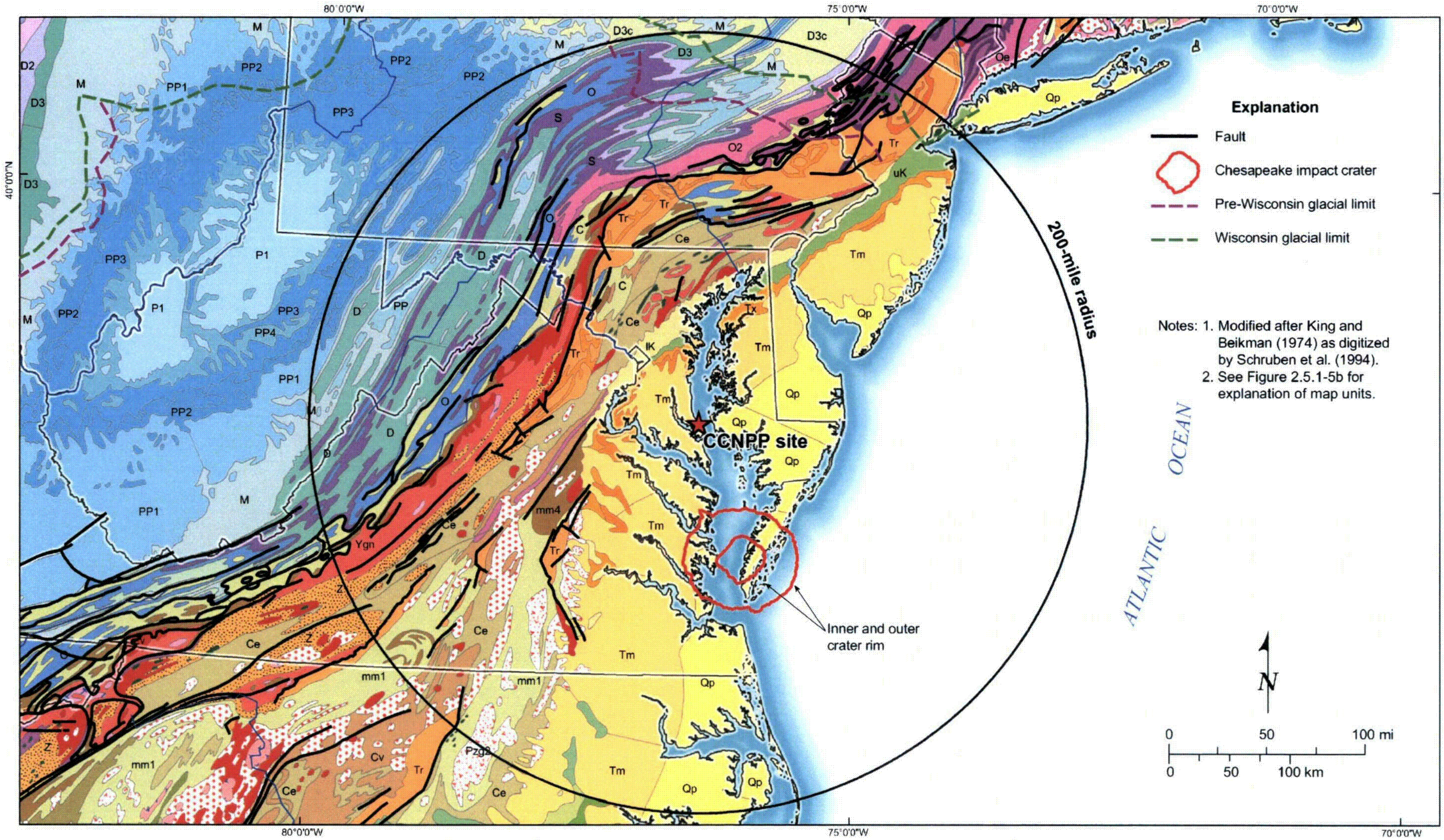


FIGURE 2.5.1-5a **Rev. 0**

REGIONAL GEOLOGIC MAP
200-MILE (320-Km) RADIUS

CCNPP UNIT 3 FSAR

Explanation

CCNPP 200-mile Geology
King and Beikman (1974) Unit Descriptions

Unit	Description
Qp	Pleistocene
Tm	Miocene
Te	Eocene
Tx	Paleocene
uK4	Navarro Group
uK	Upper Cretaceous
uK1	Woodbine and Tuscaloosa Groups
IK3	Washita Group
IK	Lower Cretaceous
Trv	Mafic Lava interbedded in Triassic Newark Group
Tri	Triassic mafic intrusives
Tr	Triassic
um	Ultramafic rocks
P1	Wolfcampian series
cat	Catacalastic rocks
PP4	Virgilian Series
Pzg3	Upper Paleozoic granitic rocks
PP3	Missourian series
PP2	Des Moinesian series
PP	Pennsylvanian
PP1	Atokan and Morrowan series
mm1	Felsic paragneiss and schist
mm2	Mafic paragneiss (= hornblendite, amphibolite)
mm3	Migmatite
Pzmi	Paleozoic mafic intrusives
mm4	Felsic orthogneiss (= granite gneiss)
M	Mississippian

Unit	Description
D3	Upper Devonian
D3c	Upper Devonian continental
Pzg2	Middle Paleozoic granitic rocks
D2	Middle Devonian
D	Devonian
De	Devonian eugeosynclinal
D1	Lower Devonian
DS	Devonian and Silurian
S3	Upper Silurian (Cayugan)
S	Silurian
lPz	Lower Paleozoic
O2	Middle Ordovician (Mohawkian)
O	Ordovician
Oe	Ordovician eugeosynclinal
Ov	Ordovician volcanic rocks
Pzg1	Lower Paleozoic granitic rocks
O1	Lower Ordovician (Canadian)
OC	Lower Ordovician and Cambrian carbonate rocks
C	Cambrian
Ce	Cambrian eugeosynclinal
Cv	Cambrian volcanics
Cq	Basal Lower Cambrian clastic rocks
Z	Z sedimentary rocks
Zg	Z granitic rocks
Zv	Z volcanic rocks
Ym	Paragneiss and schist
Ya	Anorthosite
Ygn	Orthogneiss

FIGURE 2.5.1-5b **Rev. 0**

REGIONAL GEOLOGIC MAP
200–MILE (320–Km) RADUIS
EXPLANATION

CCNPP UNIT 3 FSAR

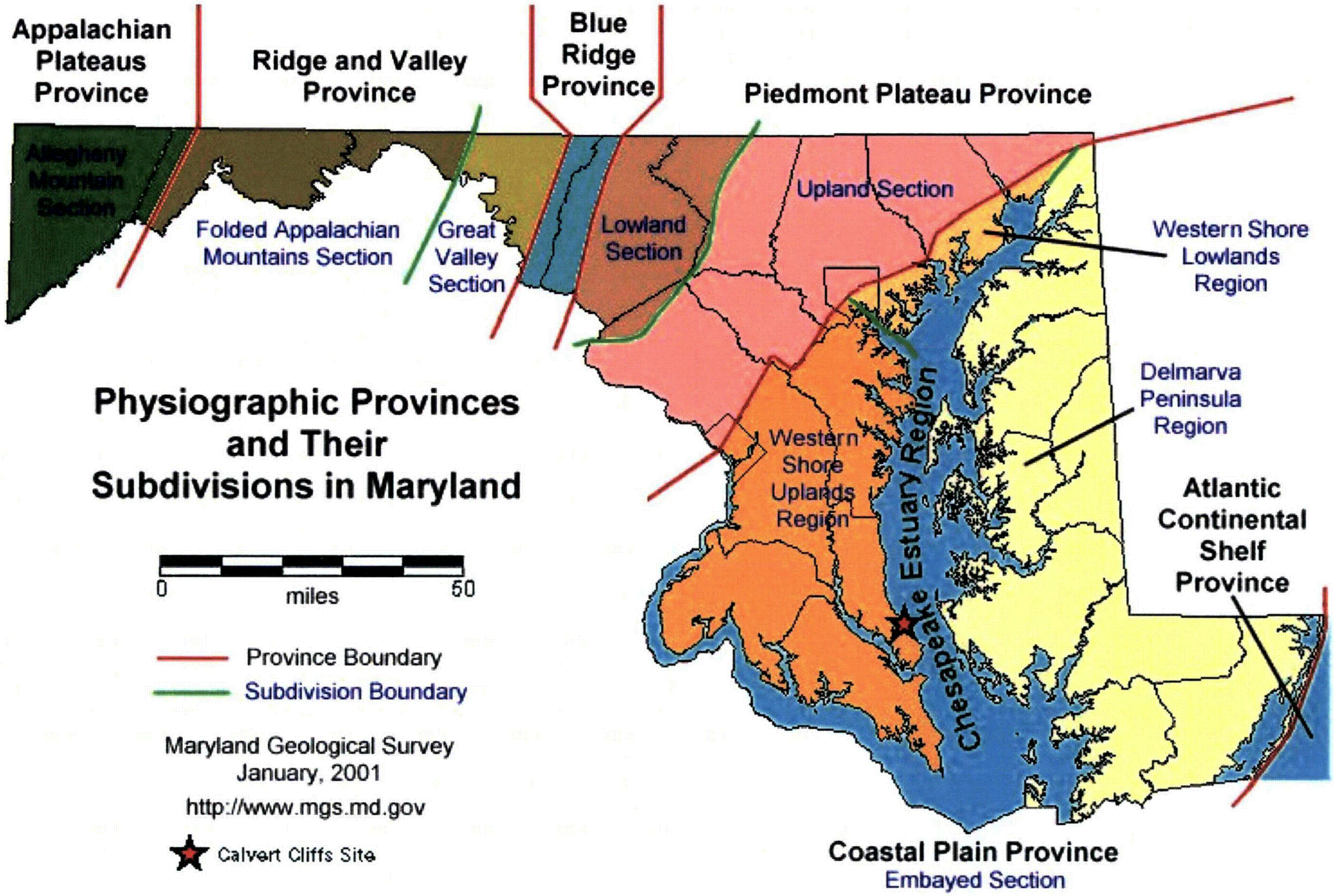


FIGURE 2.5.1-6 **Rev. 0**

PHYSIOGRAPHIC MAP OF MARYLAND

CCNPP UNIT 3 FSAR

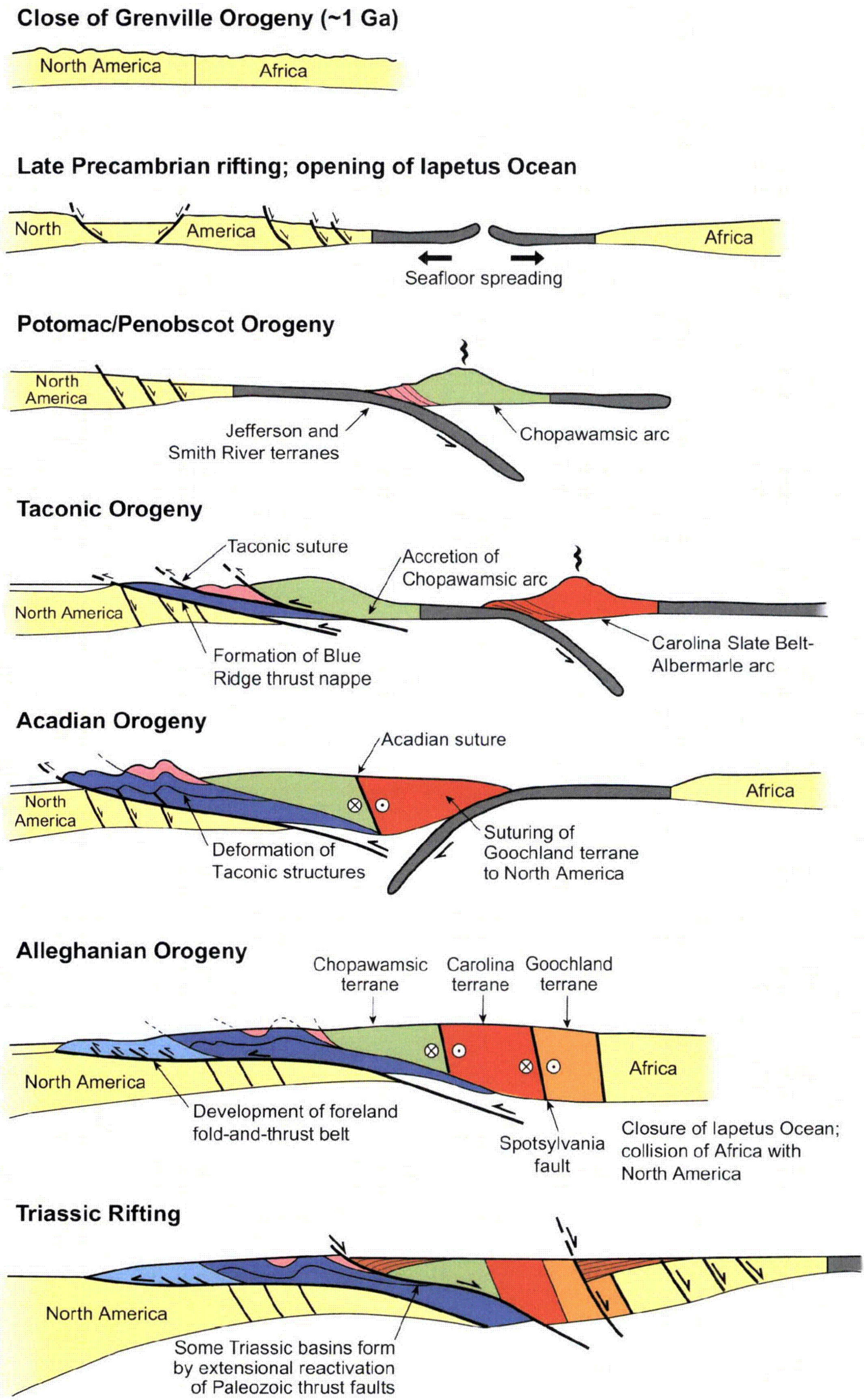
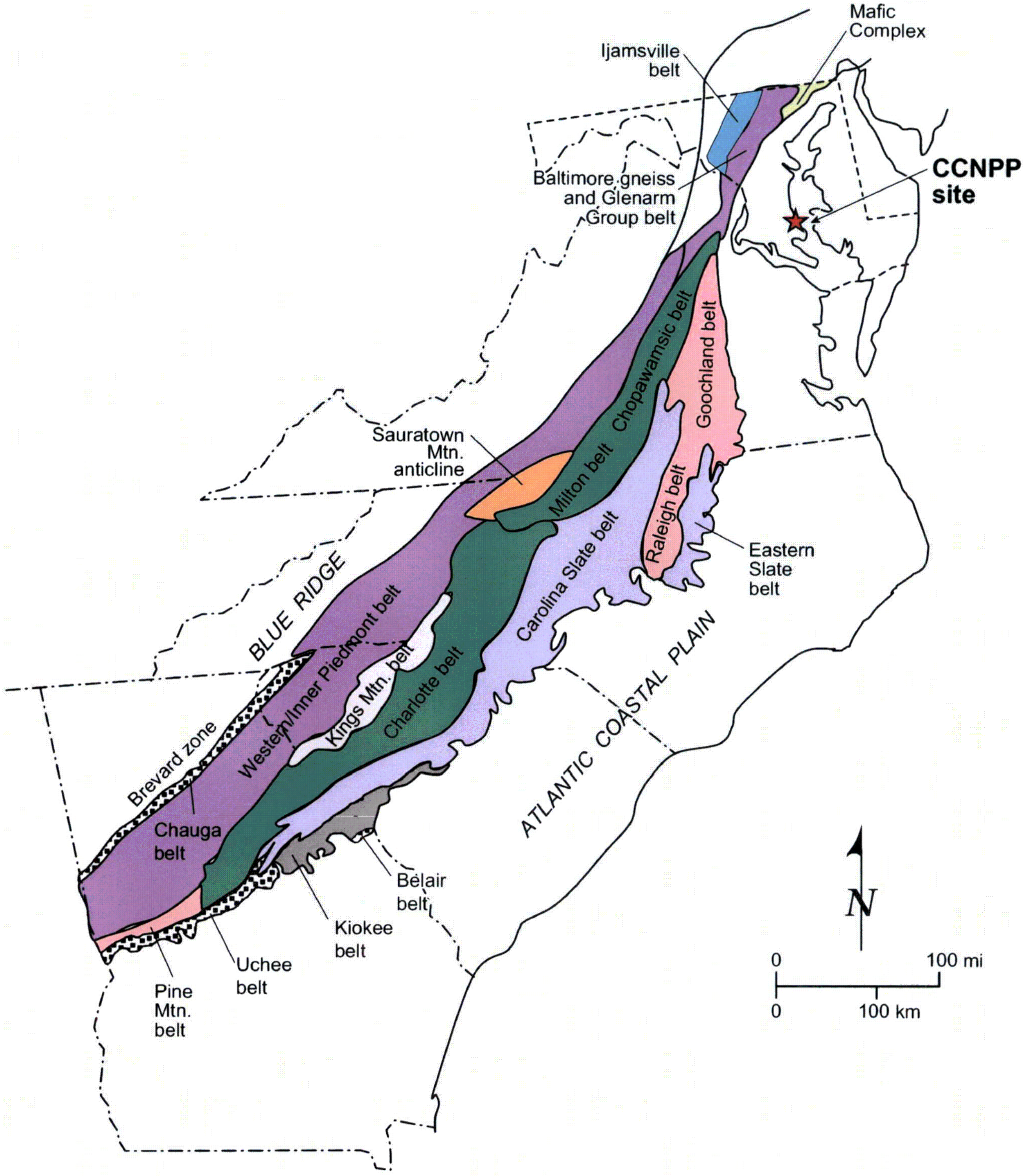


FIGURE 2.5.1-7 Rev. 0

EVOLUTION OF THE APPALACHIAN OROGEN (AFTER HATCHER, 1987)

