2.5 Geology, Seismology, and Geotechnical Engineering

This section presents information on the geological, seismological, and geotechnical engineering properties of the VEGP site. Section 2.5.1 describes basic geological and seismologic data. 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 (SSE) ground motion. Section 2.5.3 describes the potential for surface faulting in the site area, and Sections 2.5.4, 2.5.5, and 2.5.6 describe the stability of subsurface materials and foundations at the site.

NRC Regulatory Guide 1.165, *Identification and Characterization of Seismic Sources and Determination of Safe Shutdown Earthquake Ground Motion* (RG 1.165) (1977), Appendix D, *Geological, Seismological and Geophysical Investigations to Characterize Seismic Sources*, provides guidance for the level of investigation recommended at different distances from a proposed site for a nuclear facility.

The following four terms for site map areas are designated by RG 1.165:

- Site region - area within 200 mi (320 km) of the site location.
- Site vicinity - area within 25 mi (40 km) of the site location.
- Site area - area within 5 mi (8 km) of the site location.
- Site - area within 0.6 mi (1 km) of the proposed VEGP Unit 3 and 4 locations.

These terms 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 this ESP application.

SNC conducted field investigations and performed extensive research of relevant geologic literature to reach the conclusion that no geologic or seismic hazards have the potential to affect the VEGP site except the Charleston seismic zone and a small magnitude local earthquake occurring in the site region. These topics are discussed in greater technical detail in Section 2.5.2. There is only limited potential for non-tectonic surface deformation in shallow deposits within the 5-mi site area radius, and this potential can be mitigated by means of excavation.

2.5.1 Basic Geologic and Seismic Information

This section presents information on the geological and seismological characteristics of the VEGP site region and site area. The information is divided into two parts. Section 2.5.1.1 describes the geologic and tectonic setting of the site region (200 mi), and Section 2.5.1.2 describes the geology and structural geology of the site area (5 mi). The geological and seismological information was developed in accordance with the guidance presented in NRC Regulatory Guide 1.70, *Standard Format and Content of Safety Analysis Reports for Nuclear Power Plants, LWR Edition* (RG 1.70) (1978), Section 2.5.1, *Basic Geologic and Seismic Information*, and RG 1.165 and is intended to satisfy the requirements of 10 CFR 100.23(c). The geological and seismological information presented in this section is used as a basis for evaluating the detailed geologic, seismic, and man-made hazards at the site.

RG 1.165 states that the vibratory design ground motion for a new nuclear power plant may be developed using either the Electric Power Research Institute (EPRI) or Lawrence Livermore National Laboratory (LLNL) probabilistic seismic hazard methodology. As described in Section 2.5.2, the EPRI methodology has been used to develop the SSE ground motion for the VEGP site. RG 1.165 further requires that the geological, seismological, and geophysical database be updated and any new data be evaluated to determine whether revisions to the 1986 EPRI seismic source model are required (presented in Section 2.5.2). This section, therefore, provides an update of the geological, seismological, and geophysical database for the VEGP site, focusing on whether any data published since the 1980s indicates a significant change to the 1986 EPRI seismic source model. In addition, the geotechnical properties of the VEGP site location are described to evaluate the ground motion site response characteristics of the site (i.e., non-tectonic geologic and man-made hazards at the site presented in Section 2.5.4).

The geological and seismological information presented in this section was developed from a review of previous reports prepared for the existing VEGP Units 1 and 2, published geologic literature, new boreholes drilled at the potential VEGP Units 3 and 4 site and a seismic refraction and reflection survey conducted in preparation of this ESP application. A review of published geologic literature was used to supplement and update the existing geological and seismological information. A list of the references used to compile the geological and seismological information presented in the following sections is provided at the end of each major subsection within Section 2.5.

2.5.1.1 Regional Geology (200 mi radius)

This section discusses the regional geology within a 200-mi radius of the VEGP site. The physiography and geomorphology, geologic setting and stratigraphy, and tectonic setting are discussed below. The information provided is a brief summary of this broad area, with an extensive and current bibliography. The information also provides the basis for evaluating the geologic and seismologic hazards discussed in the succeeding sections.

2.5.1.1.1 Regional Physiography and Geomorphology

Physiographically, from northwest to southeast the site region includes parts of the Valley and Ridge, Blue Ridge, Piedmont, and Coastal Plain provinces. The VEGP site is located on the upper Coastal Plain, about 30 mi (48 km) southeast of the Fall Line that separates the Piedmont and Coastal Plain provinces (Figure 2.5.1-1).

The Valley and Ridge Physiographic Province

The Valley and Ridge physiographic province extends from the 25-mi wide Hudson Valley in New York State to a wider 75-mi zone in Pensylvannia, Maryland, and Virginia and is about 50 mi wide from southern Virginia southward to Alabama. This physiographic province is underlain by a folded and faulted sequence of Paleozoic sedimentary rocks. The linear valleys and ridges typical of this province are the result of differential weathering and erosion in a humid environment of lithologies that are more or less resistant to these geomorphic processes.

The Blue Ridge Physiographic Province

The Blue Ridge physiographic province is located west of and adjacent to the Piedmont province. The Blue Ridge province extends from Pennsylvania to northern Georgia. It varies from about 30 to 75 mi (48 to 120 km) wide, north to south. Elevations are highest in North Carolina and Tennessee, with several peaks in North Carolina exceeding 5,900 ft (1,800 m) msl, including Mount Mitchell, North Carolina, the highest point (6,684 ft [2,037 m] msl) in the Appalachian Mountains. The Blue Ridge front, with a maximum elevation of about 3,950 ft (1,200 m) msl in North Carolina, is an east-facing escarpment between the Blue Ridge and Piedmont provinces in the southern Appalachians. The Blue Ridge province is comprised primarily of relatively more resistant granites and granitic gneisses that form a broad mountainous upland area in Georgia.

The Piedmont Physiographic Province

The Piedmont physiographic province extends southwest from New York to Alabama and lies west of and adjacent to the Atlantic Coastal Plain. It is the eastern-most physiographic and structural province of the Appalachian Mountains. The Piedmont is a seaward-sloping plateau whose width varies from about 10 mi (16 km) in southeastern New York to almost 125 mi (200 km) in North Carolina; it is the least rugged of the Appalachian provinces. Elevation of the inland boundary ranges from about 200 ft (60 m) msl in New Jersey to over 1,800 ft (550 m) msl in Georgia.

The Piedmont province is divided into the Piedmont Upland section to the west and the Piedmont Lowland section to the east. 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 Cambrian quartzites and Precambrian crystalline 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 (referred to as the Triassic Basin Rocks).

The Coastal Plain Physiographic Province

The Atlantic Coastal Plain extends southward from Cape Cod, Massachusetts, to south-central Georgia where it merges with the Gulf Coastal Plain. The surface of the Coastal Plain slopes gently seaward. The province is underlain by a seaward-dipping wedge of unconsolidated and semiconsolidated sediments that extend from the contact with the crystalline Piedmont province at the Fall Line to the edge of the continental shelf. In Georgia, the province is known as the Coastal Plain province. Sediment thickness increases from zero at the Fall Line to about 4,000 ft (1,200 m) at the Georgia - South Carolina coastline.

The VEGP site is located within a portion of the Coastal Plain in which depositional landforms have been obliterated by fluvial erosion based on studies in South Carolina **(Colquhoun and Johnson 1968)**. The lower coastal plain south of the site is dominated by primary depositional topography that has been modified slightly by fluvial erosion. In addition, the Coastal Plain contains numerous, minor geomorphic features, called Carolina bays. Since the origin of these minor geomorphic features has been controversial they are discussed in more detail in the following paragraphs. However, it is currently believed that these features formed as a result of erosion by predominantly southwesterly winds **(Soller and Mills 1991)**.

Carolina bays are shallow, elliptical depressions with associated sand rims that are found on the surface of the Coastal Plain sediments. These features, common throughout the Atlantic Coastal Plain, are most numerous in North and South Carolina, with major axes of the depressions up to 1.1 mi (1.8 km) long **(Siple 1967)**. The depressions are found from southern New Jersey to northern Florida, with the greatest occurrence in the Carolinas **(Soller and Mills 1991)**. One hundred ninety-four confirmed or suspected Carolina bays have been identified at the Savannah River Site (SRS), a US Department of Energy (DOE) reservation located directly east of the VEGP site area on the South Carolina side of the Savannah River. The long axes of the bays are oriented S50°E **(Johnson 1942)**, and the sand rims generally are observed on the east and southeast flanks. Various authors have provided a range of hypotheses for the timing and mode of origin of these bays [e.g., **(Gamble et al. 1977b; Prouty 1952; Cooke 1954; Melton and Schriever 1933; Kaczorowski 1976)**]. Theories regarding the origin of bays include meteorite impact, sinks, wind, and water currents [e.g., **(Melton and Schriever 1933; Prouty 1952; Cooke 1954; Thom 1970; Gamble et al. 1977b; Kaczorowski 1976)**. The origin of these features remains indeterminant. Prouty (1952), Kaczorowski (1976), and Savage (1982) proposed the most likely formation mechanism, that is, Carolina bays formed from eolian erosion by strong, unidirectional, southwesterly winds **(Soller and Mills 1991)**.