CCNPP UNIT 3 FSAR



Source: Bledsoe and Marine, 1980; and Fichter and Baedke, 2000

FIGURE 2.5.1-8 Rev. 0

LITHOTECTONIC BELTS OF THE PIEDMONT PROVINCE

CCNPP UNIT 3 FSAR

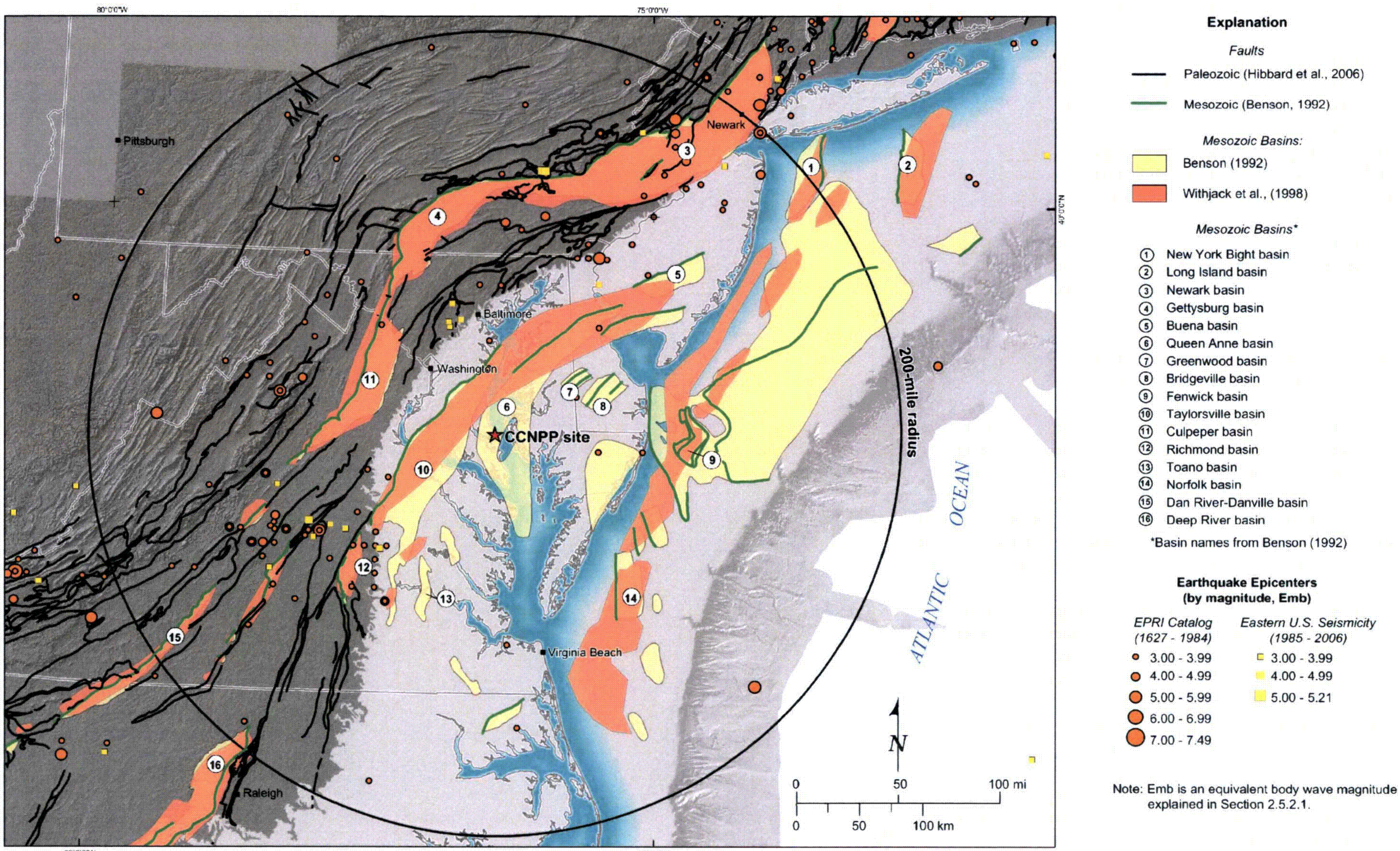


FIGURE 2.5.1-9 Rev. 0

MAP OF MESOZOIC BASINS BY VARIOUS AUTHORS

CCNPP UNIT 3 FSAR

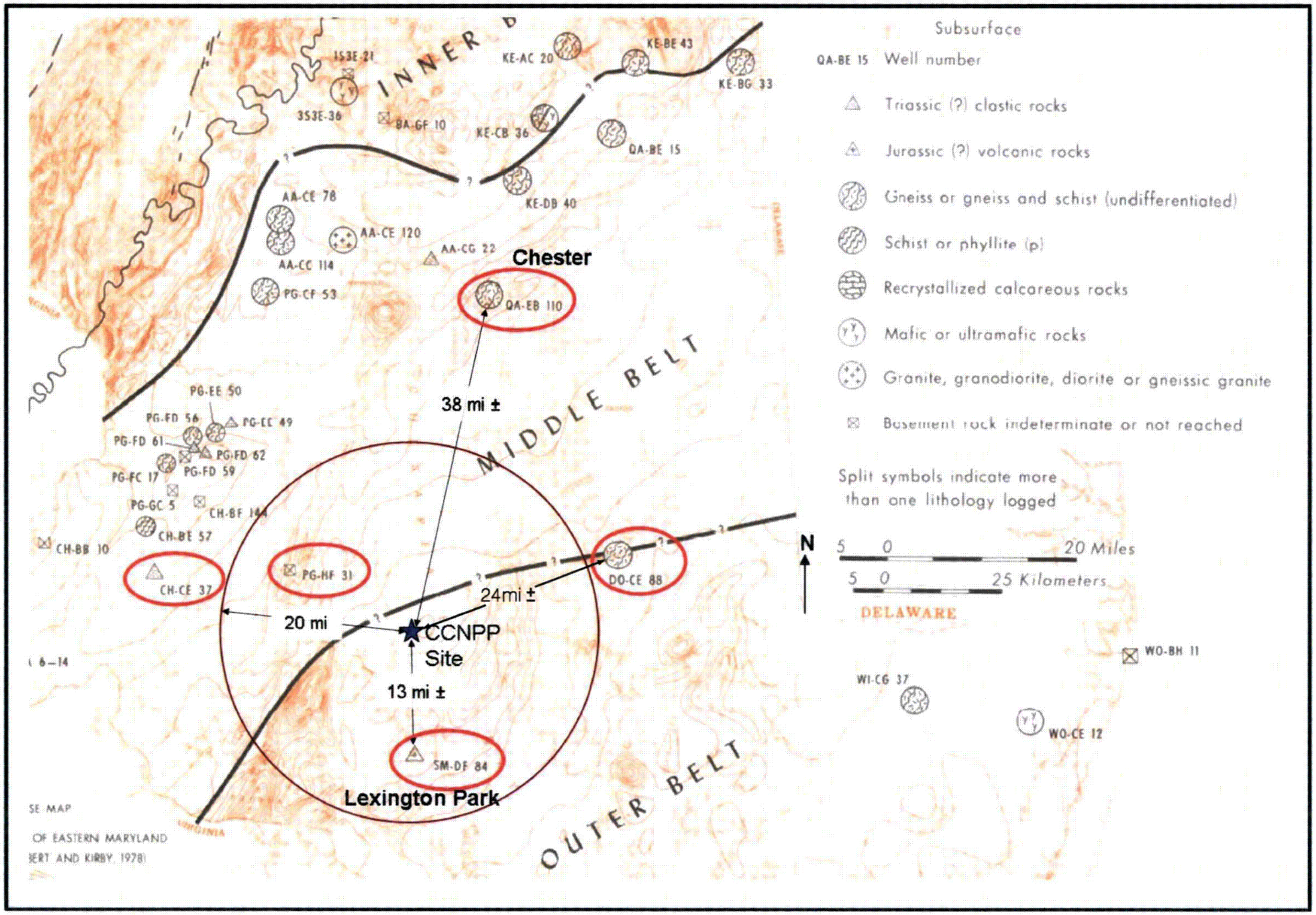
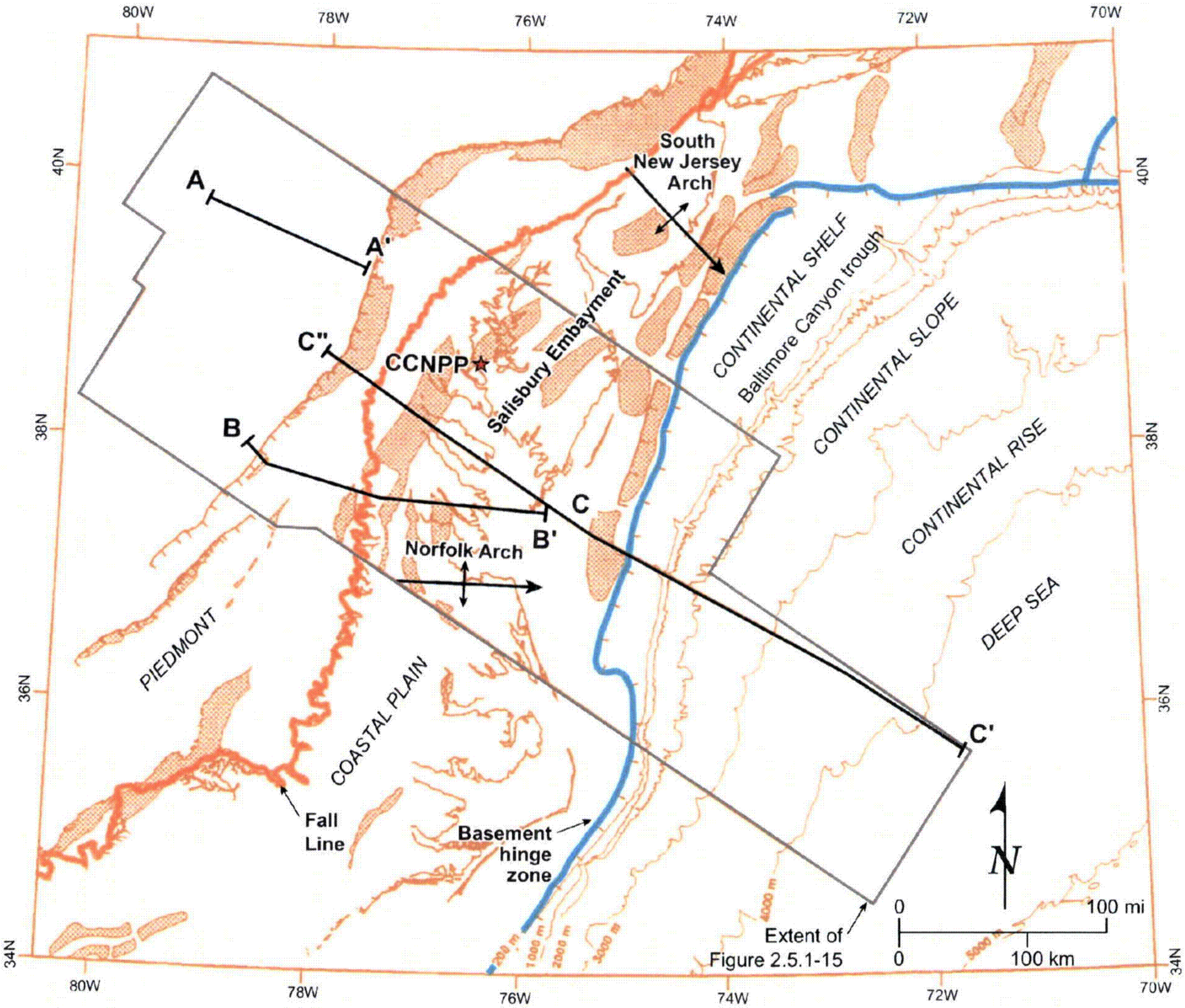


FIGURE 2.5.1-10 **Rev. 0**

LITHOLOGIES OF BASEMENT ROCKS FROM COASTAL PLAIN WELLS

CCNPP UNIT 3 FSAR



Modified from Klitgord et al. (1995)

Explanation

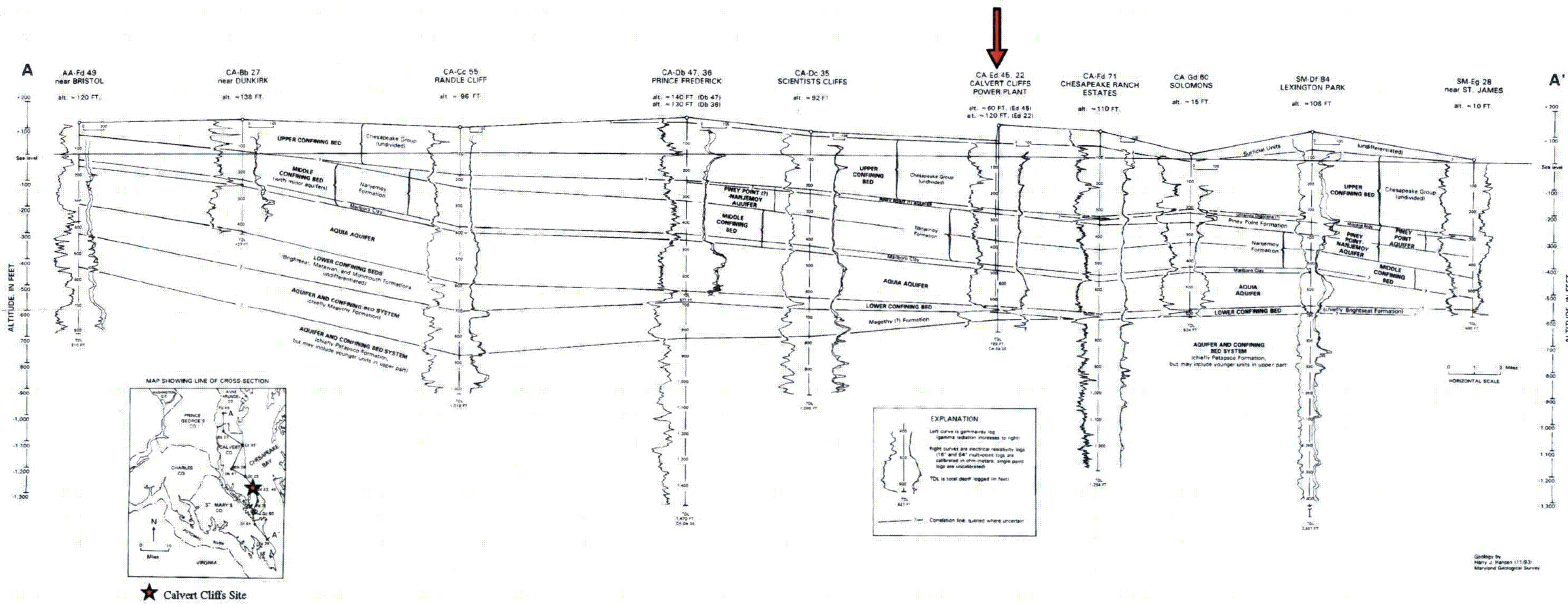
A A' Cross section (see Figures 2.5.1-16, 2.5.1-17, and 2.5.1-18)

Mesozoic rift basin

FIGURE 2.5.1-11 Rev. 0

TECTONIC FEATURES OF THE MID-ATLANTIC PASSIVE MARGIN

CCNPP UNIT 3 FSAR

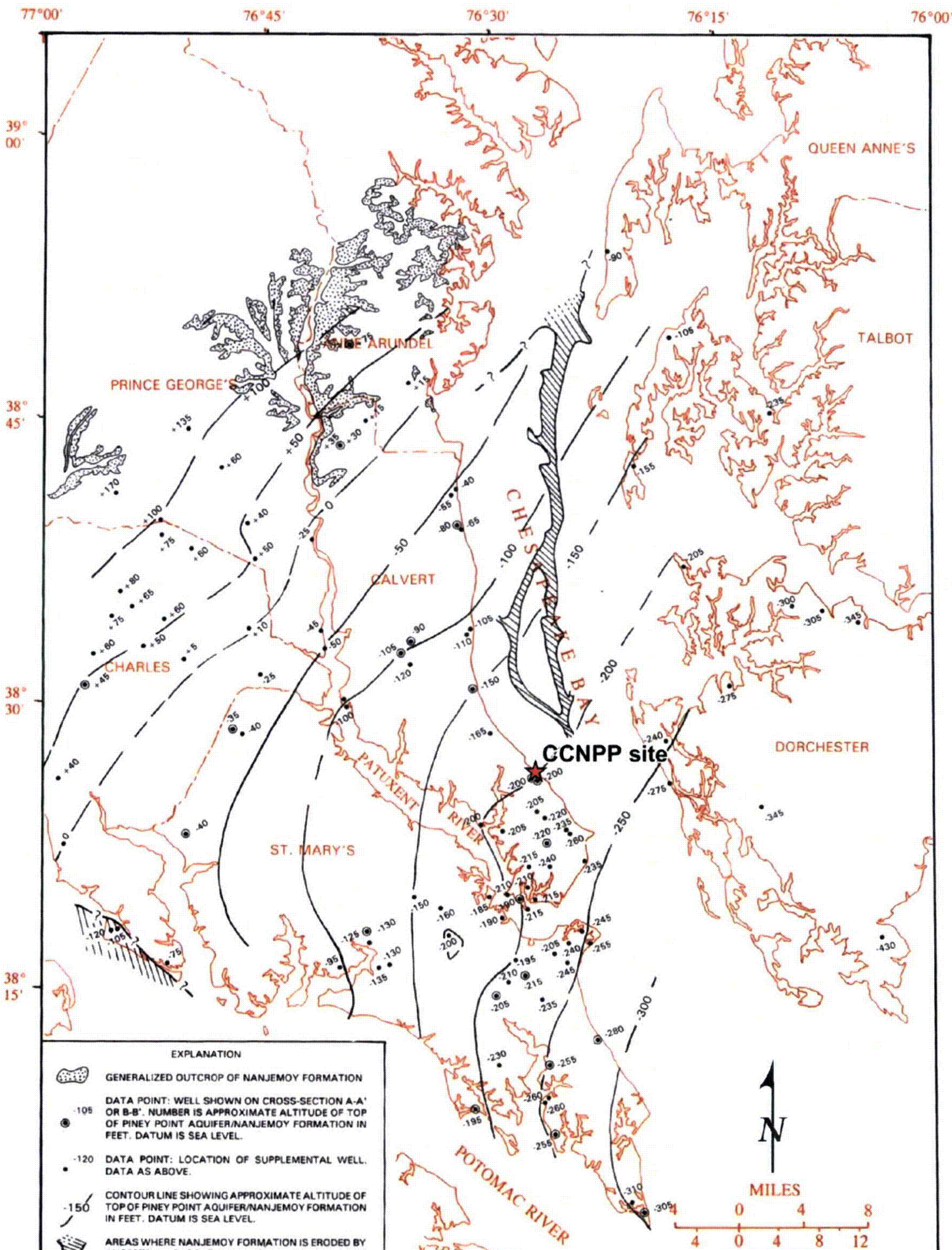


THE CCNPP SITE IS REPRESENTED
BY THE WELLS CA-Ed 45 AND 22.