Savage (1982) and Kaczorowski (1976) provide the most likely explanation of formation, that is, the bays formed by action of strong unidirectional wind on water ponded in surface depressions **(Soller and Mills 1991)**. The resulting waves caused the formation of the sand rims as shoreline features, and the sand rims formed perpendicular to the wind direction. The wind bays observed today formed in response to a southwesterly wind **(Soller and Mills 1991)**.

The Carolina bays are surficial features that have no effect on the subsurface sediments. Based on subsurface core data, a clay layer mapped beneath the bays and outside their rims has no greater relief beneath the bays than beyond them **(Gamble et al. 1977b)**. Bryant and McCracken (1964), Preston and Brown (1964), and Thom (1970) provide additional evidence of the surficial character of Carolina bays. In these studies, certain identified strata were mapped and were found to be continuous and undeformed beneath bay and interbay areas. In Horry and Marion counties, South Carolina, there was no evidence of solution-related subsidence of the Carolina bays in spite of the presence of carbonate-rich strata in the subsurface and some localized sink holes of irregular shape with depths on the order of 20 ft. Early studies suggest that the bay-like depressions in the vicinity of the VEGP site (at SRS) probably resulted from dissolution of carbonate from the underlying geological formations **(Siple 1967)**. However, this has not been substantiated.

The age of the Carolina bays is based on Soller (1988) and Thom (1970). A minimum age has been set at middle to late Wisconsinian based on radiocarbon date **(Thom 1970)**. The maximum age can be constrained by examination of the formations on which the bays rest **(Soller and Mills 1991)**. This places bay formation between 100 thousand and 200 thousand years ago (ka) **(Soller and Mills 1991)**. If there is more than one generation, then the bays could be as old as the formations on which they rest.

2.5.1.1.2 Geologic History

The VEGP site is located in the Coastal Plain Physiographic province. Portions of all the major lithotectonic divisions of the Appalachian orogen (mountain belt) are found within a 200-mile (320-km) radius of the VEGP site. The structures and stratigraphic sequences within these divisions represent a complex geologic evolution that ends in the modern day passive margin of the Atlantic continental margin. Sections 2.5.1.1.3 and 2.5.1.1.4 provide additional detail.

Within the Appalachian orogen, several lithotectonic terranes that have been extensively documented include the foreland fold belt (Valley and Ridge) and western Blue Ridge

Precambrian–Paleozoic continental margin; the eastern Blue Ridge–Chauga Belt–Inner Piedmont Terrane; the volcanic-plutonic Carolina-Avalon Terrane; and the geophysically defined basement terrane beneath the Atlantic Coastal Plain [for an expanded bibliography see **(Hatcher et al. 1990, 1994, 2002, 2005)**]. These geological divisions contain the regional geologic record for the complete cycles of Precambrian and Paleozoic orogens and subsequent opening and closing of ocean basins (Proterozoic Iapetus Ocean and the opening of the Atlantic Ocean) **(Hatcher et al. 1994)**.

The late Proterozoic rifting is recorded in rift-related sediments at the edge of the frontal Blue Ridge province and the Ocoee and Tallulah Falls basins in the western and eastern Blue Ridge, respectively. Passive margin conditions began in the middle Cambrian and persisted through early Ordovician. The Cambro–Ordovician sedimentary section in the Valley and Ridge reflects this condition. The collision-accretionary phase of the Appalachians began in the middle Ordovician and persisted with pulses through the early Permian (Penobscot/Taconic, Acadian, Alleghanian orogenies). See Section 2.5.1.1.4 and Figure 2.5.1-10 for more detail.

The modern continental margin includes the Triassic basins that record the beginning of extension and continental rifting during the early to middle Mesozoic leading to the formation of the current Atlantic Ocean. One locus of major extension during early stages was in the South Georgia rift, which extends from Georgia into South Carolina. The Dunbarton Basin, underlying the VEGP site and SRS, is likely structurally related to the South Georgia rift basin **(Stieve and Stephenson 1995)** (Figure 2.5.1-2). See Section 2.5.1.1.3.4 for more details. During the later stage of rifting (early Jurassic), the focus of extension shifted eastward to the major marginal basins that would become the site of the Atlantic Ocean basin. The extension in the onshore, western-most basins, such as the Dunbarton, waned. Eventually, rifting of continental crust ceased as sea floor spreading began in the Atlantic spreading center sometime around 175 million years ago (Ma) **(Klitgord et al. 1988)**. The oldest ocean crust in contact with the eastern continental margin is late middle Jurassic **(Klitgord and Schouten 1986)**. The significance of the transition from rifting to sea floor spreading is that the tectonic regime of rifting is no longer acting on the continental crust along the Eastern Atlantic margin.

After the continental extension and rifting ended, a prograding shelf-slope began to form over the passive continental margin. The offshore Jurassic–Cretaceous clastic-carbonate bank sequence covered by younger Cretaceous and Tertiary marine sediments, and onshore Cenozoic sediments, represent a prograding shelf-slope **(Hatcher el al. 1994)** and the final evolution to a passive margin. Cretaceous and Cenozoic sediments thicken from near zero at the Fall Line to about 1,000 ft (335 m) in the center of the VEGP area, to approximately 4,000 ft (1,219 m) at the Georgia and South Carolina coast. The fluvial-to-marine sedimentary wedge consists of alternating sand and clay with tidal and shelf carbonates common in the downdip Tertiary section. Other offshore Continental margin elements include the Florida–Hatteras shelf and slope and the unusual Blake Plateau basin and Escarpment **(Dillon and Popenoe 1988; Klitgord et al. 1988; Poag and Valentine 1988)**.

2.5.1.1.3 Regional Stratigraphy and Geologic Setting

The regional stratigraphy within each of the physiographic provinces is presented below. The generalized geology and stratigraphy within a 200-mi radius of the VEGP site is shown on Figures 2.5.1-3 and 2.5.1-4. The stratigraphy shown on Figures 2.5.1-3 and 2.5.1-4 is from a portion of The Geologic Map of the United States **(King and Beikman 1974)**. The regional stratigraphy of the rock units shown on Figure 2.5.1-3 is illustrated by the legend (Figure 2.5.1-4). The rock units are classified based on age and type.

2.5.1.1.3.1 Valley and Ridge Province

The Valley and Ridge lithotectonic terrane contains Paleozoic sedimentary rock consisting of conglomerate, sandstone, shale, and limestone (Figures 2.5.1-1, 2.5.1-3 and 2.5.1-14). This continental shelf sequence was extensively folded and thrust faulted during the Alleghanian collisional event. The physiography is expressed as a series of parallel ridges and valleys. To the east is the boundary with the Blue Ridge province, at the Blue Ridge–Piedmont fault system. This boundary is fairly distinct in most places along the strike of the Appalachians and marks the change from folded rocks that are not penetratively deformed to rocks that are penetratively deformed.

2.5.1.1.3.2 Blue Ridge Province

The Blue Ridge lithotectonic province is bounded on the southeast by the Brevard fault zone and on the northwest by a predominantly thrust fault system **(Hatcher and Goldberg 1991; Hatcher and Butler 1979; King 1955)** (Figure 2.5.1-14). The province is a metamorphosed basement/cover sequence that has been complexly folded, faulted, penetratively deformed, and intruded. These rocks record multiple late Proterozoic to late Paleozoic deformation (extension and compression) associated with the formation of the Iapetus Ocean and the Appalachian orogen **(Hatcher and Goldberg 1991; Hopson 1989; Hatcher et al. 1986a; Nelson et al. 1985; Hatcher 1978)**. The province consists of a series of westward-vergent thrust sheets, each with different tectonic histories and different lithologies, including gneisses, plutons, and metavolcanic and metasedimentary rift sequences, as well as continental and platform deposits [see **(Hatcher and Goldberg 1991** and **Hatcher et al. 1994)**] for an expanded bibliography). The Blue Ridge–Piedmont fault system thrust the entire Blue Ridge province northwest over Paleozoic sedimentary rock of the Valley and Ridge province during the Alleghanian orogeny **(Hatcher 1971, 1972; Cook et al. 1979; Coruh et al. 1987)**. The Blue Ridge geologic province reaches its greatest width in the southern Appalachians.

The Blue Ridge is divided into a western and an eastern belt separated by the Hayesville-Gossan Lead fault. Thrust sheets in the western Blue Ridge consist of a rift-facies sequence of clastic sedimentary rocks deposited on continental basement, whereas thrust sheets in the eastern Blue Ridge consist of slope and rise sequences deposited in part on continental basement and on oceanic crust **(Hatcher and Goldberg 1991; Hatcher 1978)**. Western Blue Ridge stratigraphy consists of basement gneisses and metasedimentary, metaplutonic, and metavolcanic rocks, whereas Eastern Blue Ridge stratigraphy consists of fewer lithologies, more abundant mafic rocks, and minor amounts of continental basement.