FIGURE 2.5.1-12 **Rev. 0**

STRATIGRAPHIC CROSS-SECTION
THROUGH ANNE ARUNDEL,
CALVERT AND ST. MARY'S COUNTIES

CCNPP UNIT 3 FSAR



BASE FROM MARYLAND GEOLOGICAL SURVEY, 1961, 1:250,000

Modified from Achmad and Hansen (1997)

FIGURE 2.5.1-13 Rev. 0

STRUCTURE–CONTOUR MAP OF THE TOP OF THE PINEY POINT–NANJEMOY AQUIFER

CCNPP UNIT 3 FSAR

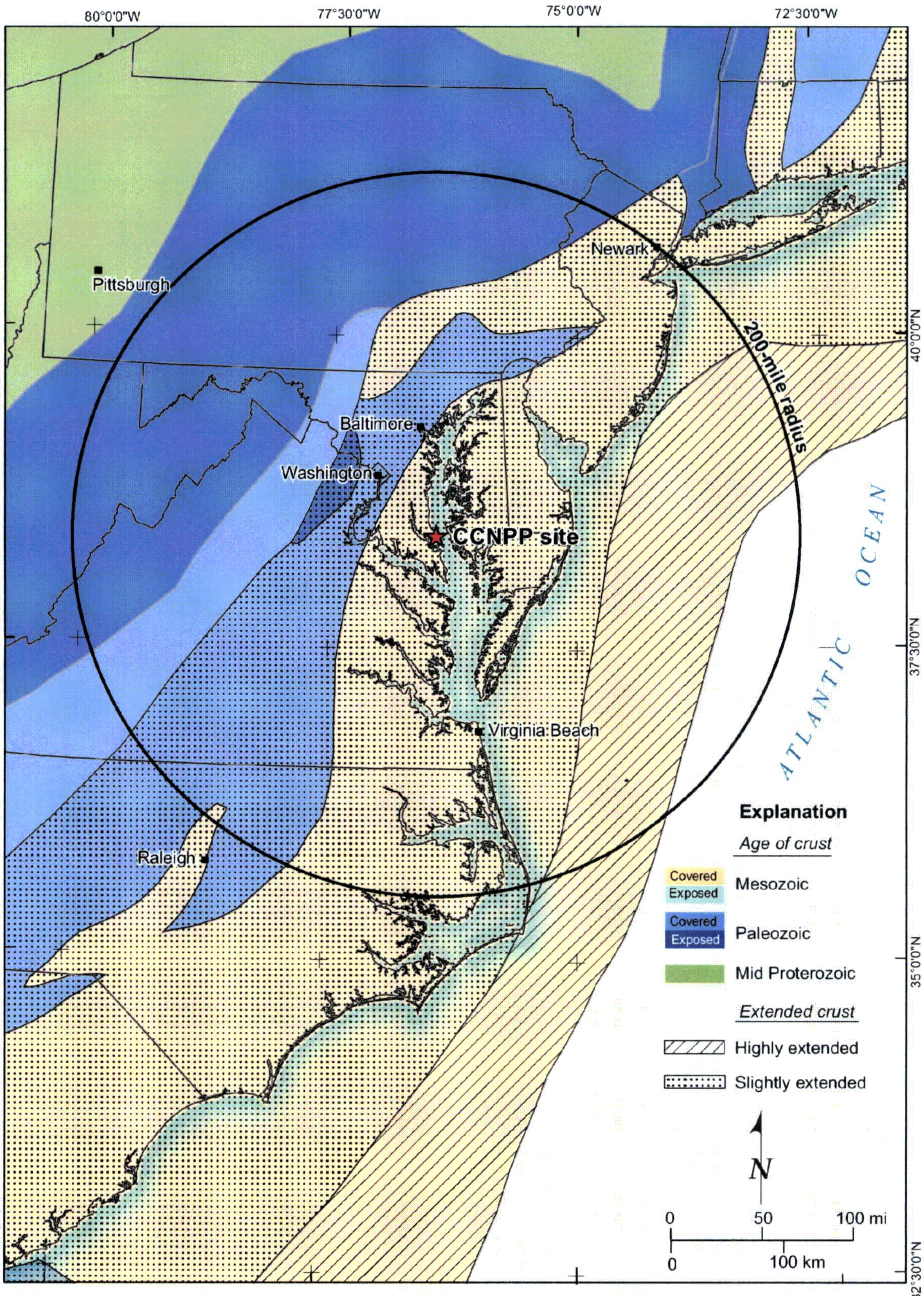
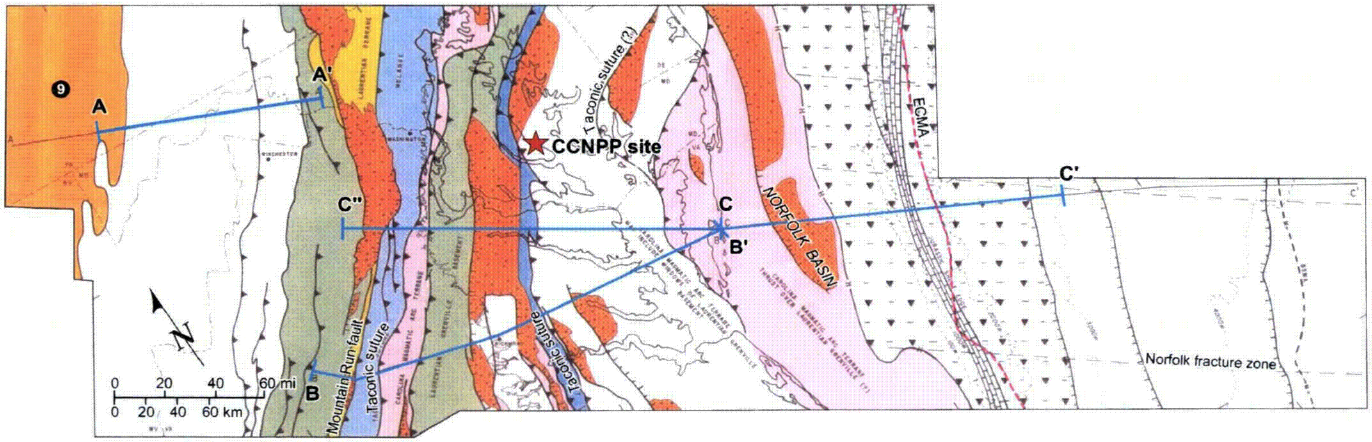


FIGURE 2.5.1-14 Rev. 0

CRUSTED AGES
FROM JOHNSTON et. al. (1994)

CCNPP UNIT 3 FSAR



Glover and Klitgord (1995)

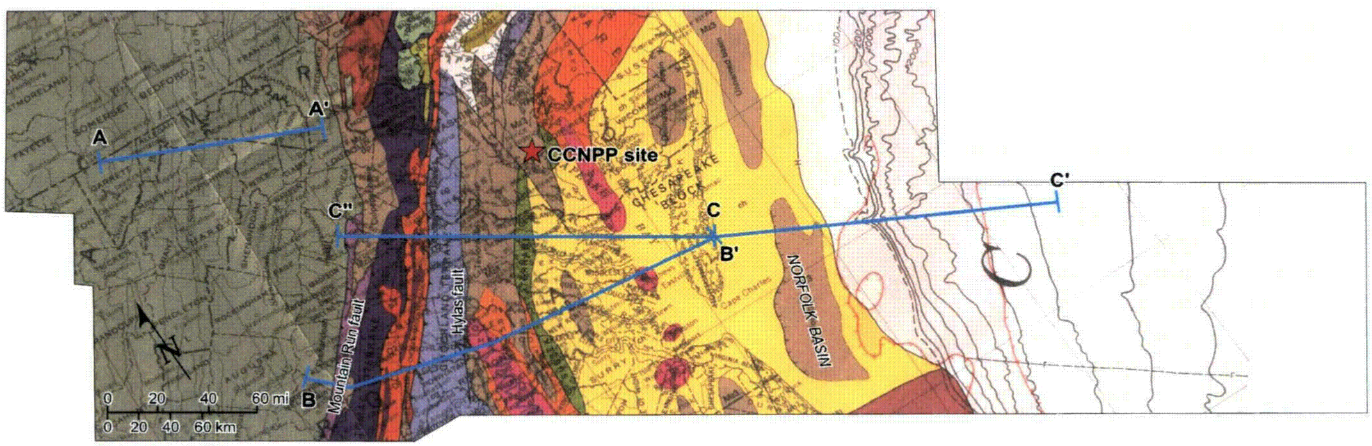
Explanation
(Glover and Klitgord, 1995)

Symbols

- Bathymetric isobath
- Normal fault
- Basement hinge zone
- C C' Cross section (see Figures 2.5.1-16, 17, and 18)
- Magnetic lineament

Map Units

- Cenozoic
- Early Mesozoic basin
- Cretaceous - stratified rock
- Permian - Upper Mississippian strafied rock
- Lower Jurassic - Upper Triassic stratified rock
- Jurassic volcanic wedge
- Cambrian - stratified rock
- Cambrian - intrusive rock



Horton et al. (1991)

Explanation
(Horton et al., 1991)

Symbols

- Basement hinge zone
- Mesozoic normal fault
- Thrust fault
- C C' Cross section line (see Figures 2.5.1-16, 17, and 18)
- Fault; solid where certain, dashed where inferred or approximate
- Magnetic low - less than 600 nanoteslas
- Magnetic high - greater than 1,000 nanoteslas

Map Units

- M_z3 Jurassic and Triassic Newark super group
- gr Late Paleozoic granitoids
- p^1, p^2 Other stitching plutons
- O_2 O_1 Late and Middle Ordovician sequences interpreted as Taconian successor basin deposits
- L Laurentia (undivided)
- aw Westminster terrane
- ib Baltimore terrane
- dje Jefferson terrane
- dp Potomac composite terrane
- ob Bel Air - Rising Sun terrane
- os Sussex terrane
- vcp Chopawamsic terrane
- vr Roanoke Rapids terrane
- cg Goochland terrane
- ch Chesapeake block

Gravity Anomalies on Delmarva Peninsula

- g1 Probable mafic complexes
- g2 Linear gravity high associated with migmatitic gneiss
- g3 Broad gravity low

FIGURE 2.5.1-15 Rev. 0

REGIONAL STRIP MAPS SHOWING TECTONOSTRATIGRAPHIC DIVISIONS AND REGIONAL CROSS-SECTION LINES

CCNPP UNIT 3 FSAR

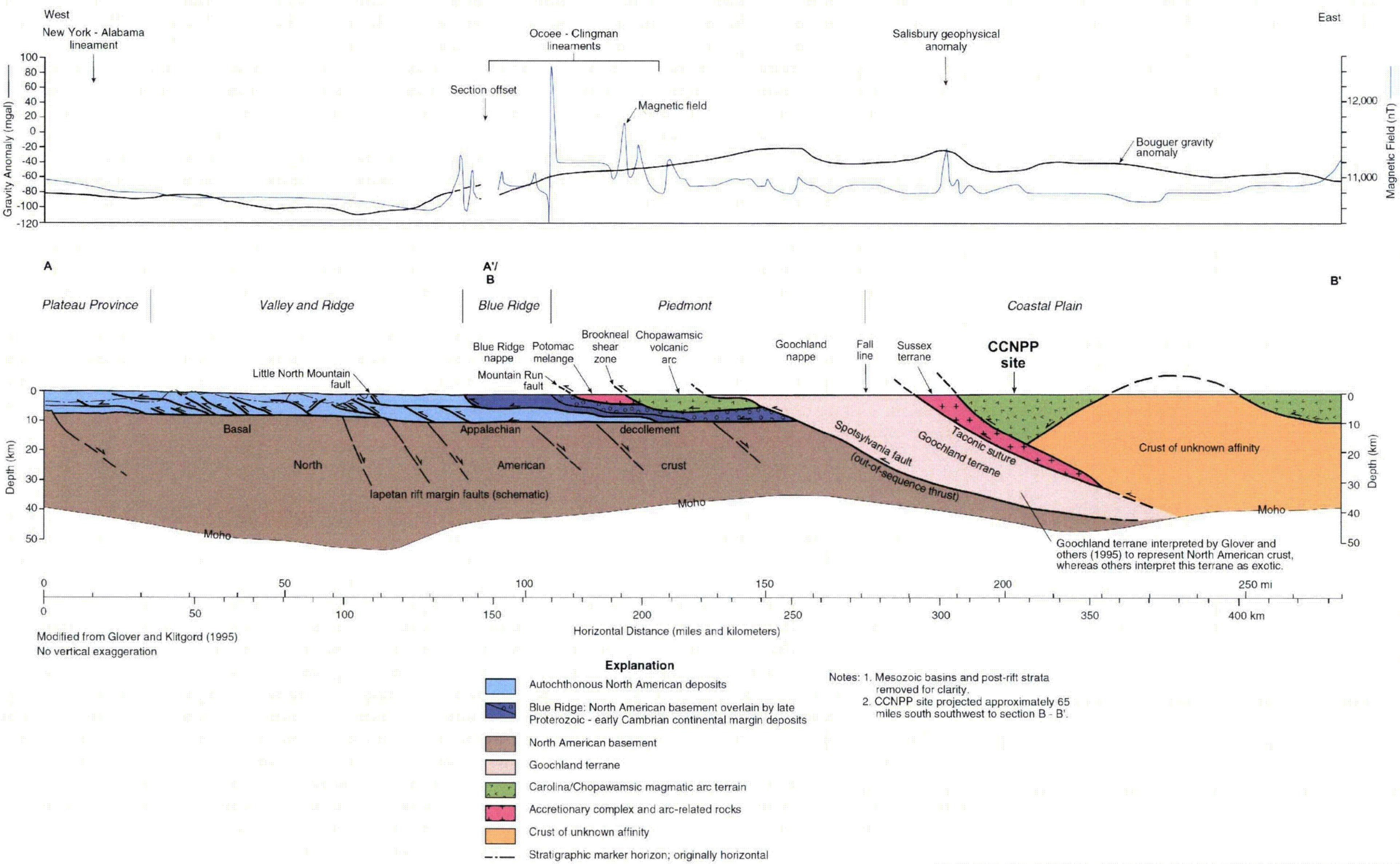
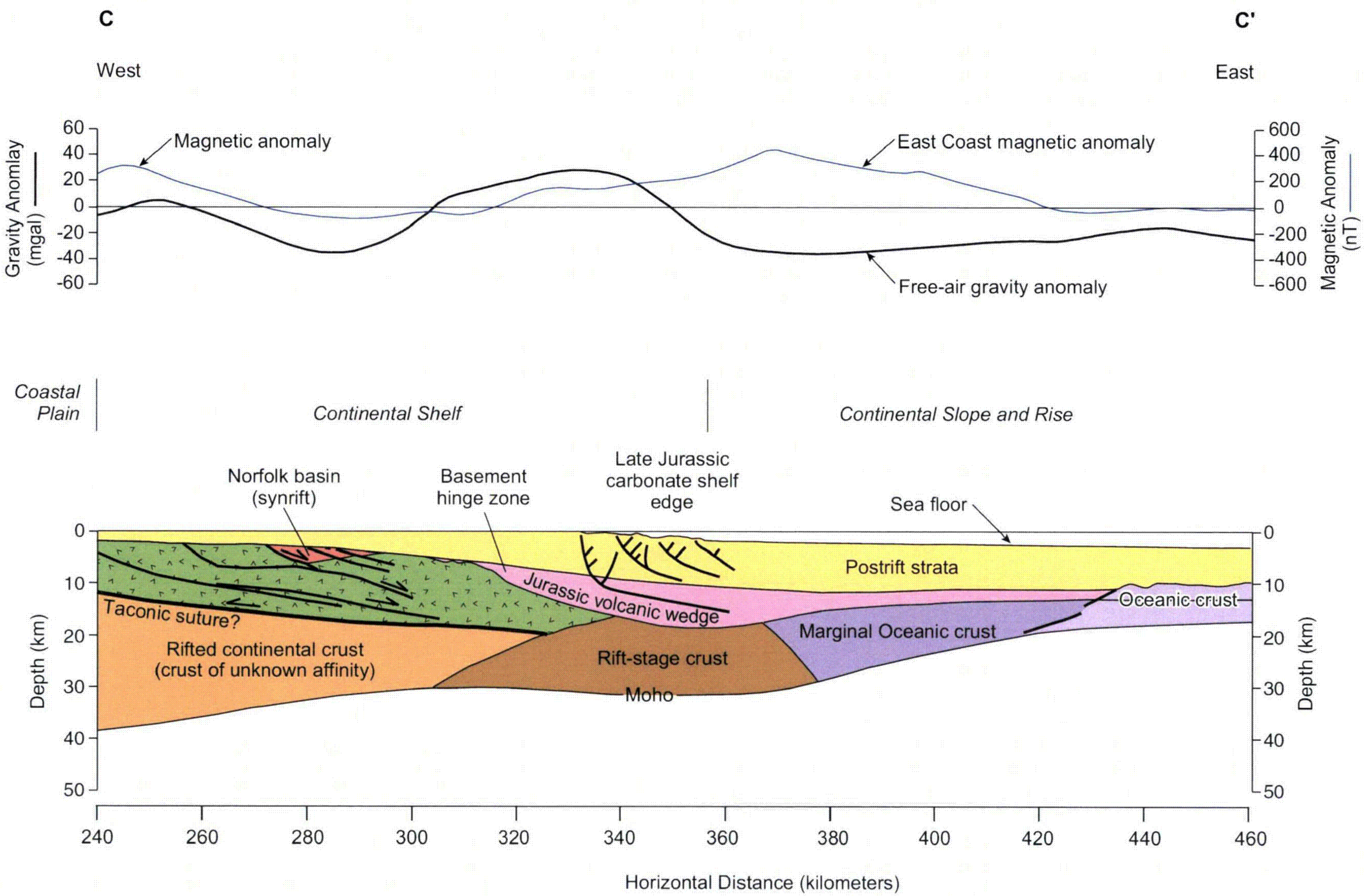


FIGURE 2.5.1-16 Rev. 0

CRUSTAL-SCALE CROSS SECTION THROUGH THE APPALACHIAN OROGEN AND COASTAL PLAIN

CCNPP UNIT 3 FSAR

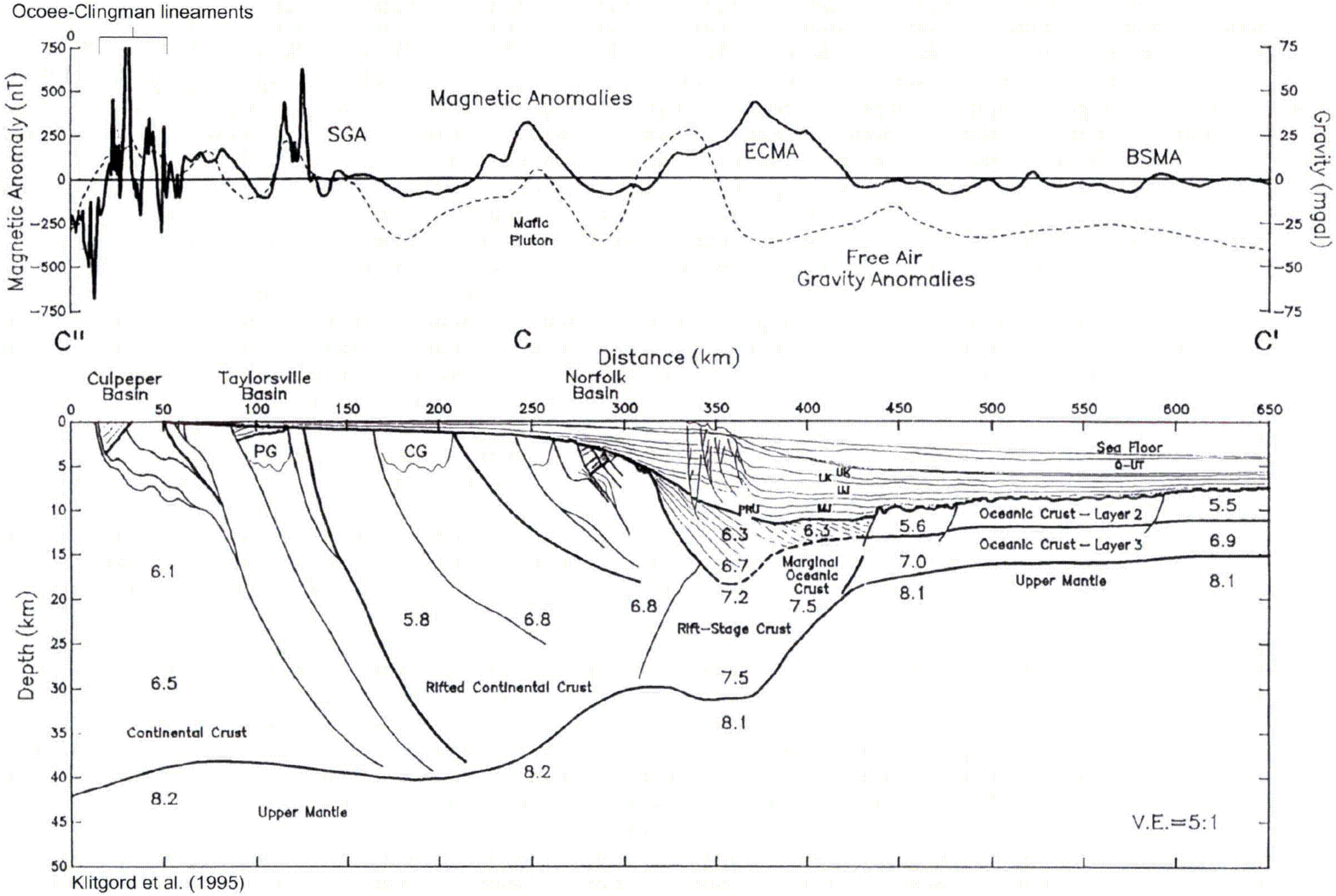


Modified from Glover and Klitgord (1995)