Western Blue Ridge

The western Blue Ridge consists of an assemblage of Middle Proterozoic crystalline continental (Grenville) basement rock nonconformably overlain by Late Proterozoic to Early Paleozoic rift and drift facies sedimentary rock **(Hatcher et al. 1994)**. The basement consists of various types of gniesses, amphibolite, gabbroic and volcanic rock, and metasedimentary rock. All basement rock is metamorphosed to granulite or uppermost amphibolite facies **(Hatcher et al. 1994)**. The calculated ages of these rocks generally range from 1,000 to 1,200 million years old (Ma) [e.g., **Fullagar et al. 1979; Fullagar and Bartholomew 1983; Fullagar and Odom 1973)**].

The rifting event during the Late Proterozoic through Early Paleozoic that formed the Iapetus Ocean is recorded in the rift-drift sequence of the Ocoee Supergroup and Chillhowie Group [e.g., **Wehr and Glover 1985; Rast and Kohles 1986; King et al. 1968; Neuman and Nelson 1965; King 1964)**]. These rocks, basement, and sedimentary cover were later affected by Taconic and possibly Acadian deformation and metamorphism. The entire composite thrust sheet was transported west as an intact package during the Alleghanian collision event on the Blue Ridge–Piedmont thrust.

Eastern Blue Ridge

The eastern Blue Ridge is located southeast of the western Blue Ridge and is separated from that province by the Hayesville–Gossen Lead fault. The Brevard fault zone forms the southeastern boundary with the Inner Piedmont. Lithologically, the eastern Blue Ridge is composed of metasedimentary rocks originally deposited on a continental slope and rise and ocean floor metasedimentary rocks in association with oceanic or transitional to oceanic crust [for expanded bibliography see **(Hatcher et al. 1994; Hopson et al. 1989)**]. This is in contrast to the previously discussed western Blue Ridge that contains metasedimentary rocks suggesting continental rift-drift facies of a paleomargin setting. The eastern Blue Ridge is structurally complex, with several major thrust faults, multiple fold generations, and two high-grade metamorphic episodes **(Hatcher et al. 1994)**. Metamorphism took place during the Taconic and possibly Acadian orogenies.

The stratigraphy within the eastern Blue Ridge includes rare Grenville (Precambrian) gneisses, metasedimentary rocks of the Tallulah Falls Formation and the Coweeta Group, metamorphosed Paleozoic granitoids, and mafic and ultramafic complexes and rocks of the Dahlonega Gold Belt. The Paleozoic granitoids are a part of a suite of similar granites found in the western Inner Piedmont, suggesting a common intrusive history. Metasedimentary rock sequences in the eastern Blue Ridge are correlated along strike as well as across some thrust fault boundaries, also suggesting a commonality in the original depositional history. Based on geochemical data, the mafic and ultramafic complexes found in particular thrust sheets in the eastern Blue Ridge have oceanic as well as continental affinities. However, their exact tectonic origin is not clear because the contacts with the host metasedimentary rock are obscure.

2.5.1.1.3.3 Piedmont Province

The Piedmont province in northwestern Georgia and South Carolina consists of variably deformed and metamorphosed igneous and sedimentary rocks ranging in age from Middle Proterozoic to Permian (1,100–265 Ma). The province consists of the Western (Inner) Piedmont and the Carolina-Avalon Terrane (Figure 2.5.1-14). This designation is made because of different tectonic origins for the western and eastern parts of the province. The province can also be subdivided into seven distinctive tectonostratigraphic belts separated by major faults, contrasts in metamorphic grade, or both. The Charlotte and Carolina Slate belts are combined and discussed as the Carolina-Avalon Terrane. The rocks of the Piedmont have been deformed into isoclinal recumbent and upright folds, which have been refolded and are contained in several thrust sheets or nappes. These metamorphic rocks extend beneath the Coastal Plain sediments in Georgia and South Carolina. The southeastern extent of the Piedmont province underneath the Coastal Plain is unknown.

Western Piedmont

The Western Piedmont encompasses the Inner Piedmont block, the Smith River Allochthon in Virginia and North Carolina, and the Sauratown Mountains Anticlinorium in north central North Carolina **(Horton and McConnell 1991)** (Figure 2.5.1-5). The Western (Inner) Piedmont is separated from the Blue Ridge province on the northwest by the Brevard Fault zone. It is separated from the Carolina–Avalon Terrane on the southeast by the Towaliga fault, a complex series of fault zones approximately coincident with the Central Piedmont suture **(Hatcher 1987)**. These faults include: Lowndesville, Kings Mountain, Eufola, Shacktown, and Chatham Fault zones **(Horton and McConnell 1991)**. The province is a composite stack of thrust sheets containing a variety of gneisses, schists, amphibolite, sparse ultramafic bodies, and intrusive granitoids **(Nelson et al. 1987; Goldsmith et al. 1988; Hatcher and Butler 1979)**. The protoliths are immature quartzo-feldspathic sandstone, pelitic sediments, and mafic lavas.

The Sauratown Mountains anticlinorium is a complex structural window of four stacked thrust sheets that have been exposed in eroded structural domes. Each sheet contains Precambrian basement with an overlying sequence of younger Precambrian-to-Cambrian metasedimentary and metaigneous rocks **(Horton and McConnell 1991)**. The Smith River Allochthon contains two predominantly metasedimentary units and a suite of plutonic rocks. It is a completely fault-bounded terrane, as is the Sauratown Mountains anticlinorium. The Inner Piedmont block is a fault-bounded, composite thrust sheet with metamorphic complexes of different tectonic affinities **(Horton and McConnell 1991)**. There is some continental basement within the block **(Goldsmith et al. 1988)** and scattered mafic and ultramafic bodies and complexes **(Mittwede et al. 1987)**, suggesting the presence of oceanic crustal material **(Horton and McConnell 1991)**. The rest of the block contains a coherent, though poorly understood, stratigraphy of metasedimentary rock, metavolcanic gneisses, and schists **(Horton and McConnell 1991)**. The eastern Blue Ridge and Inner Piedmont contain some stratigraphically equivalent rocks **(Hatcher et al. 1986b)**.

The stratigraphy and structural geologic data in the western Piedmont reflects the effects of a complex tectonic history from the Precambrian Grenville through Late Paleozoic Alleghanian orogenies. Metamorphism affected the basement rocks of the Sauratown Mountains anticlinorium at least twice: during the Precambrian Grenville and later during the Paleozoic. The metasedimentary cover sequence as well as the Smith River allochthon and the Inner Piedmont block were affected by one metamorphic event (prograde and retrograde) in the Paleozoic **(Horton and McConnell 1991)**. The Alleghanian continental collision is reflected in the thrust and dextral strike slip fault systems such as the Brevard and Bowens Creek fault zones. A few late Paleozoic granites were emplaced in the Inner Piedmont block; however, most lie further east in the Carolina Terrane. Early Mesozoic extension resulted in the formation of rift basins (Dan River and Davie County basins).

Carolina-Avalon Terrane

The Carolina Terrane is part of a late Precambrian–Cambrian composite arc terrane exotic to North America **(Secor et al. 1983; Samson et al. 1990)** and accreted sometime during the Ordovician to Devonian **(Vick et al. 1987; Noel et al. 1988)**. It consists of felsic to mafic volcanic rock and associated volcanoclastic rock. Middle Cambrian fossil fauna indicate a European or African affinity **(Secor et al. 1983)**.

The northwestern boundary of the Carolina–Avalon Terrane is formed by a complex of faults that constitute the Central Piedmont suture (Figures 2.5.1-5 and 2.5.1-6) and separate the terrane from rocks of North American affinity [see **(Hatcher et al. 1988; Hooper and Hatcher 1988; Steltenpohl 1988; Griffin 1979; Davis 1980; Rozen 1981; Horton 1981; Heyn 1988; McConnell 1988)**]. This structure was reactivated during the later Alleghanian collisional events as a dextral shear fault system **(Bobyarchick 1981)**. Subsequent investigators have further enhanced the geological knowledge of the complicated structure **(Secor et al. 1986a, 1986b; Hatcher and Edelman 1987; Hooper and Hatcher 1988)**. Dennis (1991) suggested that the Central Piedmont suture is a low-angle normal fault. The Carolina-Avalon Terrane (also known as the Carolina Terrane) is bounded on the southeast by the Modoc fault zone and the Kiokee belt (Figure 2.5.1-6).

The Carolina–Avalon Terrane is the combination of the earlier Charlotte and Carolina slate belts. The belts were initially distinguished by metamorphic grade **(King 1955)** and were later recognized as the same protolith and thus were combined **(Hatcher et al. 1994)**. Metamorphic grade increases to the northwest from lower greenschist facies to upper amphibolite facies. Pre-Alleghanian structure is dominated by large northeast trending folds with steeply dipping axial surfaces. All country rock of the Carolina-Avalon Terrane has been penetratively deformed, thereby producing axial plane cleavage and foliation **(Hatcher et al. 1994)**.

The Charlotte belt contains numerous intrusions and moderate- to high-grade metamorphic rock. Much of the belt was metamorphosed to amphibolite grade during the Taconic orogeny **(Butler 1979)**, but retrograde metamorphism is also widespread. The oldest rocks are amphibolite, biotite gneiss, hornblende gneiss, and schist, and probably were derived from volcanic, volcanoclastic, or sedimentary protoliths.

The Carolina Slate belt (Figure 2.5.1-6) is characterized by thick sequences of metasedimentary rocks derived from volcanic source areas and felsic to mafic metavolcanic rocks. The oldest rocks within the Carolina Slate belt consist of intermediate to felsic ashflow tuff and associated volcanoclastic rocks. These rocks are overlain by a sequence of mudstone, siltstone, sandstone, greywacke, and greenstone, with some interbedded volcanic tuff and flows. The belt was subjected to low- to medium-grade regional metamorphism and folding from 500 to 300 Ma and was intruded subsequently by granitic and gabbroic plutons about 300 Ma.

<u>Kiokee Belt</u>

The Kiokee belt is located between the Carolina-Avalon Terrane and the Belair belt in Georgia and South Carolina (Figure 2.5.1-6). It is referred to as the Savannah River Terrane in some of the recent literature **(Maher et al. 1991)**. The Kiokee belt is bounded on the northwest by the Modoc fault zone and on the southeast by the Augusta fault (the Modoc fault zone and the Augusta fault are discussed in Section 2.5.1.1.4.3). The Kiokee belt is a medium- to high-grade metamorphic belt with associated plutonism **(Daniels 1974)**. Snoke et al. (1980) concluded that the Kiokee belt was part of the Alleghanian metamorphic core. That core was deformed and metamorphosed prior to the development of the plastic shear zones bounding it **(Secor et al.**

1986a). The bounding faults are mylonite zones that overprint the amphibolite facies infrastructure of the core of the belt **(Hatcher et al. 1994)**.

The Kiokee belt is an antiformal structure that strikes northeast. The interior is a migmatitic complex of biotite amphibole paragneiss, leucocratic paragneiss, sillimanite schist, amphibolite, ultramafic schist, serpentinite, feldspathic metaquartzite, and granitic intrusions of Late Paleozoic age **(Secor 1987)**. Some of the lithologic units found in the Carolina slate belt may occur at higher metamorphic grade in the Kiokee belt **(Hatcher et al. 1994)**.

From extensive field studies and geochronological dating, a complex Alleghanian history can be derived from the studies of the Kiokee belt **(Bramlett 1989; Maher 1979; Snoke and Frost 1990; Dallmeyer et al. 1986; Fullagar and Butler 1979; Snoke et al. 1980; Harrison and McDougall 1980; Sacks et al. 1987)**. The pre-Alleghanian structure and stratigraphy are only partially known. The nature of the crustal rock that played a part in the metamorphism, deformation, and intrusion is still unknown. The possible role of a Precambrian basement in the Kiokee belt is a key question proposed by Hatcher et al. (1994). No rock in the Kiokee belt has been identified at this time as Precambrian basement. However, Long and Chapman (1977), suggested, based on gravity data, that a large rifted block of continental crust underlies the Kiokee belt.

<u>Belair Belt</u>

The Belair belt, also known as the Augusta Terrane **(Maher 1979, 1987)**, is locally exposed in the Savannah River valley, near Augusta, Georgia (Figure 2.5.1-6). It is largely concealed beneath the Atlantic Coastal Plain, but is exposed in several small erosional windows through the Coastal Plain sediments in eastern Georgia **(Bramlett et al. 1982)**. The Belair belt consists of intermediate to felsic volcanic tuffs and related volcanoclastic sediments penetratively deformed and metamorphosed to greenschist facies **(Bramlett et al. 1982; Maher 1978, 1979, 1987, Hatcher et al. 1977; Prowell and O'Connor 1978; Prowell 1988; O'Connor and Prowell 1976)**. The Belair belt displays similar characteristics as the Carolina-Avalon Terrane **(Maher et al. 1994)**. Geophysical and well data indicate that the Belair belt extends beneath the Atlantic Coastal Plain **(Daniels 1974)**. Near Augusta, the Augusta fault and the southeast edge of the Kiokee belt appear to be offset by the north-northeast trending Cenozoic Belair fault (Figure 2.5.1-6) (discussed in Section 2.5.1.1.4.3).

2.5.1.1.3.4 Mesozoic Rift Basins

While primarily exposed in the Piedmont Province, Mesozoic-age rift basins are found along the entire eastern continental margin of North America from the Gulf Coast through Nova Scotia (Figure 2.5.1-7). The basins formed in response to the continental rifting episode that broke up the supercontinent, Pangea, and led to the formation of the Atlantic Ocean basin. Rift basins are exposed in the Piedmont province as well as buried beneath Cretaceous and younger Coastal Plain sediments. Many underlie offshore regions. Structurally, the basins are grabens or half grabens elongated in a northeast direction and bounded by normal faults on one or both sides **(Manspeizer et al. 1978)**. Several basins were localized along reactivated Paleozoic ductile or brittle fault zones **(Petersen et al. 1984; Hutchinson and Klitgord 1986; Ratcliffe 1971; Lindholm 1978; Glover et al. 1980)**.

The basins are located in extended or rifted continental crust, identified as the Eastern Seaboard domain (subdomain of North American stable continental crust) **(Kanter 1994)**. Rifted crust is crust that has been stretched, faulted, and thinned slightly by rifting but is still recognizable as continental crust. The faulting is extensional or normal, and down-dropped blocks form rift basins. The western boundary of this zone of extended crust is defined by the western-most edge of Triassic–Jurassic onshore rift basins or the boundaries of the structural blocks in which they occur **(Klitgord et al. 1988; Keen and Haworth 1985; Kanter 1994)**. The eastern boundary is the continental shelf **(Grow et al. 1988)**.

Two belts of basins occur in the Eastern Seaboard domain, from the Carolinas to Pennsylvania **(Olsen et al. 1991)**: the eastern and the western belts. In North and South Carolina, the Deep River, Elberbe, and Crowburg basins are in the eastern belt, and the Dan River and Davie County basins are in the western belt **(Olsen et al. 1991)**. In addition, the Dunbarton, Florence, Riddleville, and South Georgia basins are buried beneath Coastal Plain sediments in the eastern belt (Figure 2.5.1-7).

The basins are generally filled with lacustrine sedimentary and igneous rock. Sedimentary strata consist mainly of non-marine sandstone, conglomerate, siltstone, and shale. Carbonate rocks and coal are found locally in several basins. Igneous rocks of basaltic composition occur as flows, sills, and stocks within the basins and as extensive dike swarms within and outside the basins **(King 1971)**. These basin fill strata have been described and named the Newark Supergroup **(Olsen 1978; Froelich and Olsen 1984; Olsen et al. 1991)**. In general, the stratigraphy can be broken out into three sections. The lower section is characteristically fluvial **(Smoot 1985; Gore 1986)** and contains reddish-brown, arkosic, coarse-grained sandstone and conglomerate. The middle section mainly includes sediments of lacustrine origin **(Smoot 1985)**. These sediments include grey-black fossiliferous siltstone, carbonaceous shale, and thin coal beds **(Olsen et al. 1991)**. The upper section is a complex of deltaic, fluvial, and lacustrine environments **(Olsen and Schlische 1988; Schlische and Olsen 1990)**. These sediments include red-brown siltstone, arkosic sandstone, pebble sandstone, red and grey mudstone, and conglomerate **(Olsen et al. 1991)**.

In Georgia and South Carolina, the Dunbarton Triassic rift basin, a sub-basin within the South Georgia Basin, is located beneath Coastal Plain sediments below the SRS and the VEGP sites

(Figure 2.5.1-7). The Dunbarton Basin was first identified based on aeromagnetic and well data **(Siple 1967)**. Subsequent seismic reflection surveys, potential field surveys, and additional well data have lead to the current understanding of the basin **(Marine 1974a, 1974b; Marine and Siple 1974; Chapman and DiStefano 1989; Anderson 1990; Stephenson and Stieve 1992; Luetgert et al. 1994; Domoracki 1995)**. The structure is currently interpreted as an asymmetric graben approximately 31 mi long and 6 to 9 mi wide. The axis of the basin strikes northeast, parallel to the regional strike of crystalline basement **(Marine and Siple 1974)**. The basin extends 5 mi southwest of the Savannah River and 25 mi to the northeast of the SRS, where it terminates against a granite body interpreted from magnetic data **(Siple 1967; Marine 1974a, 1974b; Anderson 1990)**. The primary fault controlling basin formation, the Pen Branch fault, bounds the northwest side of the basin. The fault appears to have been an earlier Paleozoic reverse fault that was reactivated as an extensional normal fault during Mesozoic continental rifting. The fault was subsequently reactivated in the Cenozoic as a reverse fault or right-oblique slip fault **(Price et al. 1989; Snipes et al. 1993a; Stieve and Stephenson 1995)**. The Pen Branch fault dips to the southeast. The master fault to the Riddleville Basin in Georgia also dips to the southeast **(Peterson et al. 1984)**. For further discussion of the Pen Branch fault, see Section 2.5.1.2.4.1. The southeast boundary of the basin is poorly constrained but is interpreted as fault bounded **(Faye and Prowell 1982; Snipes et al. 1993b)**.