FIGURE 2.5.1-17 Rev. 0

CRUSTAL-SCALE CROSS SECTION ACROSS THE MID-ATLANTIC CONTINENTAL SHELF, SLOPE, AND RISE

CCNPP UNIT 3 FSAR

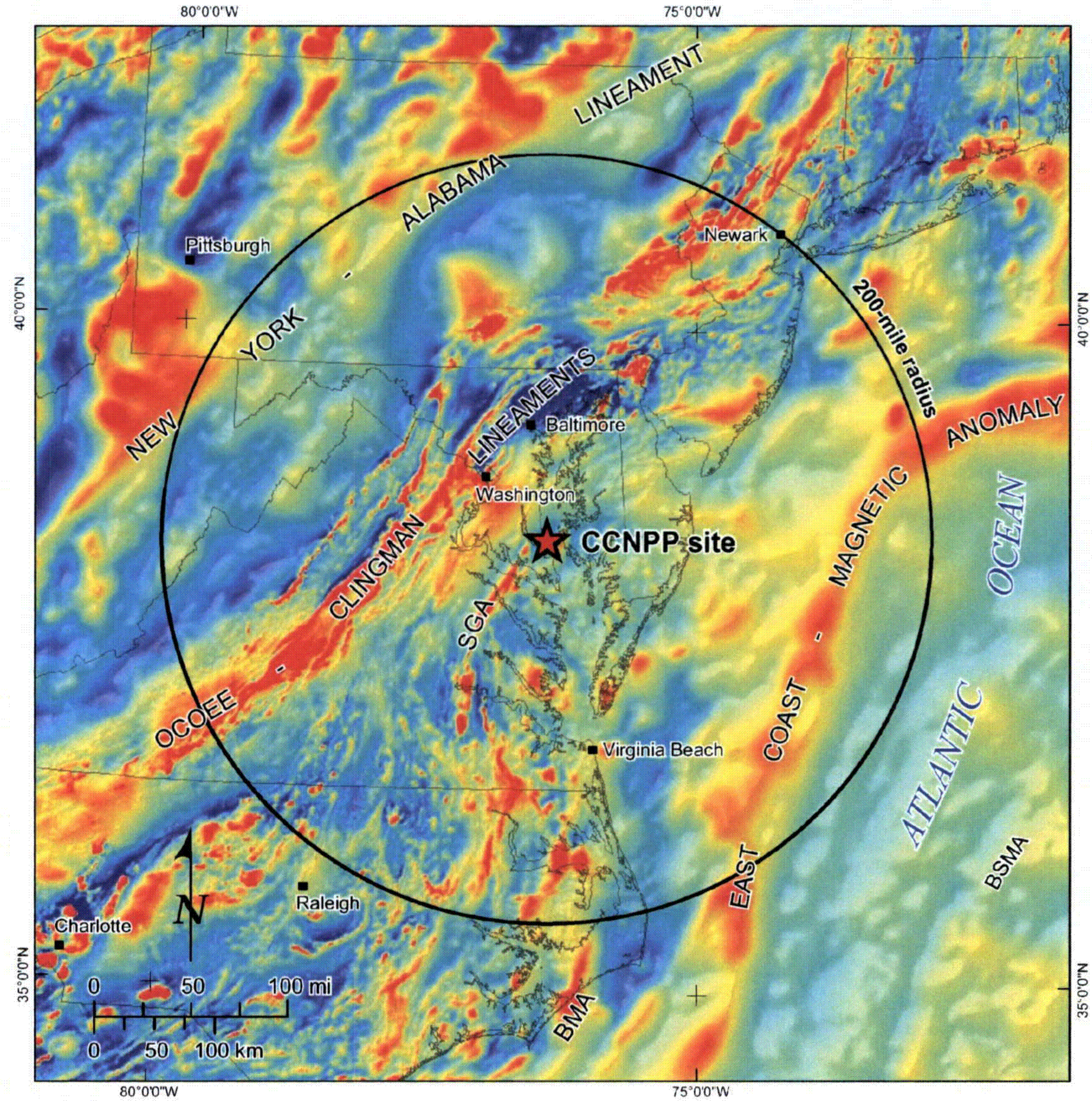


Cross section along line C" - C - C' displaying selected crustal fractures. Surface features along segment C" - C are taken directly from the geologic map panel. Subsurface features have been projected northward onto the profile from cross section B - B'. Magnetic and gravity anomaly profiles along the section and selected refraction velocity values (in km/sec) are shown. Major sub-horizontal crustal boundaries are indicated by heavy lines. Sedimentary strata are indicated by the light lines above the upper heavy line. SGA - Salisbury geophysical anomaly; ECMA = East Coast magnetic anomaly; BSMA = Blake Spur magnetic anomaly; PG = Petersburg Granite; CG = Chesapeake Granite. See Figure 2.5.1-15 for section location. C- C' is the same as Figure 2.5.1-17, but represents an alternative interpretation.

FIGURE 2.5.1-18 **Rev. 0**

CRUSTAL-SCALE CROSS SECTION OF THE MID-ATLANTIC PASSIVE MARGIN

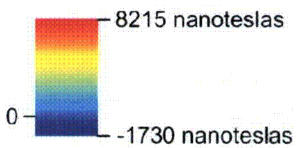
CCNPP UNIT 3 FSAR



Explanation

BMA	Brunswick magnetic anomaly
BSMA	Blake Spur magnetic anomaly
SGA	Salisbury geophysical anomaly

Aeromagnetics
(Bankey et al., 2002)

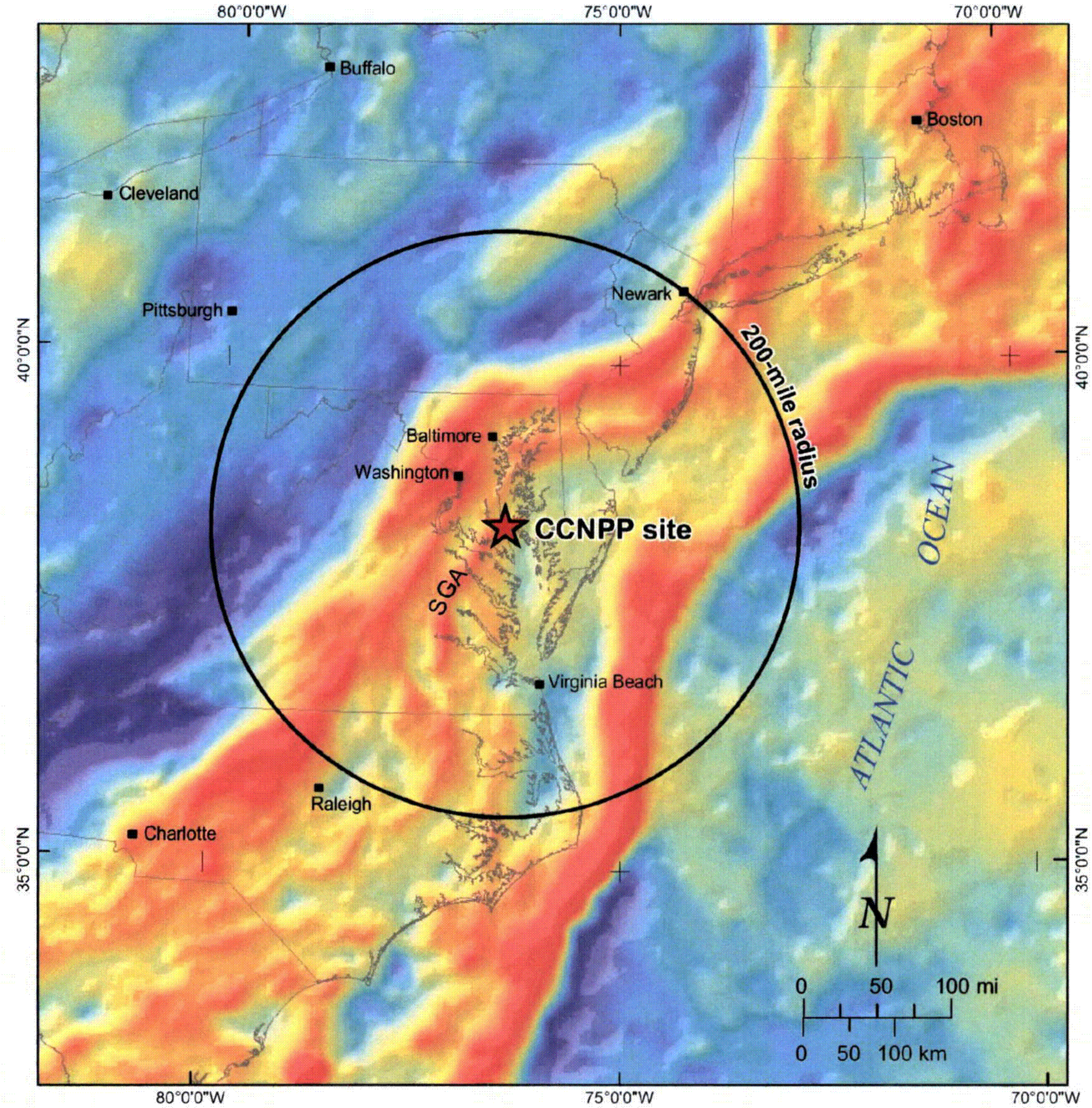


Note: Aeromagnetic data from Bankey et al. (2002).

FIGURE 2.5.1-19 Rev. 0

REGIONAL MAGNETIC ANOMALY MAP

CCNPP UNIT 3 FSAR



Explanation

SGA Salisbury geophysical anomaly

Gravity Anomaly

87 mgals

0

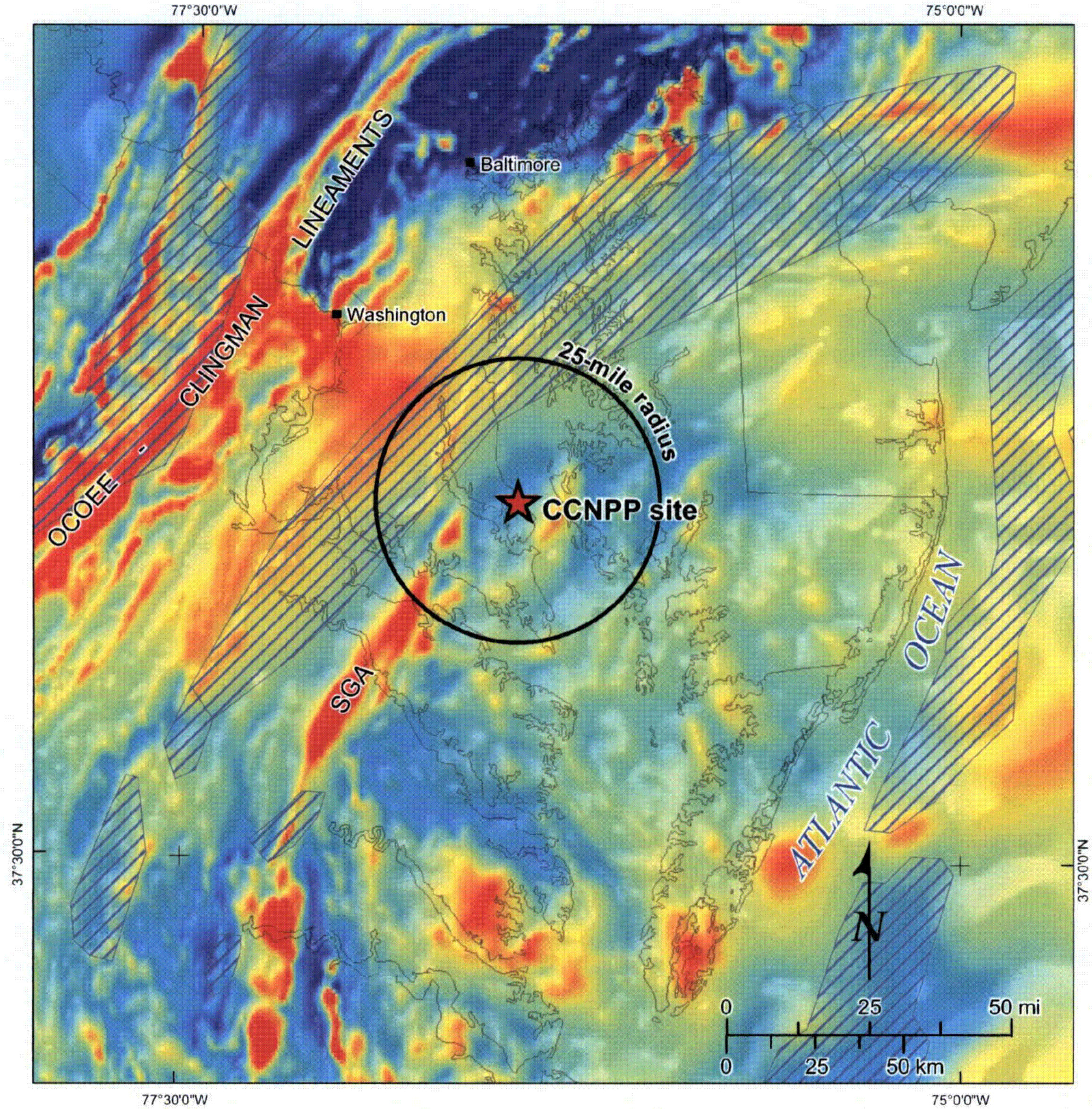
-110 mgals

Notes: 1. Gravity data from Hittelman et al. (1994).
2. Gravity measurements over land are Bouger gravity anomalies.
3. Gravity measurements over water are free-air anomalies.

FIGURE 2.5.1-20 Rev. 0

REGIONAL GRAVITY ANOMALY MAP

CCNPP UNIT 3 FSAR



Explanation

Mesozoic basin (Withjack et al., 1998)

SGA Salisbury geophysical anomaly

Aeromagnetics

8215 nanoteslas

0

-1730 nanoteslas

Note: Aeromagnetic data from Bankey et al. (2002).