Fourteen wells drilled at the SRS penetrated sedimentary rocks of the Dunbarton Basin. Recovered core is Mesozoic clastic rock **(Parsons Brinckerhoff 1973; Marine and Siple 1974)**. Conglomerate, fanglomerate, sandstone, siltstone, and mudstone are the dominant lithologies. These rocks are similar to the clastic facies in other Newark Supergroup basins. Conglomerate and red clayey siltstone are the dominant lithologies in those cores. Parsons Brinckerhoff (1973) concluded that the lithology and stratigraphy identified in the core indicate that the proximal side of the basin is to the northwest. A larger component of coarse-grained rock types occurs to the northwest than on the southeast side of the basin. This suggests an asymmetric basin with greater cumulative separation on the northwest than to the southeast. This asymmetry led to greater local relief along the northern boundary, where high energy fluvial processes dominated, and the resulting sediments are coarser grained than farther out in the basin.

Gravity and magnetic modeling suggests that the Triassic section in the Dunbarton Basin is at least 1.2 mi (2 km) thick. Boreholes have encountered up to 899 m of Triassic fill, but the base of the Dunbarton was not encountered **(Marine and Siple 1974)**. Seismic reflection data do not unequivocally constrain the base of the basin; the transition between the Triassic rock and the crystalline terrane is unclear. However, interpreted Triassic reflectors are at least as deep as 1 to 2.3 mi (1.6 to 3.7 km) **(Domoracki 1995)**.

The South Georgia Basin, further east and south in Georgia and South Carolina, is a much larger, deeper, and more complex basin than the Dunbarton Basin. The basin is as wide as 62 mi and as deep as 4 mi **(McBride 1991)**. It is not a single basin, but a complex of isolated synrift grabens with limited major crustal extension. The major border fault dips northward **(McBride 1991)**, as opposed to southeastward, for the controlling faults bounding the Dunbarton Basin.

Further to the northeast, in North Carolina, two major basins are exposed: the Dan River and Deep River basins. Both basins exhibit half-graben geometry, bounded on one side by a major normal fault zone. However, the border faults on the two basins are on opposite flanks of the basin: Dan River's Chatham fault zone dips to the southeast and Deep River's Jonesboro fault zone dips northwest **(Olsen et al. 1991)**.

2.5.1.1.3.5 Coastal Plain

In the Coastal Plain, rocks of Paleozoic and Triassic age have been beveled by erosion and are unconformably overlain by unconsolidated to poorly consolidated Coastal Plain sediments **(Cooke 1936; Siple 1967; Colquhoun and Johnson 1968)**. The sediments of the Coastal Plain in Georgia and South Carolina are stratified sand, clay, limestone, and gravel that dip gently seaward and range in age from Late Cretaceous to Recent. The sedimentary sequence thickens from essentially zero at the Fall Line to more than 4,000 ft (1,219 m) at the coast **(Colquhoun et al. 1983)**. Regional dip is to the southeast, although beds dip and thicken locally in other directions because of locally variable depositional regimes and differential subsidence of basement features.

Many investigations have provided insight into the evolution of the southeastern United States Coastal Plain, including: Cooke (1936), Siple (1967), Huddleston and Hetrick (1978), Colquhoun and Steele (1985), Prowell et al. (1985a, 1985b), Dennehy et al. (1988), Fallaw et al. (1990a, 1990b, 1992, 1995), Nystrom et al. (1990), Aadland and Bledsoe (1990), Huddleston and Summerour (1996) and, most recently, Falls and Prowell (2001). The Coastal Plain section is divided into several rock-stratigraphic groups based principally on age and lithology (Figure 2.5.1-8).

The proposed VEGP Units 3 and 4 site is located on the Coastal Plain. Numerous geologic studies have been conducted in the surrounding area since initial studies were conducted for the existing VEGP units. Most of these studies were conducted in the vicinity of the SRS and focused on correlating both geologic and hydrogeologic formations present in Georgia and South Carolina resulting in an updated stratigraphic nomenclature. A correlation chart showing the Vogtle FSAR, current USGS, Georgia Geological Survey, South Carolina and SRS nomenclature is provided in Figure 2.5.1-8. The following sections describe each geologic unit (from oldest to youngest), largely taken from the recent work of the USGS **(Falls and Prowell 2001)** and the Georgia Geological Survey **(Huddlestun and Summerour 1996)**.

Cretaceous Sediments

Five deep test holes were drilled in the Georgia Coastal Plain sediments of Burke and Screven counties by the USGS and the Georgia Geologic Survey to determine the stratigraphy of the Upper Cretaceous and Tertiary sediments in eastern Georgia near the Savannah River (Figure 2.5.1-9) **(Falls and Prowell 2001)**. The Cretaceous sections in the cores are divided into the Cape Fear Formation, the Pio Nono/Middendorf Formation, the Gaillard/Black Creek Formation (regionally, the Black Creek Group), and the Steel Creek Formation (Figure 2.5.1-8). These units consist of siliciclastic sediments. Evidence for possible unconformities led the investigators to recognize two subunits in the Middendorf and three subunits in the Black Creek Group. Each contact between units is a regional unconformity and denotes considerable hiatus in sedimentation. The depositional environment for all four units is interpreted as a set of large deltaic systems that prograded across the continental shelf in east-central Georgia and western South Carolina.

Cape Fear Formation

The Cape Fear Formation consists of partially lithified to unlithified, poorly to very poorly sorted clayey sand and sandy clay with a few beds of silty clay. The sand is fine to very coarse with granules and pebbles and is predominantly angular to subangular quartz with some feldspar. Cristobalite in the clay matrix results in lithologies that are harder and denser than sediments in the other Cretaceous units. The cristobalitic clay matrix imparts a yellowish-green to greenish-gray color to most of the lithologies and occludes most of the intergranular porosity in the sand beds. In three out of the five test holes evaluated by the USGS, the bottom of the Cape Fear Formation was drilled through and the formation ranges in total thickness from 96 to 212 ft, thickening in the downdip direction.

The Cape Fear Formation contains multiple fining-upward cycles that range in thickness from 3 to 15 ft. Each cycle grades upward from a basal pebbly coarse sand to a clayey sand or clay. The clays are oxidized and are generally stained with reddish-brown and yellowish-brown patches of iron oxide. A root-trace pattern is present at the top of a few of the fining-upward cycles. The upper contact sediments are typically stained with iron oxides.

Most of the strata in this unit are without fossils; however, a silty clay sample from one of the cores contains low-abundance and low-diversity pollen assemblages. Palynologic analysis indicates a Coniacian microflora **(Frederiksen et al. 2001)**. This is consistent with the microflora of the Cape Fear Formation of South Carolina and North Carolina **(Christopher et al. 1979; Christopher 1982; Sohl and Owens 1991)**. Prowell et al. (1985a) and Fallaw and Price

(1995) suggested a Santonian age for unit UK1 and the Cape Fear Formation at the SRS. Huddleston and Summerour (1996) suggested that the Cape Fear Formation is equivalent to the Cenomanian–Turonian Tuscaloosa Formation of western Georgia.

The presence of a terrestrial microflora and the absence of dinoflagellates (marine fossils) in the Cape Fear Formation suggest deposition in a nonmarine environment. The multiple fining-upward cycles, the coarse texture of the sand layers, the iron-oxide staining, and the root-trace patterns in the clay layers suggest that most of this unit was deposited in channel and overbank environment during aggradation of a fluvially dominated, subaerially exposed part of a delta-plain environment.

Pio Nono/Middendorf Formation

The Pio Nono/Middendorf Formation unconformably overlies the Cape Fear Formation with a distinct contact. The formation is marked by an abrupt change from the moderately indurated clay and clayey sand of the underlying Cape Fear to the slightly indurated sand and lesser clayey sand of the Pio Nono/Middendorf. The Pio Nono/Middendorf sand units are moderately to poorly sorted and are very porous and permeable in comparison to the sand units in the underlying Cape Fear. The Pio Nono/Middendorf consists predominantly of unlithified sand, which is locally fine to very coarse or fine to medium quartz. The mineral assemblage includes smoky-quartz granules and pebbles, mica, lignite, and generally very little clay matrix. The basal zone is often pebbly. The formation has a maximum known thickness of about 520 ft (158 m) in Georgia **(Clarke et al. 1985)**. Total thickness of the unit in the USGS test holes ranges from 157 to 207 ft.

The Pio Nono/Middendorf Formation contains two distinct subunits (informal, ascending Subunits 1 and 2) in the Millhaven, Girard, and Millers Pond cores (Figure 2.5.1-9). Each includes a basal lag deposit of poorly sorted sand that grades up to interbedded and interlaminated clay and sand. Micaceous and lignitic sand laminae are common in the Middendorf sections, particularly near the top of each subunit. Clay beds in some of the cores display more abundant iron-oxide staining near the top of the subunits. A root-trace pattern is observed in the clay bed at the top of Subunit 2 in the Girard core.