FIGURE 2.5.1-21 Rev. 0

CHESAPEAKE BAY REGION MAGNETIC ANOMALIES WITH MESOZOIC BASINS

CCNPP UNIT 3 FSAR

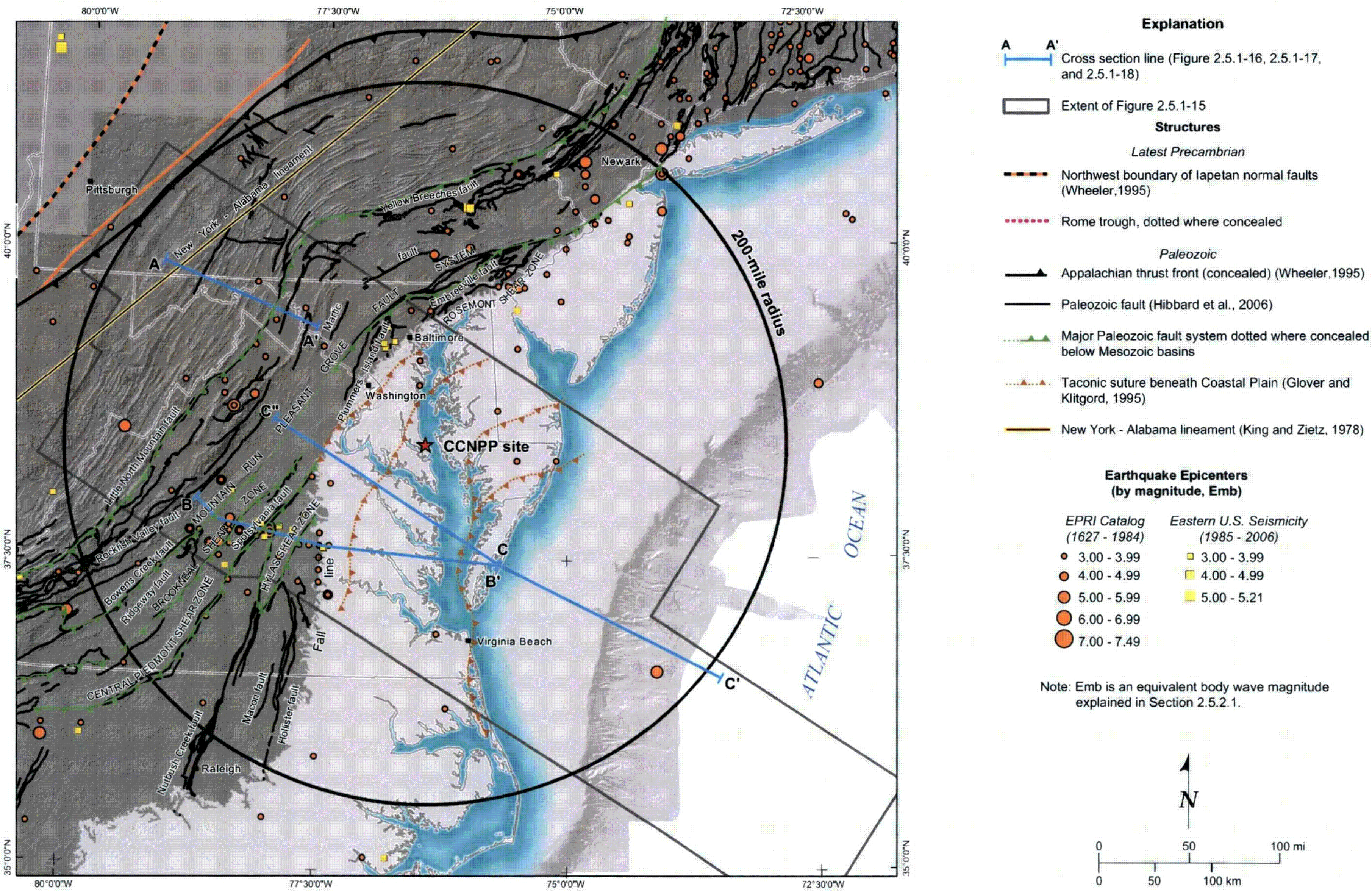


FIGURE 2.5.1-22 Rev. 0

LATE PROTEROZOIC AND PALEOZOIC TECTONIC FEATURES

CCNPP UNIT 3 FSAR

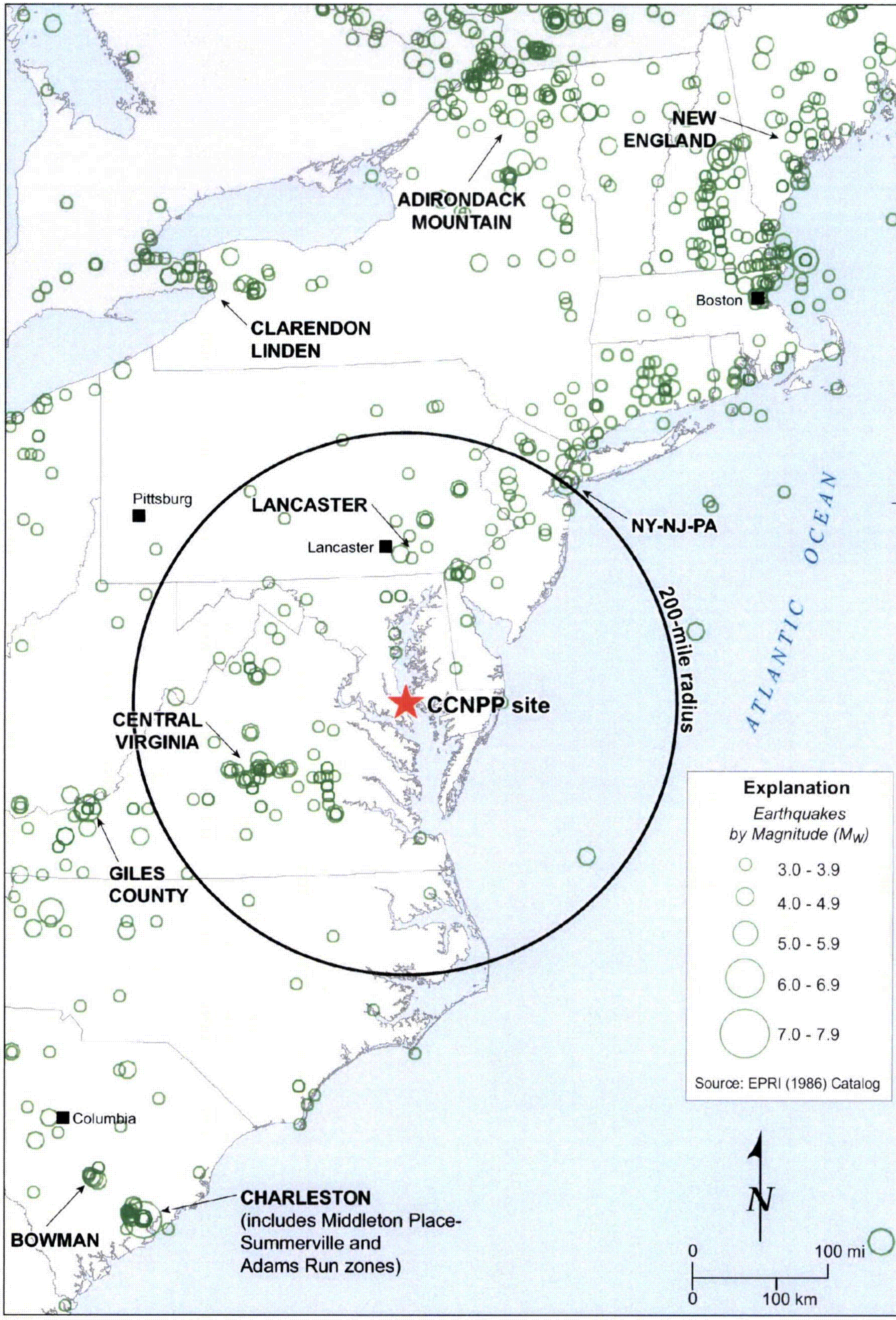
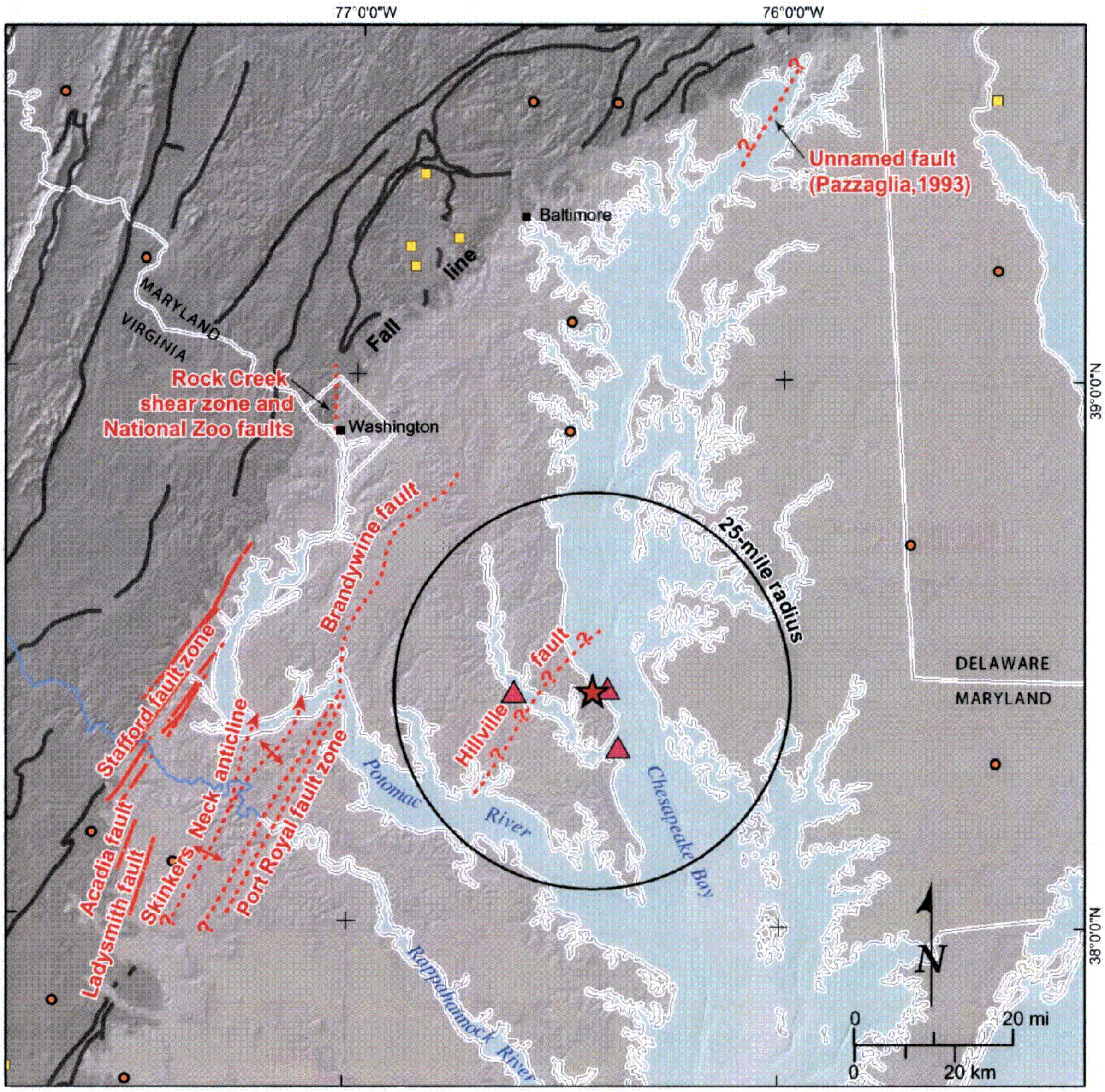


FIGURE 2.5.1-23 Rev. 0

SEISMIC ZONES AND SEISMICITY IN CEUS

CCNPP UNIT 3 FSAR



Explanation

Symbol	Description
★	CCNPP site
▲	McCartan et al. (1995) features
— - - ·· ?	Tertiary fault; dashed where uncertain; dotted where buried
—	Pre-Tertiary fault (Hibbard et al., 2006)
···↕···➤	Anticline

Physiographic Provinces

- Coastal Plain
- Piedmont

Earthquake Epicenters
(by magnitude, Emb)

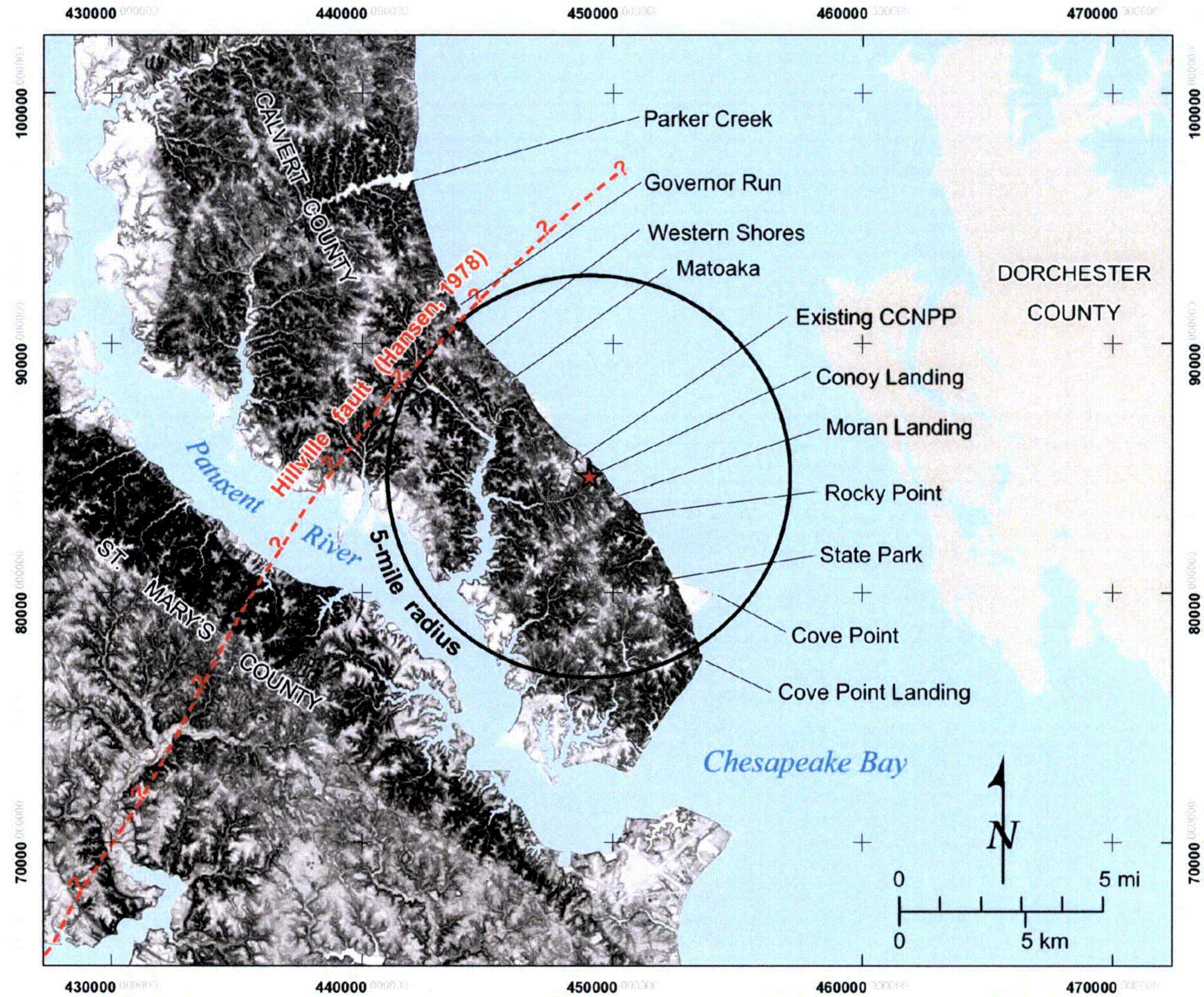
EPRI catalog (1627-1984)	*Eastern U.S. Seismicity (1985-2006)*
3.00 - 3.99	3.00 - 3.99
4.00 - 4.99	4.00 - 4.99
5.00 - 5.99	5.00 - 5.21
6.00 - 6.99	
7.00 - 7.40	

Note: Emb is an equivalent body wave magnitude explained in Section 2.5.2.1.

FIGURE 2.5.1-24 **Rev. 0**

MAP OF TERTIARY TECTONIC FEATURES

CCNPP UNIT 3 FSAR



Maryland State plane projection, NAD 83

Explanation

★ CCNPP Unit 3

? - - - - Tertiary fault buried

LiDAR slope map

Degrees slope

High : 83.5

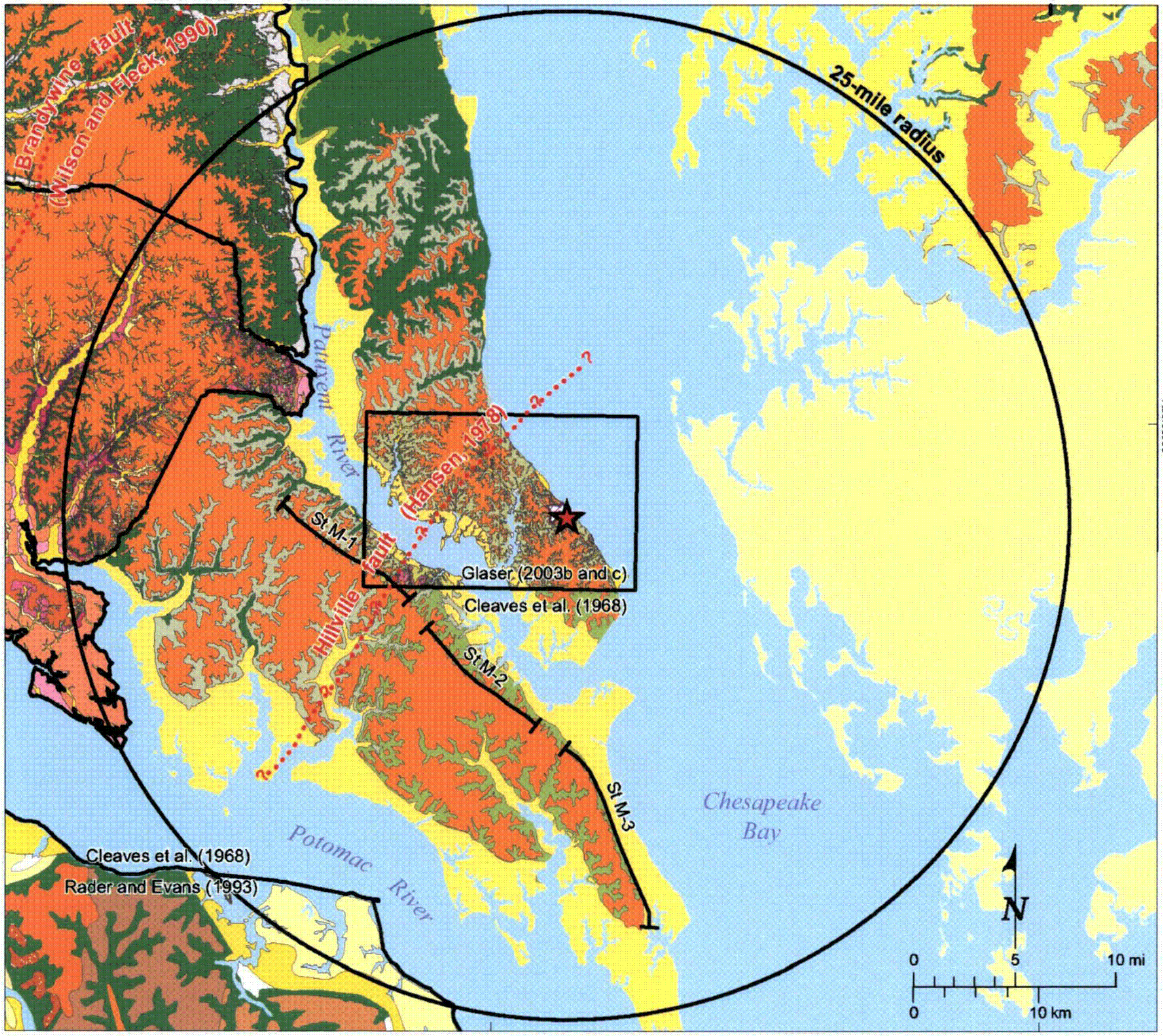
Low : 0.0

Note: LiDAR data for Calvert and St. Mary's County has an ~2-meter resolution. Text leadlines refer to reference locations from Kidwell (1997).

FIGURE 2.5.1-25 Rev. 0

LiDAR DATA FOR CALVERT AND ST. MARY'S COUNTIES

CCNPP UNIT 3 FSAR

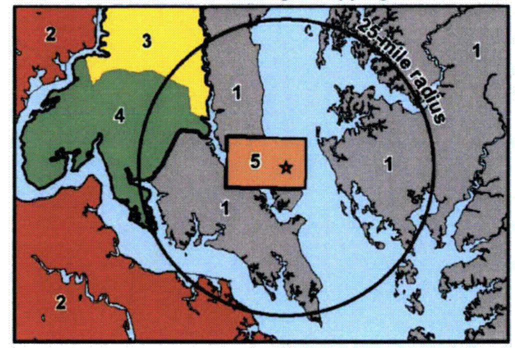


Projection: UTM Zone 18 NAD83

76°30'0"W

76°0'0"W

Index of Geologic Mapping



1 State Map of Maryland (Cleaves et al., 1968) digitized by Dicken et al. (2005)

2 State Map of Virginia (Rader and Evans, 1993) digitized by Dicken et al. (2005)

3 Geologic map of Prince George's County (Glaser, 2003a)

4 Geologic map of Charles County (McCartan, 1989a)

5 Geologic maps of Cove Point and Broomers Island 7.5-minute quadrangles (Glaser, 2003b and c)

Explanation

CCNPP Unit 3

Source map boundary

Tertiary fault (buried)

St M-1 Seismic line (St M-1 through ST M-3) (Hansen, 1978)

For description of map units see Figure 2.5.1-26b

FIGURE 2.5.1-26a Rev. 0

SITE VICINITY GEOLOGIC MAP
25-MILE (40-Km) RADIUS

CCNPP UNIT 3 FSAR

Explanation

Maryland
(Cleaves, et al., 1968)

Quaternary – Pleistocene to recent
- Qdu – Quaternary undivided - sand
- Ql – Lowland deposits - gravel
- Qu – Upland deposits (Eastern Shore) - gravel

Pliocene(?)
- QTu – Quaternary and Tertiary deposits (undifferentiated) - gravel

Chesapeake Group

Miocene
- Ts – St. Marys Formation - clay
- Tch – Choptank Formation - sand
- Tc – Calvert Formation - sand

Eocene
- Tn – Nanjemoy Formation - sand

Paleocene
- Ta – Aquia Formation - sand

Virginia
(Rader and Evans, 1993)

Quaternary
- sb – Beach and dune sand deposits
- al – Quaternary, alluvium
- Qtlp – Lynnhaven and Poquoson members undifferentiated (Tabb Formation) - sand
- Qtl – Lynnhaven Member (Tabb Formation) - sand
- Qts – Sedgefield member (Tabb Formation) - sand
- QTu – Tertiary-Quaternary deposits undifferentiated - sand
- Qsh – Shirely Formation (Omar Fm equivalent) - gravel
- Qcc – Charles City Formation - sand

Pliocene
- Tm – Moorings unit of Oaks and Coch (1973) - sand
- Tb1 – Bacons Castle Formation - gravel
- Tb2 – Bacons Castle Formation - sand
- psg – Pliocene sand and gravel

Chesapeake Group

Miocene
- Tc – Chesapeake group - sand

Prince George's County
(Glaser, 2003a)

Quaternary – Holocene
- Qal – Alluvium - interbedded sand silt-clay and subordinate gravel

Quaternary – Pleistocene
- Qt – Terrace deposits (informal unit) - interbedded sand gravel and silt-clay

Upper Miocene Pliocene
- Tu – Upland deposits (informal unit) - sand pebbly sand and gravel capped by sandy pebbly loam in places

Chesapeake Group

Lower to Middle Miocene
- Tc – Calvert Formation - fine- to very fine-grained sand and silt, as well as diatomaceous silt

Lower Eocene
- Tn – Nanjemoy Formation - clay and minor interbedded silt

Paleocene
- Tm – Marlboro clay, clay or silty clay; sparse lenses silt
- Ta – Aquia Formation, sand (variably glauconitic)

Charles County
(McCartan, 1989a)

Quaternary – Holocene
- Qh – Holocene deposits undivided - sand gravel silt and clay
- Czcu – Cenozoic colluvium undivided - clay to cobble-sized material; poorly sorted

Quaternary – Pleistocene
- Qk – Kent Island Formation - silty sand (fine-to medium-grained)
- Qm – Maryland Point Formation - sand, silty clay, and clay
- Qo – Omar Formation - mud and muddy fine sand grading downward to fine gravel
- Qcx – Chicamuxen Church Formation - silty clay and muddy fine sand grading downward to pebbly mud and sand

Pliocene
- Tp – Park Hall Formation, silty sand, sand, and clay interbedded with coarse sand, pebbles, and boulders
- Tug4 – Upland gravel 4, sand; lenses of muddy sand and clay grading downward to gravelly mud or coarse sand
- Tug3 – Upland gravel 3 - mainly interbedded sand and gravel with coarse sand matrix

Chesapeake Group

Middle Miocene
- Tch – Choptank Formation - fine sand to silty sand interbedded with silty clay
- Tc – Calvert Formation - very fine sand and fine sandy clay; rare medium to coarse sand

Lower Eocene
- Tn – Nanjemoy Formation - glauconitic sand (fine- to med-grained) and silty clay

- Tm – Marlboro clay, clay and minor interbedded silt
- Ta – Aquia Formation, micaceous glauconitic sand (fine- to coarse-grained)

Broomers Island and Cove Point Quadrangles
(modified from Glaser, 2003 b and c)

Quaternary – Holocene
- af – Artificial fill - heterogeneous unstratified materials, such as sand, gravel, clay, slag, construction debris, and dredge spoil
- Qal – Alluvium - interbedded sand, gravel, and silt-clay; organic matter in places; rare thin peats
- Qb – Holocene beach deposits - loose, cross stratified sand and marshes

Quaternary – Pleistocene
- Qz – Lowland deposits (informal unit) - interbedded sand, silt-clay and minor pebbly sand; fossiliferous in places; thin gravels at base
- Qcx – Chicamuxen Church formation - interbedded sand and gravel; uppermost 4 to 5 feet consist of silty clay or silty fine sand
- Qt – Terrace deposits (informal unit) - sand and gravel

Pliocene
- Tp – Park Hall Formation - interbedded sand and minor clay-silt and pebbly sand and gravel
- Tu – Upland deposits (informal unit) - sand, gravel and minor clay-silt

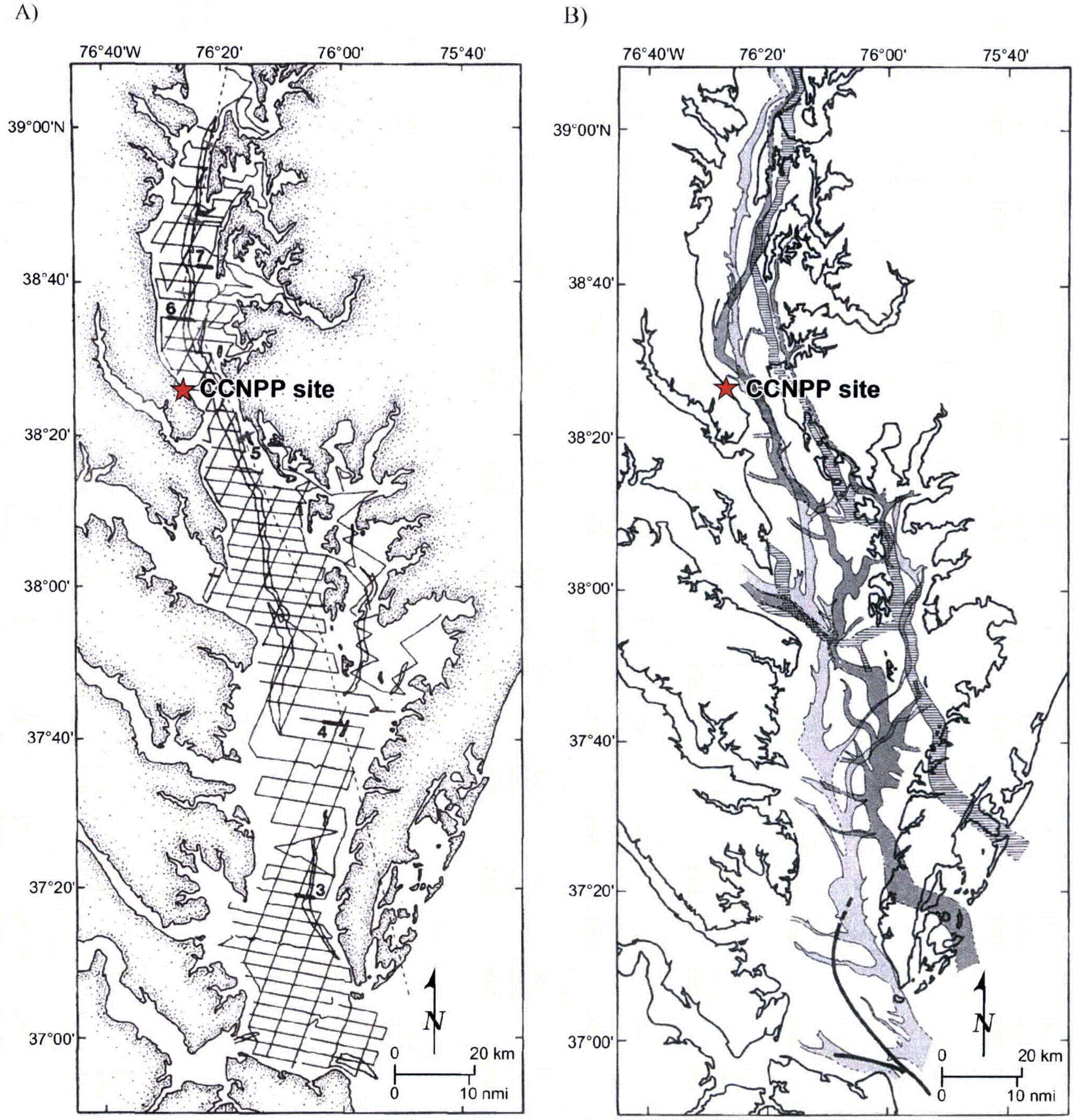
Chesapeake Group

Middle Miocene
- Tcsm – Choptank and St. Marys Formations undivided - sand, sandy clay and clay; variably fossiliferous
- Tcp – Calvert Formation, Plum Point marls - sand and clayey sand; fossiliferous at several horizons yet sparingly fossiliferous throughout
- Tcf – Calvert Formation, Fairhaven member - sand and diatomaceous silt with clay - diatomaceous silt in lower portion; homogeneous sand and silty sand in upper portion (mapped in subsurface only)

FIGURE 2.5.1-26b Rev. 0

SITE VICINITY GEOLOGIC MAP
25-MILE (40-Km) RADIUS
UNIT DESCRIPTIONS

CCNPP UNIT 3 FSAR



Note: Black lines represent track lines of seismic reflection profiles within Chesapeake Bay used to map paleo-Susquehanna River channels.

Explanation

- Cape Charles
- Eastville
- Exmore
- Modern tidal

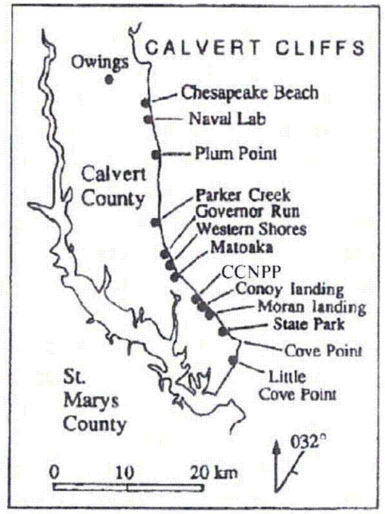
Notes: 1. Shading marks paleochannel extents.
2. Paleo-channel ages described in text.

FIGURE 2.5.1-27 Rev. 0

MAP OF SEISMIC LINES (A) AND PALEO–SUSQUEHANNA RIVER CHANNELS (B) (COLEMAN et. al., 1990)

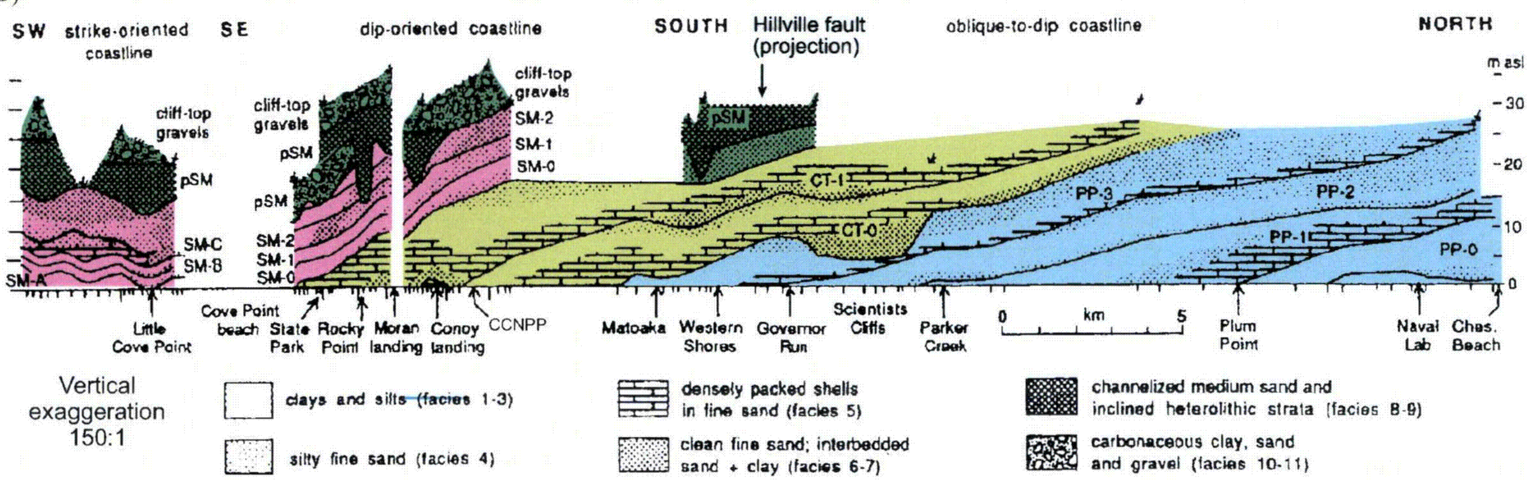
CCNPP UNIT 3 FSAR

A)



Location of landmarks along the Calvert Cliffs (refer to Figure 2.5.1-25).

B)

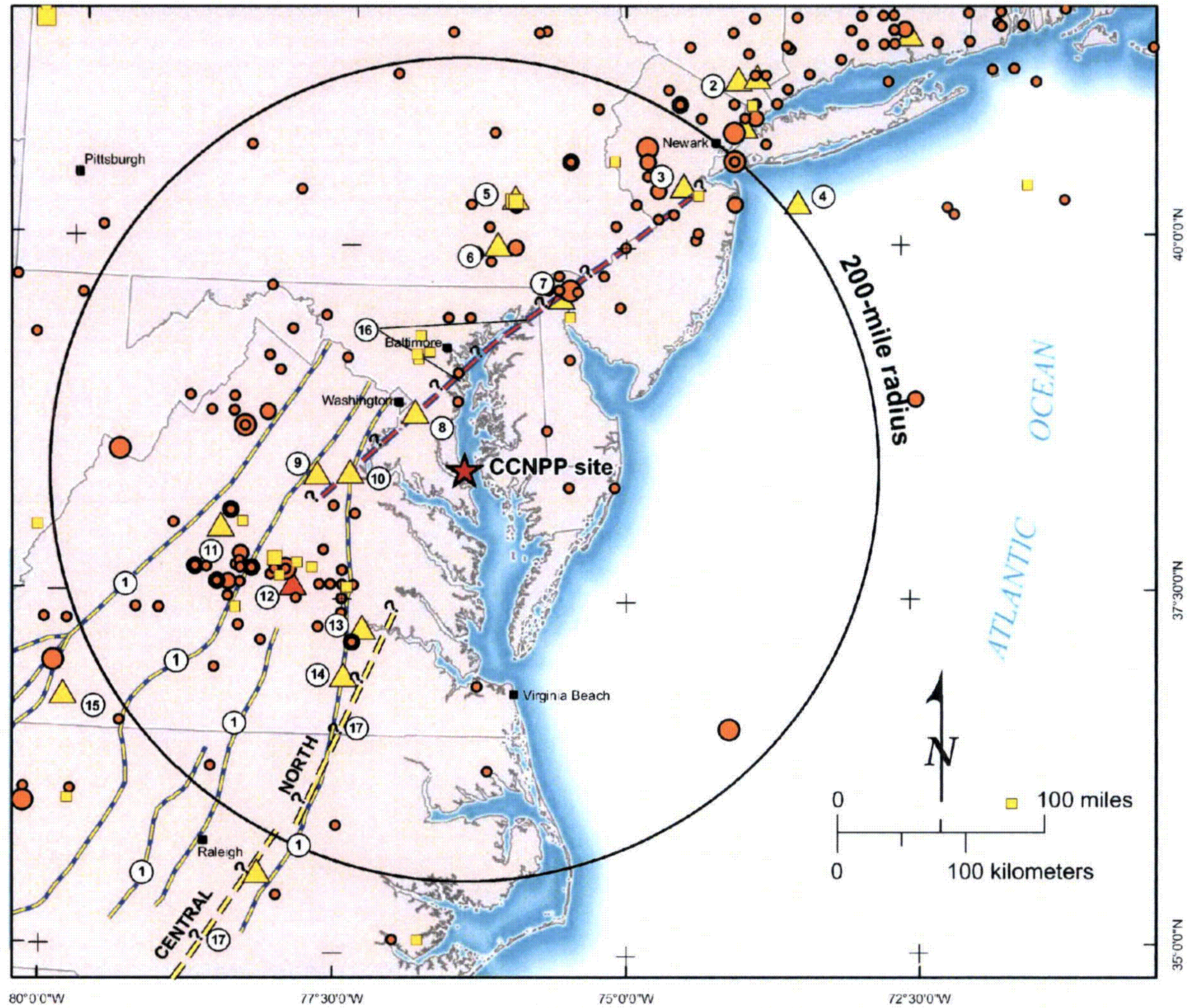


Cross section of the Calvert Cliffs showing disconformity-bounded units. PP, Plum Point Member of the Calvert Formation; CT, Choptank Formation; SM, St. Marys Formation; pSM, post-St. Mary's Neogene strata.

FIGURE 2.5.1-28 Rev. 0

LOCATION MAP (A) AND CROSS SECTION (B) OF CALVERT CLIFFS, MARYLAND

CCNPP UNIT 3 FSAR



Explanation

Features* compiled from Crone and Wheeler (2000) and Wheeler (2005)

Class

- A
- C
- C East coast fault system (Marple and Talwani, 2000)
- C Weems fall lines (1998) (Marple, 2004)
- C Stafford fault system (Marple, 2004)

Earthquake Epicenters (by magnitude, Emb)

EPRI Catalog (1627 - 1984)	*Eastern U.S. Seismicity (1985 - 2006)*
3.00 - 3.99	3.00 - 3.99
4.00 - 4.99	4.00 - 4.99
5.00 - 5.99	5.00 - 5.21
6.00 - 6.99	
7.00 - 7.49	

*Refer to table below for feature names

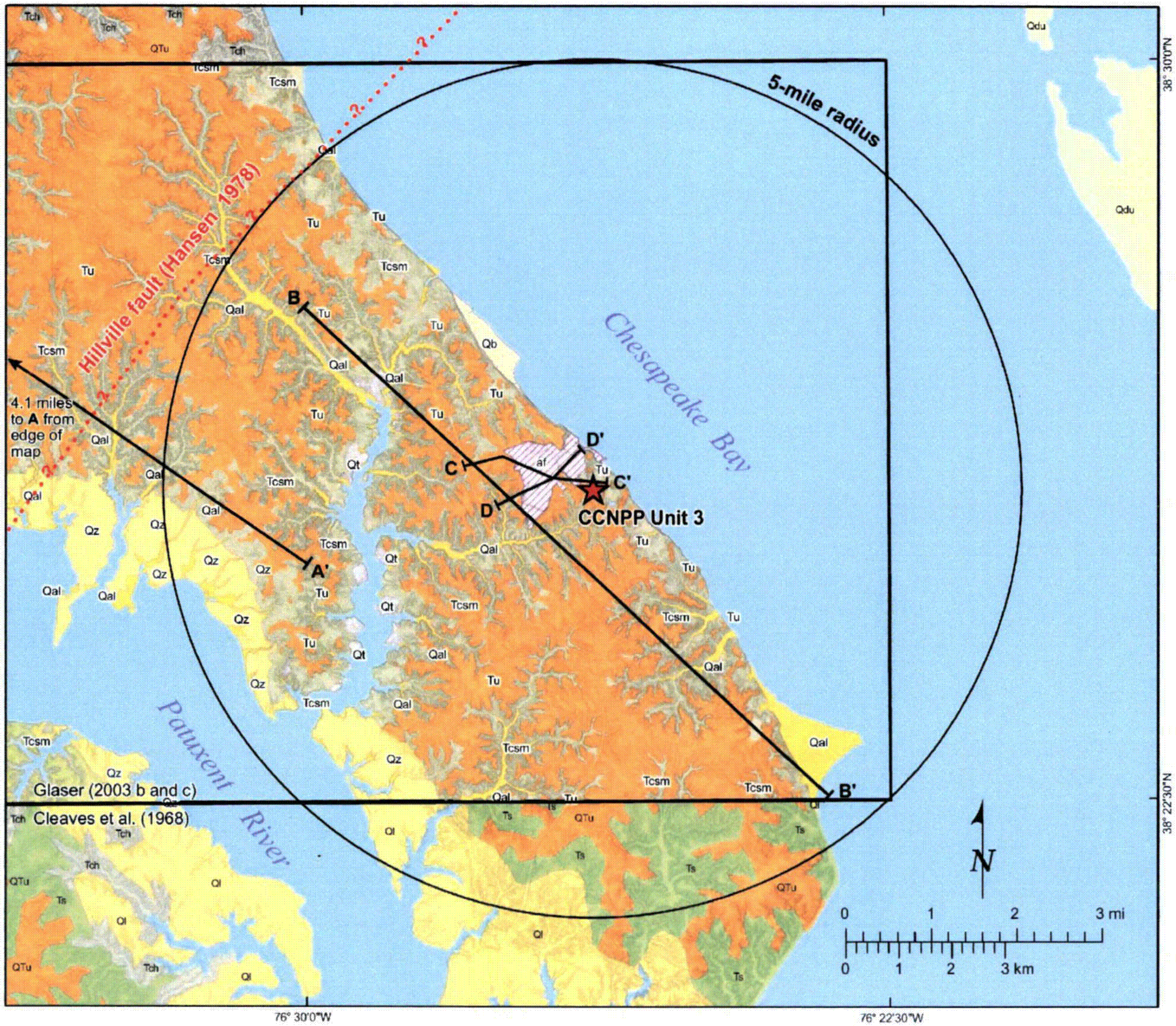
1. Fall lines (Weems, 1988)
2. Ramapo fault system
3. Kingston fault
4. New York Bight fault (offshore)
5. Cacoosing Valley earthquake
6. Lancaster seismic zone
7. New Castle County faults
8. Upper Marlboro faults
9. Everona fault-Mountain Run fault zone
10. Stafford fault (Mixon et al., 2000)
11. Lebanon Church fault
12. Central Virginia seismic zone
13. Hopewell fault
14. Old Hickory faults
15. Stanleytown-Villa Heights faults
16. Stafford fault system of Marple (2004)
17. East Coast fault system (Marple and Talwani, 2007)

Note: Emb is an equivalent body wave magnitude explained in Section 2.5.2.1.

FIGURE 2.5.1-29 Rev. 0

POTENTIAL QUATERNARY FEATURES IN THE SITE REGION

CCNPP UNIT 3 FSAR



Explanation

?...... Tertiary fault (buried)

Geologic map boundary

B B' Cross section (see Figures 2.5.1-28, 2.5.1-29, and 2.5.1-30)

Cove Point and Broomers Island 7.5-minute geologic maps (Glaser, 2003 b and c)

af	Artificial fill
Qb	Holocene beach deposits
Qal	Alluvium
Qz	Lowland deposits (informal unit)
Qt	Terrace deposits (informal unit)
Tu	Upland deposits (informal unit)
Tcsm	Choptank and St. Marys Formations undivided

Geologic map of Maryland (Cleaves et al., 1968)

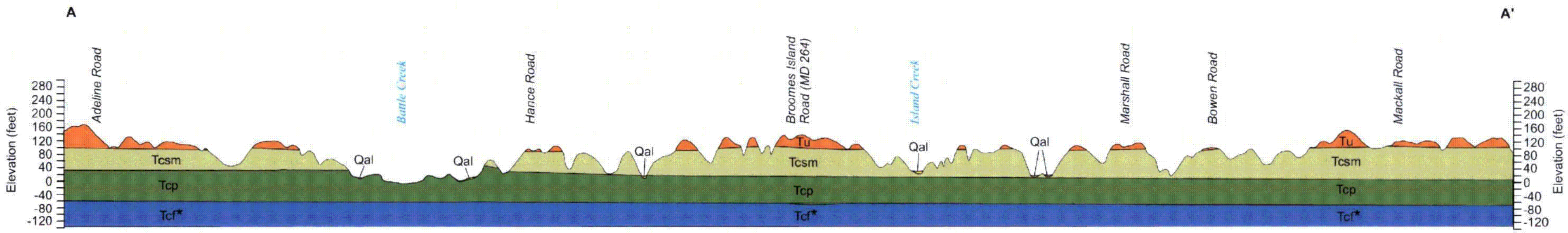
Qdu	Quaternary, sand
Ql	Quaternary, gravel
QTu	Tertiary-Quaternary, sand
Ts	Miocene, sand
Tch	Miocene, sand

* For a detailed description of map units see Figure 2.5.1-26b

FIGURE 2.5.1-30 **Rev. 0**

SITE AREA GEOLOGIC MAP
5-MILE (8-Km) RADIUS

CCNPP UNIT 3 FSAR

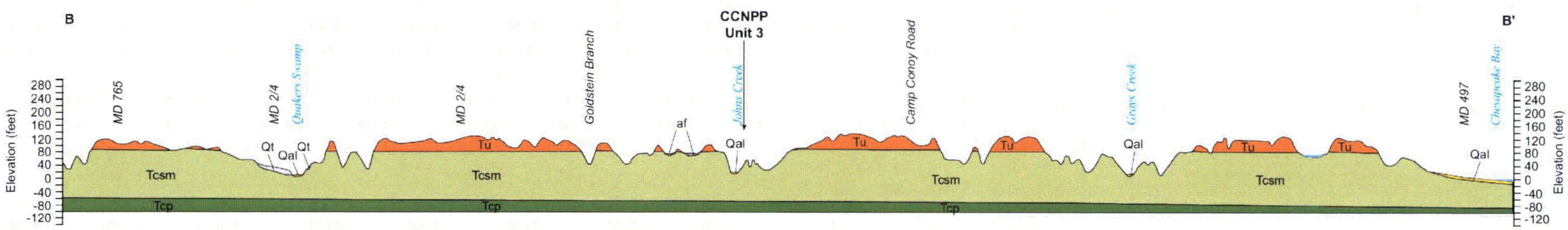


Vertical exaggeration 10X

Modified from Glaser (2003b)

Tcf* - Calvert Formation - Fairhaven member (mapped in subsurface)

For unit descriptions see Figure 2.5.1-26b Site Vicinity Geologic Map (25-mile radius) Unit Descriptions.