A Santonian age is suggested by current research **(Falls and Prowell 2001)**. Lithologic and geophysical log patterns indicate the upper contact of the Pio Nono/Middendorf Formation in Georgia correlates to the UK2/UK4 boundary of Prowell et al. (1985a) and to the upper contact of the Pio Nono/Middendorf in South Carolina **(Fallaw and Price 1995)**.

Dinoflagellates and other marine indicators are sparse and suggest a marginal-marine environment in the downdip cores and a nonmarine environment for this unit in the other cores.

Near Bamberg, South Carolina, the Pio Nono/Middendorf Formation consists of poorly sorted, gray, medium to very coarse grained, angular to subangular quartz sand with quartz pebbles and sparse feldspar grains **(Logan and Euler 1989)**. Silt and fine grained sand are present. The angularity and large overall grain size of the quartz and the presence of feldspar indicate that deposition occurred relatively close to the source area, most likely in an upper delta plain environment. In southeastern Georgia, the Middendorf includes some shallow shelf sediments. Farther downdip, sediments of the Middendorf become finer grained. In Allendale County, South Carolina, in the vicinity of Millet, the unit consists of light gray to colorless, fine to coarse grained quartzose sand, clayey sand, and silty clay. The sand is unconsolidated and poorly to moderately sorted. Trace amounts of heavy minerals and lignite are present. Deposition most probably occurred on a lower delta plain **(Logan and Euler 1989)**.

Gaillard /Black Creek Formation (regionally the Black Creek Group)

The Gaillard/Black Creek Formation (Black Creek Group) is distinguished from the overlying and underlying Cretaceous units by its better-sorted sand, fine grained texture, and relatively high clay content. It is generally darker, more lignitic, and more micaceous, especially in the updip part of the section, than the other Cretaceous units. The total thickness of the unit in the study area ranges from 270 to 333 ft. Falls and Prowell (2001) found that, in Georgia, the Gaillard Formation typically consists of three subunits:

- Subunit 1 - Basal, lignitic sand
- Subunit 2 - Laminated black clay and sand
- Subunit 3 - Coarsening upward sand sequence

The lag deposits observed at the bases of these subunits suggest unconformities. However, the Gaillard Formation/Black Creek Group in the Millers Pond core (updip location) is coarser and sandier, with no recognizable subunits.

Subunit 1 is similar to the underlying Pio Nono/Middendorf Formation. The sediments consist of moderate to poorly sorted, fine to coarse grained quartz sand. The sequence grades up into fine and very fine grained sand with thin beds of clay. Sand layers contain fine lignite and mica with little clay matrix.

Subunit 2 has two sharp contacts within the subunit, observed in the downdip cores. Each contact is overlain by a basal lag of very poorly sorted sand. Most of Subunit 2 is calcareous and contains laminae and lenses of very fine sand and sand-filled burrows. There are thick sections of silty laminated black clay. The sand layers typically contain mica, lignite, and minor glaconite.

Bioturbation features in Subunit 2 include clay-lined burrows, mottled textures, and discontinuous laminae of clay in the sands. Subunit 2 has the most abundant and diverse marine macrofaunas, microfaunas, and microfloras in the Cretaceous section in the study area, including shark teeth, pelecypods, ostracodes, benthic and planktonic foraminifers, spicules, dinoflagellates, pollen, and calcareous nannofossils.

Subunit 3 typically consists of a coarsening upward sequence with very poorly sorted lag deposit at the base. Above the basal lag, moderate to well sorted, very fine to medium grained sand occur. Higher in the section, fine to coarse grain sand is found. There are laminae and thin beds of dark grey clay, large and small lignite pieces, mica and crossbedding.

Paleontological data from the Gaillard Formation/Black Creek Group indicate a Campanian age for the unit **(Frederiksen et al. 2001; Bukry 2001; Gohn 2001)**. Units UK4 and UK5 **(Prowell et al. 1985a)** and the Black Creek Group at the SRS **(Fallaw and Price 1995)** are assigned an age of Campanian to Maastrichtian. The diversity and abundance of dinoflagellates, the abundance of marine faunas, and the presence of glauconite at Millhaven and Girard coreholes suggest a strong marine influence during the deposition of Subunit 2, probably in the distal part of a deltaic complex. Dinoflagellates in Subunit 3 suggest a marginal-marine depositional environment. The composition of the microflora and the absence of other marine indicators suggest that Subunit 1 at Millhaven and Girard, and the entire section of the Gaillard Formation/Black Creek Group at Millers Pond reflect sedimentation in a nonmarine part of the delta **(Frederiksen et al. 2001)**.

Steel Creek Formation

The contact between the Steel Creek Formation and the underlying Gaillard Formation/Black Creek Group is subtle in most of the test cores in the region **(Kidd 1996; Huddleston and Summerour 1996)**. The contact may be conformable, paraconformable, or gradational at various locations in the region. The USGS study **(Falls and Prowell 2001)** identified the contact as unconformable and placed the boundary based on projection of the contact from a downdip test hole. The Steel Creek Formation in the study area ranges from 38 to 197 ft thick.

The Steel Creek Formation in Georgia typically consists of poorly to very poorly sorted, fine to very coarse grained sand with granules and pebbles of smokey quartz. Clay matrix ranges from 5 to 15 percent. Basal lags are characteristically overlain by thick sections of oxidized clay. There are multiple fining-upward sequences in the formation. Cross-bedding is common. Lignite and mica are common accessories.

The formation is mostly barren of fossils. Some core samples yielded Cretaceous and Paleocene palynomorphs **(Fredericksen et al. 2001)**. Paleontology data from the Black Creek

Group below and the Tertiary Black Mingo/Ellenton Formation above restrict the age to Maastrichtian.

Coarse grained sediments, fining-upward sequences, iron oxide staining, and indications of root-bearing zones indicate channel and overbank deposits in a delta-plain environment.

<u>Tertiary Sediments</u>

Tertiary sediments in the site area range in age from Early Paleocene to Miocene and were deposited in open marine shelf environments (downdip) to marginal-marine environments (updip) **(Huddleston and Summerour 1996, Falls and Prowell 2001)**. The Tertiary sequence is divided into five units: Black Mingo/Ellenton Formation, Snapp Formation, Congaree Formation [regionally, the Fourmile Branch/Congaree/Warley Hill (FM/C/W) unit], Lisbon Formation (regionally, the Santee Limestone), and Barnwell Group/unit (Figure 2.5.1-8). The section is generally more calcareous in the downdip area than in the updip area, with greater diversity and abundance of marine microflora and fauna. The thickness of the entire Tertiary section in the study area ranges from 284 to 642 ft.

Black Mingo Formation

The Black Mingo Formation varies from a predominantly clay-rich section in the downdip area into a more sandy, clayey, and calcareous area further updip and finally to a predominantly sandy section in the most updip locations. The formation varies in total thickness from 33 to 72 ft. In the downdip core, the section contains well sorted, fine to medium grained sand with large portions of well laminated clay and silty clay. The top of the section in the downdip area is calcareous and non-calcareous clay. Further updip in the Girard core, the section contains sandy carbonate, limestone, calcareous sand, and abundant glauconite, along with well laminated, non-calcareous silty clay. The sand layers are typically fine to medium grained. Several lag deposits examined have 10 to 20 percent glauconite, rounded phosphatic clasts, and shark teeth. In the updip core, the section is generally fine to very coarse grained sand interbedded with sandy clay. Midway in this section are interlaminated black lignitic clay and very fine to medium grained sand. The top of the section is clayey, fine to medium grained sand.

Paleontologic studies identify a diverse microflora of dinoflagellates, pollen, and calcareous nannofossils and a faunal component of ostacodes, planktonic foraminifers, pelecypods, and gastropods in the downdip sections **(Edwards et al. 2001)**. The updip section at Millers Pond contains a low-diversity microflora of dinoflagellates and pollen **(Clarke et al. 1994)**. The marine fossils, glauconite, and carbonate in downdip sediments indicate an open-marine environment, possibly distal prodelta. The low diversity and the low abundance of dinoflagellates and the absence of other marine indicators at Millers Pond suggest a change to a more restricted marginal-marine environment.

Snapp Formation

The Snapp Formation in all test holes in the study area consists of moderate to poorly sorted, fine to very coarse grained sand. The sand typically contains granules and pebbles and less than 10 percent clay matrix. The formation in the study area is from 50 to 67 ft thick. The sand section is overlain by a very light grey colored kaolin zone that is oxidized and stained red and yellow by iron oxides. Pedogenic structures are found in the clay and include root traces and dessication cracks. Pyrite is disseminated in the clay and also concentrates along dessication cracks. In general, there is a strong fining-upward sequence from bottom to top within the formation.

The Snapp Formation pinches out in the vicinity of the VEGP site. Snapp was not found in the Thompson Oak core. This is consistent with the described updip limit of Snapp in South Carolina, near the Upper Three Runs creek in Aiken County **(Fallaw and Price 1995)**.

Paleontology samples from the downdip core hole (Millhaven) provided sparse dinoflagelates **(Edwards et al. 2001)**.