Vertical exaggeration 10X

Modified from Glaser (2003c)

For unit descriptions see Figure 2.5.1-26b Site Vicinity Geologic Map (25-mile radius) Unit Descriptions.

FIGURE 2.5.1-31 Rev. 0

SITE AREA GEOLOGIC CROSS SECTIONS A–A' and B–B' 5–MILE (8–Km) RADIUS

CCNPP UNIT 3 FSAR

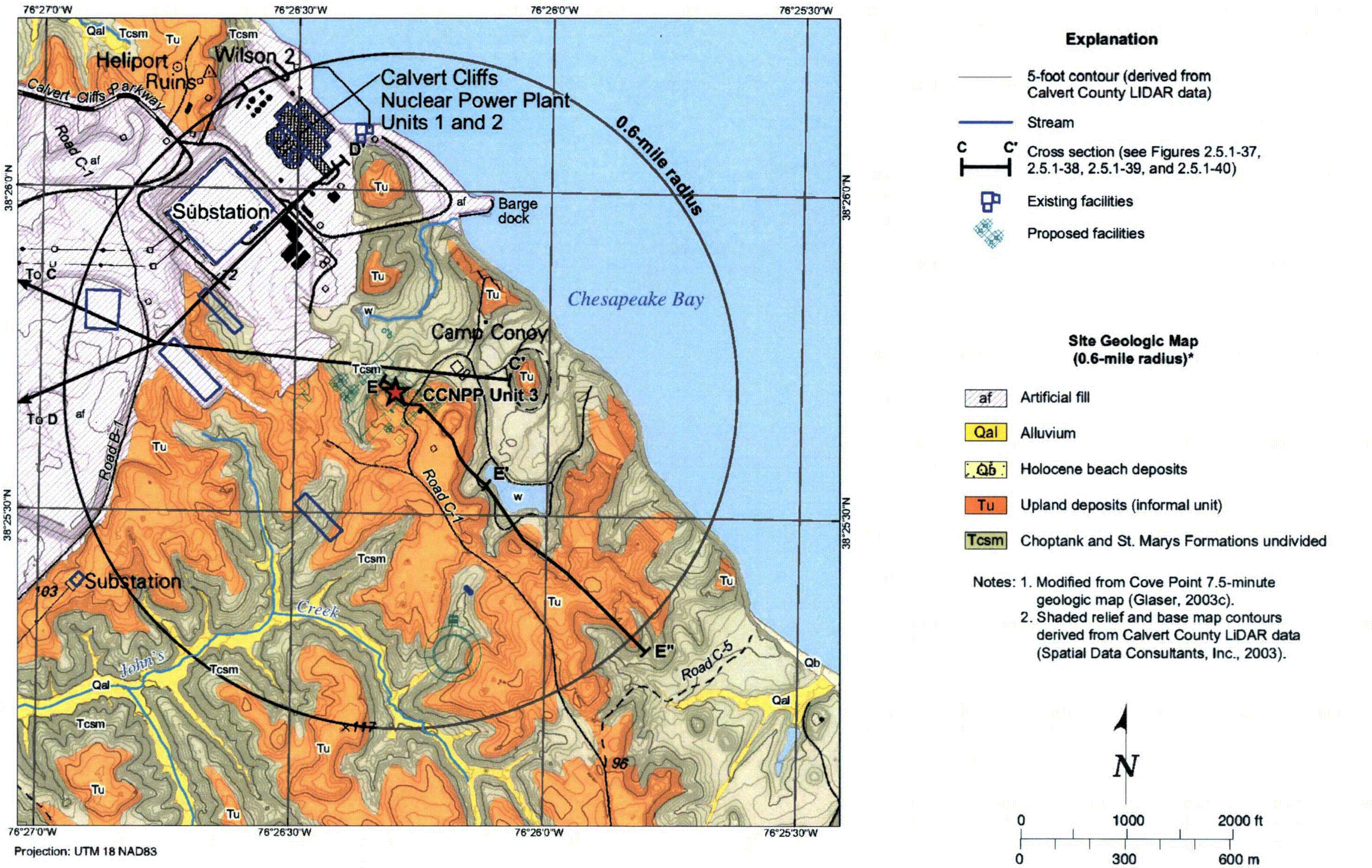


FIGURE 2.5.1-32 **Rev. 0**

SITE GEOLOGIC MAP
0.6-MILE (1-Km) RADIUS

CCNPP UNIT 3 FSAR

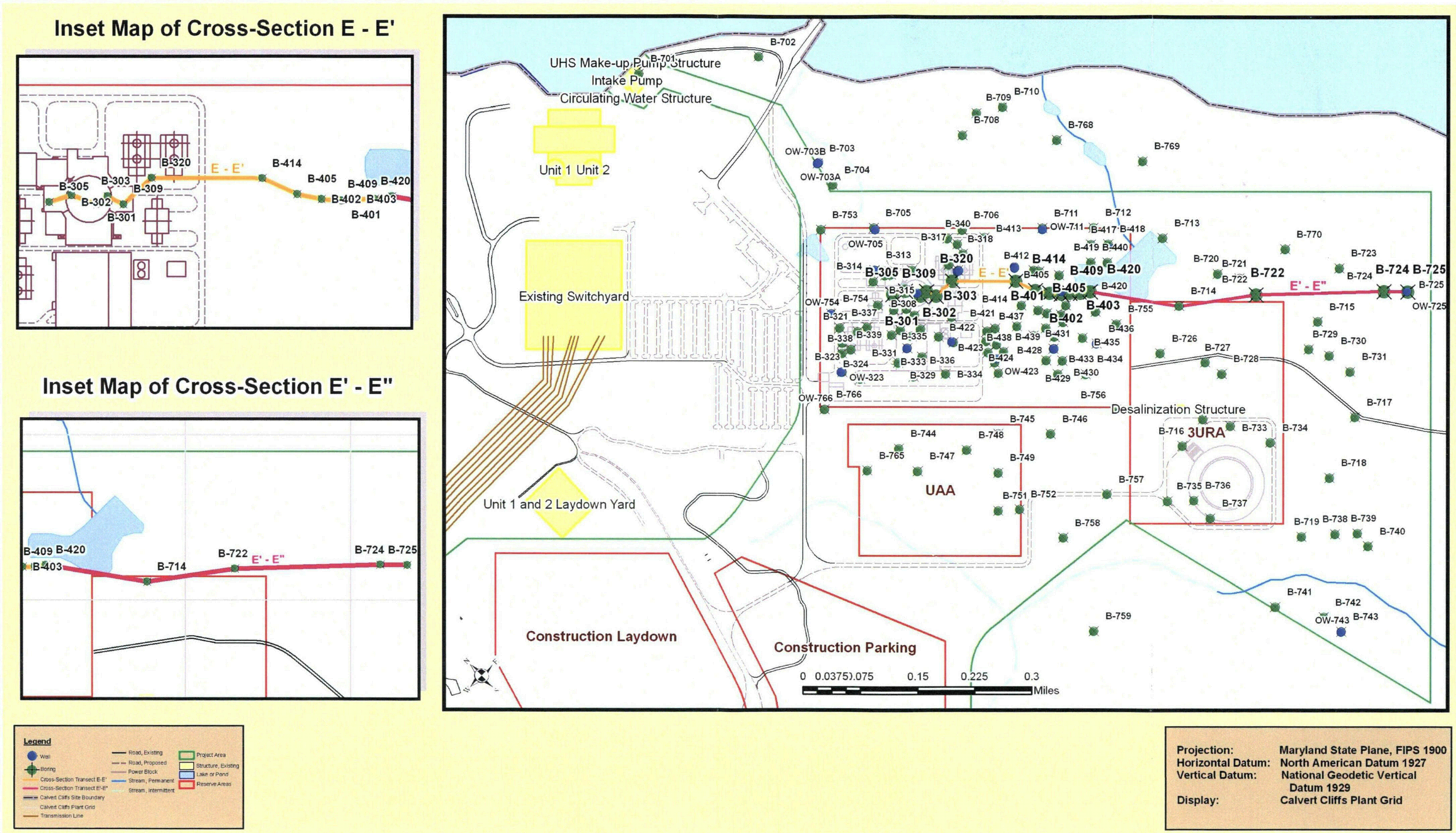


FIGURE 2.5.1-33 **Rev. 0**

BORING LOCATION MAP CALVERT CLIFFS NUCLEAR POWER PLANT UNIT 3

CCNPP UNIT 3 FSAR

ERA	PERIOD	EPOCH		AGE (Ma)	UNIT	THICKNESS (FT)
Cenozoic	Quarter-nary	Holocene		0.01	**Alluvium & Beach Deposits**	0-50
		Pleistocene		1.8	**Terrace & Lowland Deposits**	
	Tertiary	Pliocene		5.3	**Upland Deposits**	0-50
		Miocene	Upper	11.2		
			Middle	16.4	**Chesapeake Group** St. Marys Formation Choptank Formation Calvert Formation	245-280
		Eocene	Middle	49	**Piney Point Formation**	20
			Lower	54.8	**Nanjemoy Formation**	180
		Paleocene	Upper	61	**Marlboro Clay** **Aquia Formation**	165-170
			Lower	65	**Brightseat Formation**	10-20
Mesozoic	Cretaceous	Upper		99	**Magothy, Monmouth, Matawan Formations undifferentiated**	30?
		Lower		144	**Potomac Group** **Patapsco Formation** **Arundel/Patuxent Formations (undivided)**	1000-1100 750-900
Proterozoic/ Paleozoic				543+	**Metamorphic/Igneous**	

FIGURE 2.5.1-34 **Rev. 0**

CCNPP SITE SPECIFIC STRATIGRAPHIC COLUMN

CCNPP UNIT 3 FSAR

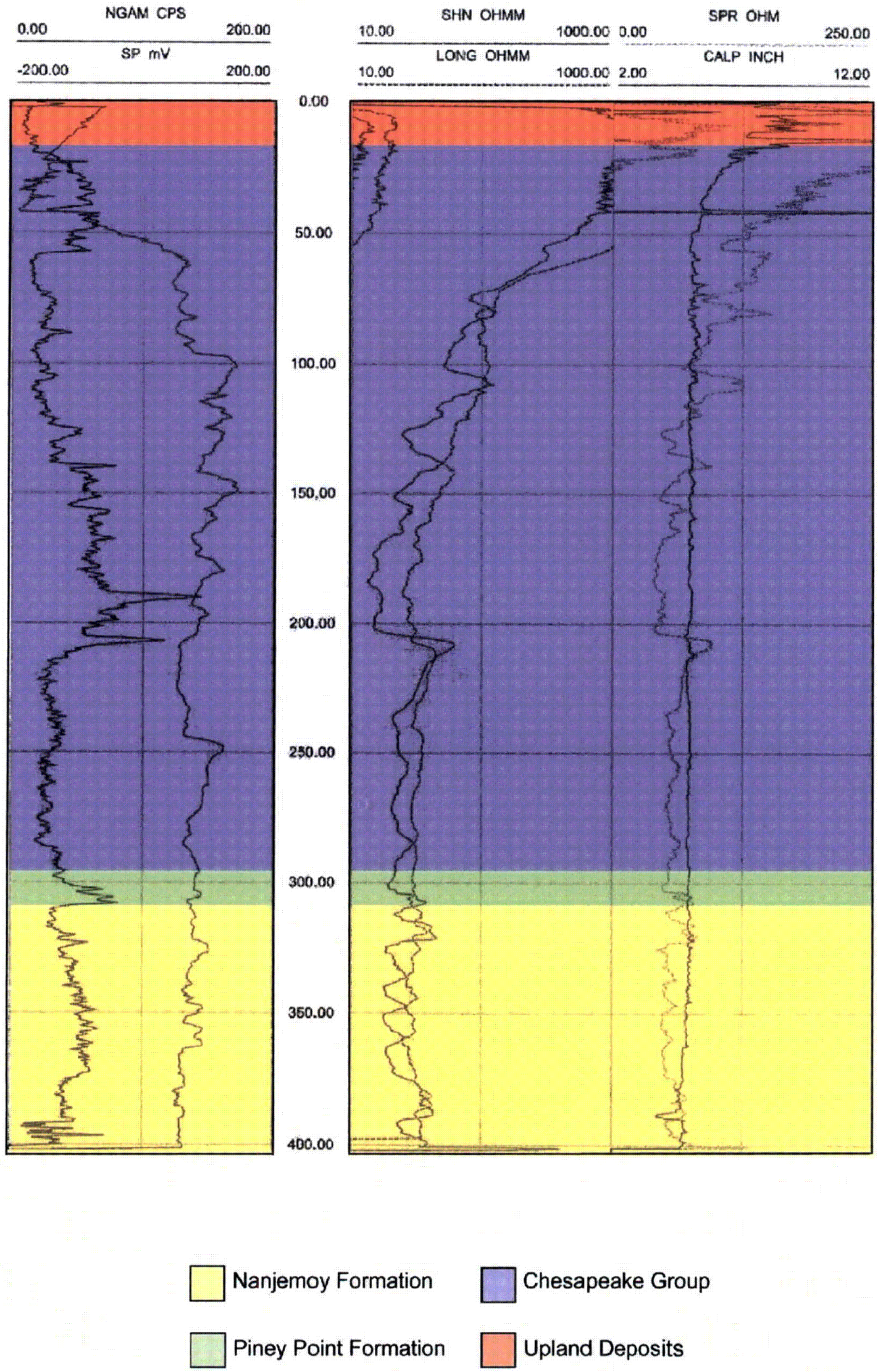


FIGURE 2.5.1-35 **Rev. 0**

BORING B-301, CALIPER, NATURAL GAMMA, RESISTIVITY AND SP LOGS

CCNPP UNIT 3 FSAR

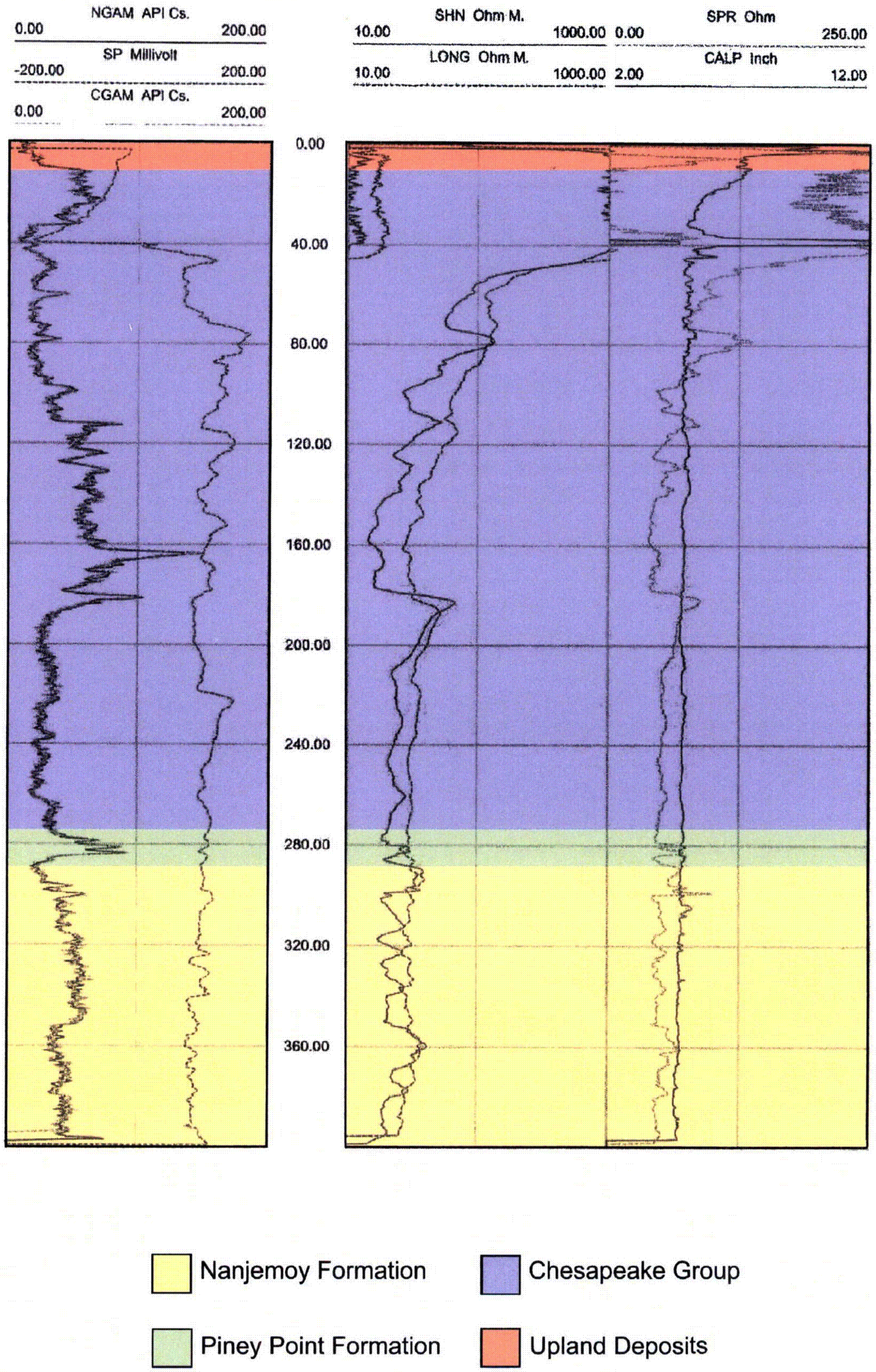


FIGURE 2.5.1-36 **Rev. 0**

BORING B-401, CALIPER, NATURAL GAMMA, RESISTIVITY AND SP LOGS

CCNPP UNIT 3 FSAR

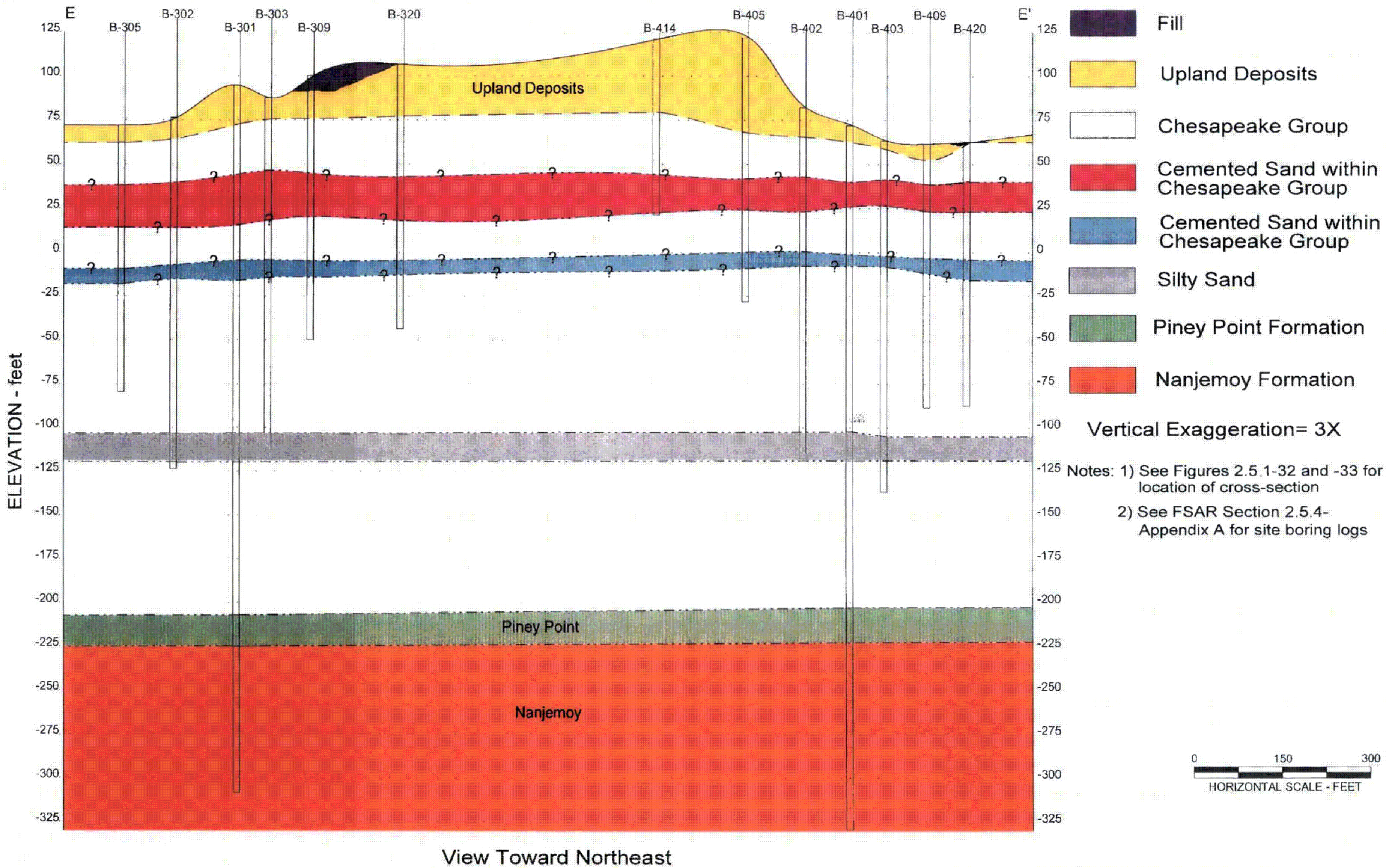
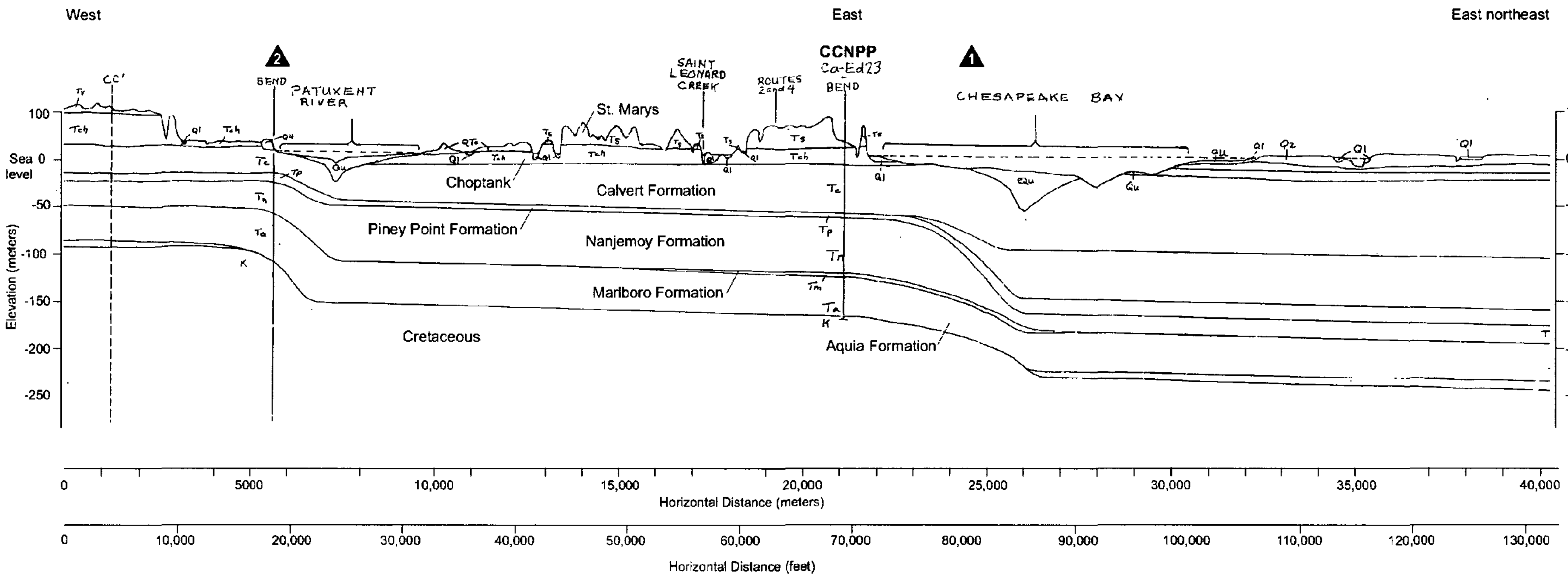


FIGURE 2.5.1-37 **Rev. 0**

SUBSURFACE PROFILE E–E'

CCNPP UNIT 3 FSAR



Explanation

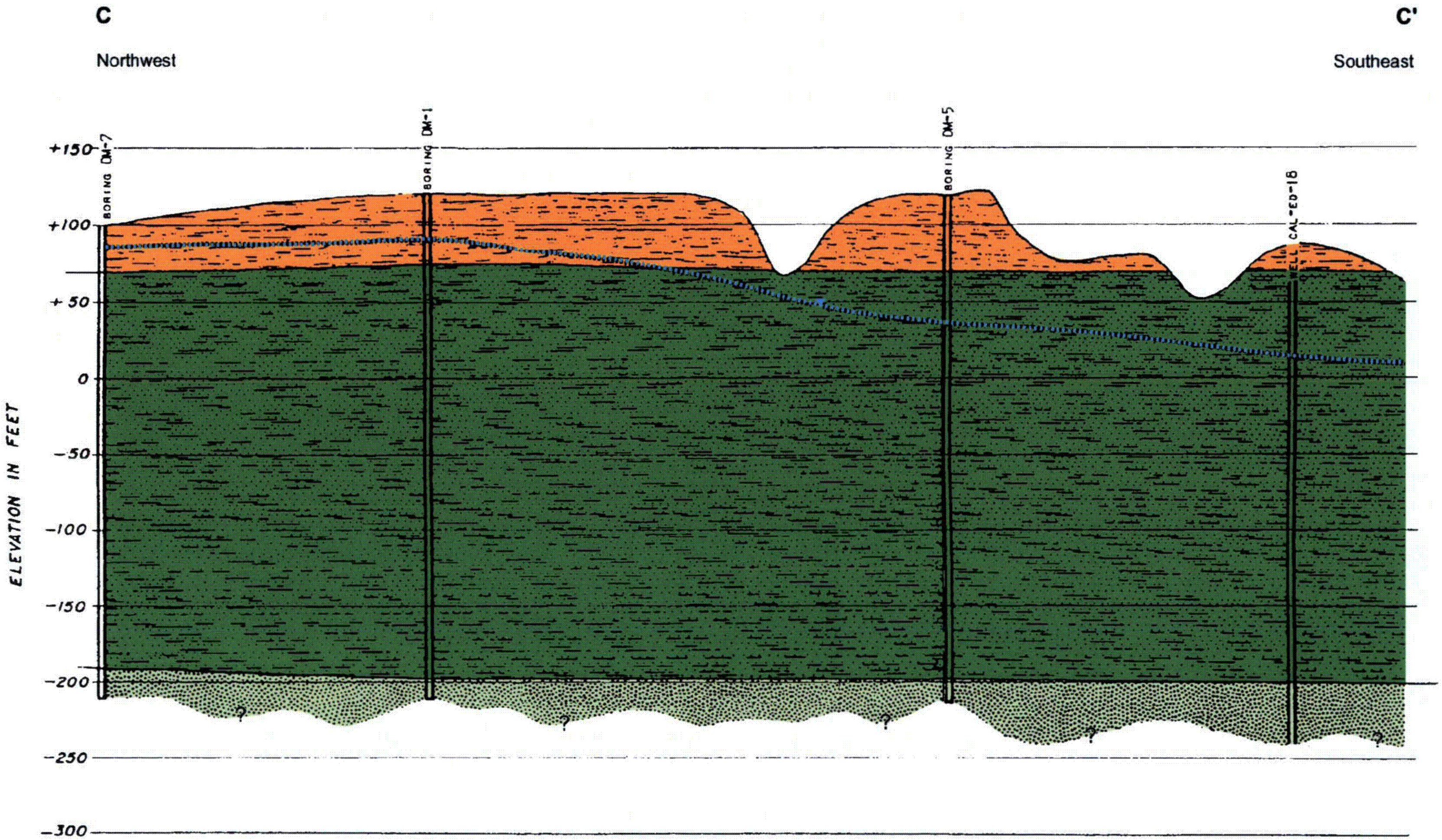
1 — Location of warp shown on Figure 2.5.1-24; vertical exaggeration = ~25X

Note: The above portion of section A - A' was modified from McCartan et al. (1995).

FIGURE 2.5.1-38 **Rev. 0**

McCARTAN et al. (1995) CROSS SECTION

CCNPP UNIT 3 FSAR



Explanation

- Pliocene Upland deposits - primarily silt and sand with some gravel
- Miocene Chesapeake Group - primarily sandy and clayey silt with some interbedded sand; shell layers in upper portion
- Eocene Piney Point Formation - glauconitic sand

0 500 1000 feet

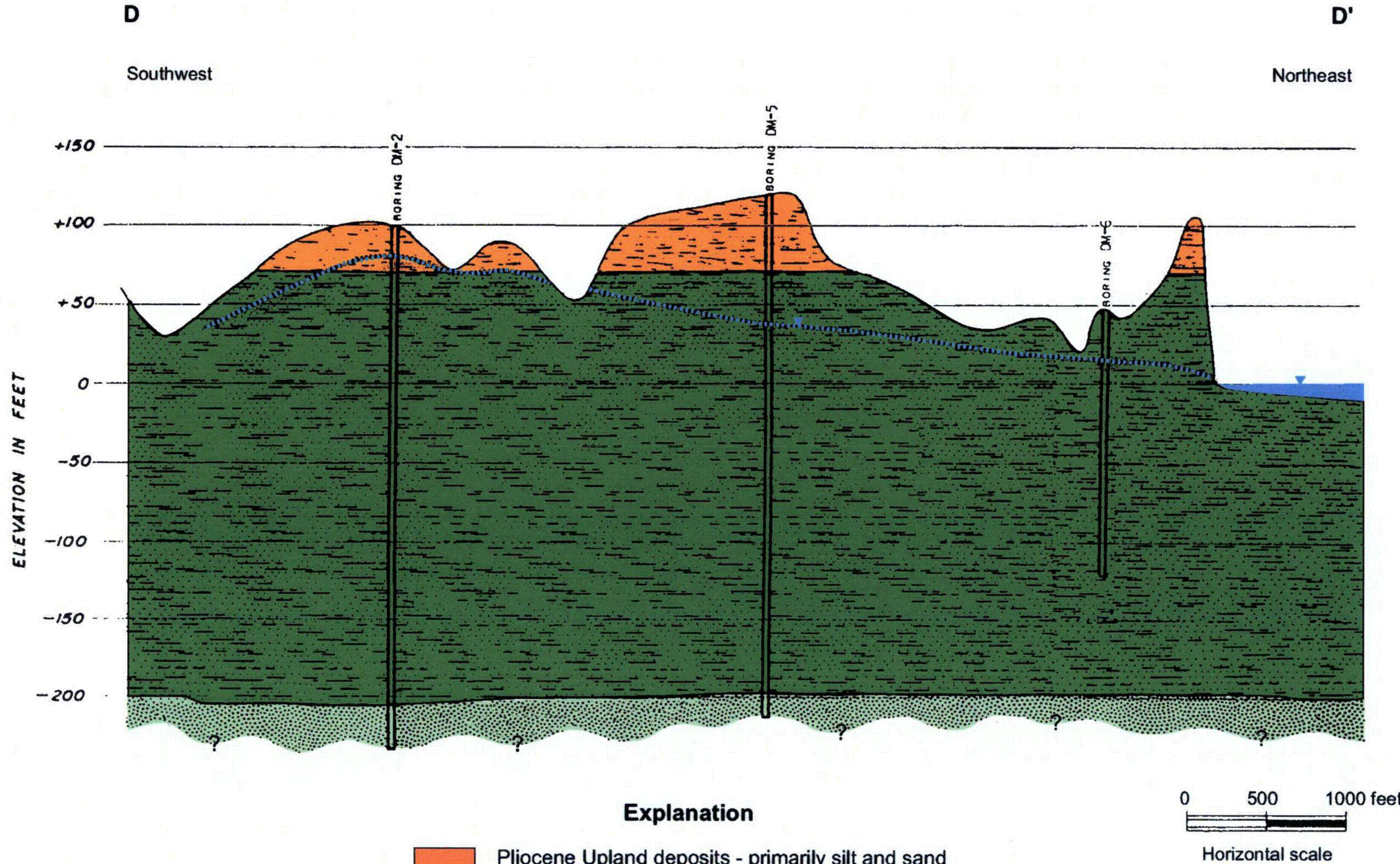
Horizontal scale

Note: Cross section originally A - A' from Dames & Moore.

FIGURE 2.5.1-39 **Rev. 0**

GEOLOGIC SECTION C – C'

CCNPP UNIT 3 FSAR



Explanation

(orange)	Pliocene Upland deposits - primarily silt and sand with some gravel
(dark green)	Miocene Chesapeake Group - primarily sandy and clayey silt with some interbedded sand; shell layers in upper portion
(light green)	Eocene Piney Point Formation - glauconitic sand

Note: Cross section originally B - B' from Dames & Moore.

FIGURE 2.5.1-40 Rev. 0

GEOLOGIC SECTION D–D'

CCNPP UNIT 3 FSAR

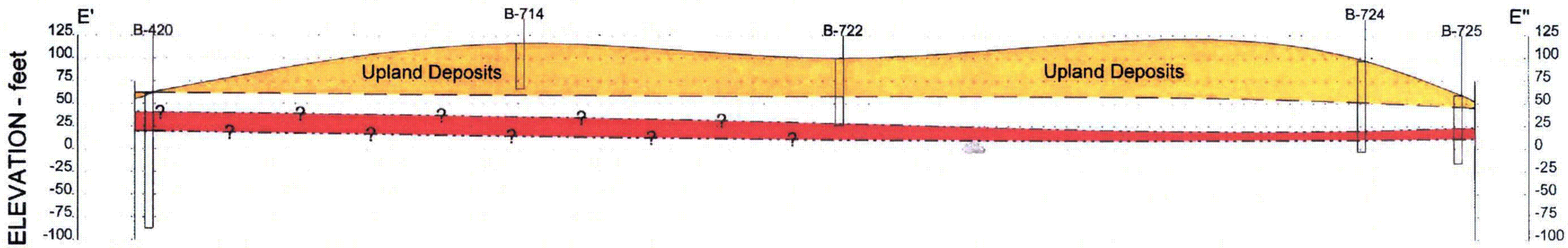


Figure 2.5.1-41

View Toward Northeast

- Fill
- Upland Deposits
- Chesapeake Group
- Cemented Sand within Chesapeake Group
- Cemented Sand within Chesapeake Group
- Silty Sand
- Piney Point Formation
- Nanjemoy Formation

Vertical Exaggeration= 5X

Notes: 1) See Figures 2.5.1-32 and -33 for location of cross-section

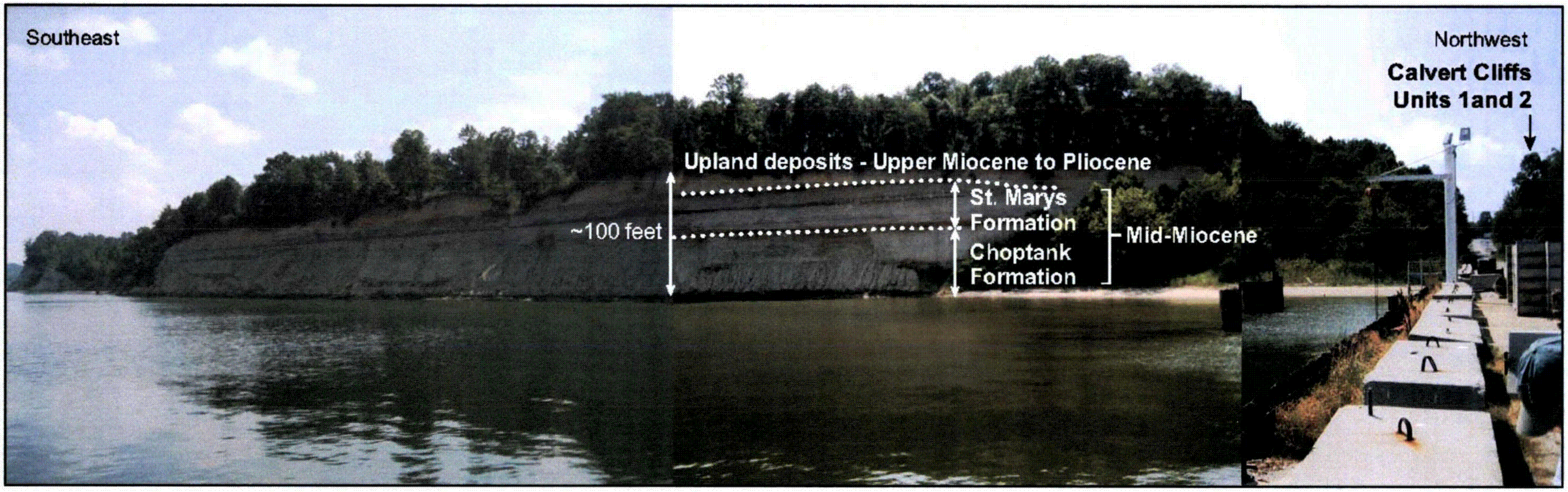
2) See FSAR Section 2.5.4-Appendix A for site boring logs

0 600 1200
HORIZONTAL SCALE - FEET

FIGURE 2.5.1-41 **Rev. 0**

SUBSURFACE PROFILE E–E'

CCNPP UNIT 3 FSAR



Note: St. Marys and Choptank Formation contact from Kidwell (1977).

FIGURE 2.5.1-42 **Rev. 0**

VIEW OF CALVERT CLIFFS
TOWARD THE SOUTHWEST
FROM THE BARGE

CCNPP UNIT 3 FSAR