The stratigraphic position of the Snapp Formation between the Black Mingo Formation and the overlying early Eocene part of the Congaree Formation suggests that the age of the strata is either late Paleocene **(Prowell et al. 1985a; Fallaw and Price 1995)** or early Eocene **(Harris and Zullo 1992)**.

Sedimentary characteristics suggest a fluvially dominated depositional environment in either an upper delta plain or an incised alluvial valley. The presence of dinoflagellates in the Millhaven core suggests a marginal-marine environment in the downdip part of the study area.

Congaree Formation

The Congaree Formation unit consists of mixed siliciclastic/carbonate sections in the downdip test cores and grades to siliciclastic sediments in updip holes. The total section varies from 9 to 103 ft thick.

Downdip, the Congaree Formation unit contains interbedded quartz sand, marl, and limestone. The sand is moderately well sorted, very fine to fine at the bottom, grading to fine to medium grained higher in the same section. Carbonate layers, some of which are lithified, include glauconite, and have a clay matrix and fossils. Extensive burrowing is described in the sandy carbonate material.

Further updip in the Girard core, the sand layers range from fine to medium grained at the bottom of the section and grade up into medium to coarse grained sand interbedded with sandy carbonate, and limestone. Unlithified sandy carbonate and partially lithified calcareous sand are typical in the top portion of the section. In the most updip test hole, the unit is a 9-ft section of well sorted, very fine to fine sand with less that 5 percent clay matrix **(Falls and Prowell 2001)**. However, there are clay-lined burrows.

Dinoflagellates, pollen, and calcareous nannofossils were recovered from the core samples of the Congaree Formation unit at Millhaven and Girard. Dinoflagellates and pollen were recovered from the Thompson Oak and Millers Pond cores. Paleontologic examination of these core samples indicate that this unit is early Eocene to early middle Eocene in age and that it includes more than one biostratigraphic unit **(Bybell 2001; Frederiksen 2001; Edwards 2001)**. Other fossils observed in the Millhaven core included bryozoans, pelecypods, and foraminifers below a depth of 462 ft, and pelecypods and foraminifers above a depth of 462 ft. In the Girard core, pelecypods, bryozoans, and shark teeth were observed above a depth of 342 ft. Biomoldic pores indicate that gastropods also were present **(Falls and Prowell 2001)**.

Sedimentary characteristics of the Still Branch Formation sections in the Girard and Thompson Oak cores suggest a nearshore-marine environment. Sedimentary characteristics of the overlying Congaree beds suggest an open-marine shelf environment for deposits in the downdip core and a fluvially dominated to marginal-marine environment for deposits in the updip cores in the vicinity of Millers Pond. However, the Warley Hill Formation at Millhaven also was deposited in an open-marine shelf environment **(Falls and Prowell 2001)**.

Lisbon Formation

The Lisbon Formation is predominantly calcareous clay with limestone with with a few beds of calcareous sand and clay. As discussed, the Lisbon Formation in eastern Georgia **(Huddleston and Summerour 1996)** includes lithologies assigned to other units **(Steele 1985; McClelland 1987; Fallaw and Price 1992; 1995, and Falls and Prowell 2001)**; the Blue Bluff Marl of the Lisbon Formation **(Huddleston and Hetrick 1985)**, the Santee Limestone **(Sloan 1908)**, and the McBean Limestone Formation **(Veatch and Stephenson 1911; Huddleston and Summerour 1996)**. The unit ranges between 52 and 173 ft in thickness in the VEGP area.

In the downdip Millhaven core, the Lisbon Formation can be divided into lower, middle, and upper sections. The lower section is a medium to coarse grained calcareous sand that grades into a fine to medium grained sandy carbonate with large oyster shells. The contact between the lower and middle sections is distinguished by a layer that is phosphatized and pyritized. The middle portion of the section is predominantly carbonate sediments that vary from marl to carbonate, with little sand in either lithology. The carbonate is well lithified with biomoldic porosity. The marl is burrow mottled to wavy laminated with minor amounts of lignite and pyrite. Fossils include foraminifers, spicules, shark teeth, pelecypods, gastropods, bryozoans, echinoids, and brachiopods. The upper section contains fine to medium grained sand with glauconite and marine fossils, including pelecypods, gastropods, and bryozoans.

The Lisbon Formation in the Girard test hole further updip contains a very sandy limestone with glauconite base layer. This lithology has biomoldic porosity. Overlying limestone is a marl. The contact between them is pyrite rich. The marl is a fine grained sandy carbonate with up to 30 percent clay matrix. The sand content continues to increase to 25 percent and is very fine to fine grained. There are few macrofossils in this portion of the Lisbon.

The Lisbon Formation in the updip Millers Pond core consists of a lower sandy limestone and calcareous sand. A thin basal lag of very poorly sorted sand separates the calcareous material from a more sandy section above. The quartz sand is fine to very coarse in calcareous sand beds and fine to medium in sandy limestone beds. The limestone below a depth of 121 ft is finely crystalline and contains glauconite and marine fossils, including pelecypods, spicules, foraminifers, and shark teeth. Marine fossils in the limestone above a depth of 100 ft include oysters and other pelecypods, foraminifers, and echinoid fragments. Biomoldic porosity also is present above a depth of 100 ft and reflects dissolution of aragonitic pelecypods and gastropods.

Calcareous nannofossils, planktonic foraminifers, dinoflagellates, and pollen from the core localities indicate a late middle Eocene (late Claibornian) age for the Lisbon sections **(Edwards et al. 2001)**. Marine fossils and carbonate suggest that this unit was deposited in an open-marine, shallow-shelf environment. The distribution of siliciclastic sediments and the diversity of marine fossils in the carbonate facies suggest that the updip Millers Pond core is more proximal to a source of siliciclastic sediments than the downdip Millhaven core.

Barnwell Unit

The Barnwell unit, as described in Falls and Prowell (2001), includes the Barnwell Groups plus post-Eocene strata in the study area. The Barnwell Group includes the Utley Limestone Member, Clinchfield and Dry Branch Formations, and the Tobacco Road Sand. The unit is 82 to 250 ft thick in the study area.

In the downdip Millhaven core, the section begins with calcareous clay overlain by moderately well to well sorted, fine to medium grained, calcareous quartz sand followed by partially lithified sandy limestone. Glauconite is a typical accessory mineral in the sand. Higher in the section, thin beds of silica-replaced limestone are common. Other lithologies in the upper section include unlithified carbonate, partially lithified limestone, and irregularly shaped phosphatized limestone clasts. The top of this unit consists of a coarsening-upward sequence of clayey sand and sandy clay.

Fossils observed in the core include pelecypods, bryozoans, echinoids, and foraminifers. Biomoldic pores are present and reflect dissolution of aragonitic pelecypods and gastropods.

The Barnwell unit in the Girard core consists of a lower portion of clay, sand, and carbonate layers. A basal calcareous clay is overlain by partially silicified, phosphatized, and glauconitic limestone; a calcareous quartz sand; a sandy limestone; a marl; and a sandy limestone that grades into an overlying quartz sand. Sand is fine to coarse near the base and grades into very fine to fine for the rest of the section. The clay matrix ranges from 20 to 40 percent. The top part of the Barnwell unit section is noncalcareous and contains clayey sand and clay. The sand ranges from fine to coarse grained.

Fossils include pelecypods and bryozoans. Biomoldic porosity ranges from 5 to 20 percent in the limestone and reflects dissolution of aragonitic pelecypods.

At the updip Millers Pond test hole, the unit contains siliciclastic sediments with essentially no limestone or carbonate. It includes thin beds of fine to medium grained sand and fine to very coarse grained sand and thin beds of well-laminated clay. The sand contains fine lignite, clay clasts, and 10 to 20 percent clay matrix. The top of the section is a coarsening-upward sequence of sand that ranges from fine to medium grained sand up to fine to very coarse sand with a clay matrix from 5 to 25 percent. Sedimentary structures include clay laminae and clay wisps. The Barnwell unit is mapped as the uppermost stratigraphic unit at the Millers Pond site **(Falls and Prowell 2001)**, where it includes the Tobacco Road Sand and Irwinton Sand members of the Dry Branch Formation.

Paleontologic data for the Millhaven and Girard cores suggest a late Eocene to questionably early Oligocene age for the Barnwell sections **(Edwards et al. 2001)**. The Barnwell Group contains eqivalent units to E6, E7, E8, and MI **(Prowell et al. 1985a)** in addition to the Clinchfield Formation, Dry Branch Formation, and Tobacco Road Sand of the Barnwell Group at the SRS **(Fallaw and Price 1995)**. Throughout the study area, the abundance of carbonate, the presence of glauconite and phosphate, and the abundance of marine macrofossils and microfossils in the calcareous part of the section indicate that the Barnwell strata were deposited in open-marine environments. The calcareous sand probably was deposited in a shallow-shelf environment, and the fossil bed at the base is a lag deposit produced by a late Eocene marine transgression. The noncalcareous sand and clay, the ovoid flattened pebbles, and the clay wisps in the upper part of the Barnwell unit suggest that these strata were deposited in nearshore-marine environments.

Quaternary Surfaces and Deposits

The Quaternary record in the VEGP area is preserved primarily in the fluvial terraces along the Savannah River and its major tributaries and in deposits of colluvium, alluvium, and eolian sediments on upland interfluve areas.

The drainage systems within the site consist entirely of streams tributary to the Savannah River. A series of nested fluvial terraces are preserved along the east side of the Savannah River (Figure 2.5.1-29). Fluvial terraces are the primary geomorphic surface that can be used to evaluate Quaternary deformation within the Savannah River area. However, limited data are available for the estimation of ages of river terraces in both the Atlantic and Gulf coastal plains **(Markewich and Christopher 1982; Colman 1983; Mixon et al. 1989; Markewich 1985; Soller 1988)**.

The development of major stream terraces is represented by a sequence of erosional and depositional events in response to tectonism, isostasy, and climatic variations that upset the fluvial system equilibrium. Streams respond to uplift by cutting down into the underlying substrate to achieve a smooth longitudinal profile that grades to the regional base level. Aggradation or deposition occurs when down-cutting is reversed by a rise in base level. The stream channel is elevated and isolated from the underlying marine strata by layers of newly deposited fluvial sediments. Down-cutting may resume again, and the aggraded surface is abandoned. The result is a land form referred to as a fill terrace.

Along the east side of the Savannah River in the vicinity of the VEGP site, there are two prominent terraces above the modern flood plain (Qal): the Bush Field (Qtb) and the Ellenton (Qte) terraces (Figure 2.5.1-29) **(Geomatrix 1993; Hanson et al. 1993)**. These designations are based on morphology and relative height above local base level, which is the present elevation of the Savannah River channel. In addition, there are other minor terraces, one lower than the Bush Field and several higher older terrace remnants. Other investigators have delineated essentially the same terraces **(Stevenson 1982; Brooks and Sassaman 1990)**; however, there are significant differences in the estimated ages of these terraces.

The modern flood plain is a broad alluvial surface that is 6 to 9 ft (2 to 3 m) above the channel. Local relief is about 3 ft (1 m). Refer to Stevenson (1982) for more extensive details. The next terrace at a structurally higher position is the Qty terrace, but that terrace is minor and not laterally continuous **(Geomatrix 1993)** (Figure 2.5.1-29).

The Bush Field terrace (Qtb) surface ranges from 26 to 43 ft (8 to 13 m) above the Savannah River at SRS. This terrace is preserved primarily on the northeast side of the river. Limited subsurface data indicate the terrace is 29 to 49 ft (9 to 15 m) thick. The Ellenton terrace (Qte) is the higher of the two major terraces and includes surfaces that range from 56 to 82 ft (17 to 25

m) above the river channel. Subsurface data indicate the terrace is at least 6 to 10 ft (2 to 3 m) thick.

Estimated ages of the terraces are based on several techniques, including radiometric Carbon-14 dates, soil chronosequences, relative position above base level, and correlation to other dated river or marine terraces. The modern flood plain is as old as the latest Pleistocene to Holocene **(Geomatrix 1993)**. Others have indicated a much younger age of 4,000 years **(Brooks and Sassaman 1990)**. The Qtb terrace appears to be about 90 thousand years old (ka), based on correlation, relative position, and terrace morphology; based on soil chronosequences, the Qtb terrace surface may be as much as 200 ka considering the various methods for estimating the age of the Qtb terrace, Geomatrix (1993) assigned an age estimate of 100 to 250 ka for the Qtb terrace. The Qte terrace is about 200 ka to more than 760 ka, based on regional correlation and morphology; based on soil chronosequences, the Qte terrace surface is at least 400 ka to perhaps 1 Ma. Considering the various methods for estimating the age of the Qte terrace, Geomatrix (1993) assigned an age estimate of 350 ka to 1 Ma for the Qte terrace. The terraces on Upper Three Runs range from 11 ka for the lower (1.6 to 14.8 ft [0.5 to 4.5 m]) terrace to 38 to 47 ka for the higher (greater than 19.7 ft [6 m]) terrace.

2.5.1.1.4 Regional Tectonic Setting

The regional tectonic setting in which the VEGP site is located is presented below. This section includes discussions of regional plate tectonic evolution, regional tectonic stresses, and principal regional tectonic structures.

2.5.1.1.4.1 Plate Tectonic Evolution of the Appalachian Orogenic Belt at the Latitude of the Site Region

The VEGP site lies within the southern part of the northeast-southwest-trending Appalachian orogenic belt, which extends nearly the entire length of the eastern United States from Alabama to southern New York State. The Appalachian orogenic belt formed during the Paleozoic Era and records the opening and closing of the proto-Atlantic along the eastern margin of ancestral North America. The geologic history of the region surrounding the VEGP site is discussed in Section 2.5.1.1.2.

Depending on the focus of a given study, the Appalachian orogenic belt has been subdivided in a variety of ways by various researchers. These subdivisions include provinces, belts, and terranes. Provinces, which are generally more regional in extent, are defined based on both physiography (landforms) and geology (Figure 2.5.1-1). Five physiographic provinces have been defined across the Appalachian belt. From west to east, these include the Appalachian Plateaus (the “Cumberland Plateau” at the latitude of the site region) and the Valley and Ridge, Blue Ridge, Piedmont, and the Coastal Plain physiographic provinces (Figure 2.5.1-1). The Blue Ridge and Piedmont physiographic provinces are further divided into different lithotectonic belts of similar rock type and/or tectonic origin (Figure 2.5.1-14). Some geologists further divide the lithotectonic belts into individual fault-bounded terranes [e.g., **(Horton et al. 1989, 1991; Hatcher et al. 1994)**]. The terms "belt" and "terrane" are used interchangeably in this section to describe fault-bounded crustal blocks that share a common tectonic history.

Since the publication of EPRI (1986), geologists such as Hatcher (1987) and Horton et al. (1989, 1991) have proposed models for the development of the Appalachian orogen and its component provinces in the context of collisional tectonic events that added new fragments of crust to the eastern margin of North America and finally closed an ancestral ocean basin between ancestral North America ("Laurentia") and ancestral Africa ("Gondwana"). The most recent synthesis of the Appalachian orogen at the latitude of the ESP study region is by Hatcher et al. (1994) and incorporates analysis of gravity, magnetic, and seismic reflection data in the interpretation. From northwest to southeast, Hatcher et al. (1994) recognized the following principal tectonic elements of the Appalachian orogen:

- The Valley and Ridge Province. This province encompasses a sequence of sedimentary rocks originally deposited on North American crust and subsequently deformed by folds and thrust faults during the Alleghanian orogeny, which resulted from the collision of Gondwana with Laurentia at the end of the Paleozoic Era.
- The Blue Ridge Province: The western part of the Blue Ridge province is a thrust-bounded sheet of crystalline rocks with overlying sedimentary strata that originally lay along the Paleozoic eastern margin of Laurentia. The eastern part of the Blue Ridge province is underlain by high-grade metamorphic rocks, some or all of which may be exotic to ancestral North America. At the latitude of the VEGP site region, the eastern and western parts of the Blue Ridge province are separated by the Hayesville fault. The eastern tectonic boundary between the Blue Ridge and Piedmont provinces is the Brevard fault.
- The Piedmont Province: The Piedmont province is subdivided into the Inner Piedmont belt on the west, and the Carolina–Avalon Terrane **(Hatcher et al. 1994)** on the east. The boundary between these two units is the Towaliga fault. Rocks of the Carolina–Avalon Terrane extend east of the Piedmont province beneath the overlying sedimentary cover of the Coastal Plain province.
- The Coastal Plain Province: This province is defined by a sequence of predominantly Cretaceous and Tertiary marine sediments overlying Paleozoic crystalline rocks and Mesozoic sedimentary rocks. The Coastal Plain strata record development of a passive continental margin along the eastern United States following Mesozoic rifting and opening of the Atlantic Ocean basin.

Modern plate tectonic reconstructions of the southern Appalachian orogenic belt published since EPRI (1986) interpret that at least some of the major regional Paleozoic deformation events (i.e., Penobscottian, Taconic, Acadian, and Alleghanian) are associated with collisions of exotic or suspect terranes with ancestral North America [e.g., **(Horton et al. 1989; Hatcher et al. 1994)**] (Figure 2.5.1-10). Most geologists generally agree that folded strata in the Valley and Ridge province and the crystalline rocks in the western Blue Ridge province are native to North America [e.g., **(Horton et al. 1991; Hatcher et al. 1994)**] and that these units have been transported westward from their original position along Paleozoic east-dipping, west-verging thrust faults. Key differences between plate tectonic models arise from varying interpretations of which belts and terranes represent exotic or suspect terranes and the location of primary tectonic boundaries or sutures that juxtapose such exotic terranes against North American crust and to one another. At the latitude of the VEGP site, there is general agreement among geologists regarding which belts and terranes are native to North America and which are exotic. The primary differences among models concern the precise location and downdip geometry of the major faults and sutures [e.g., alternative interpretations in **(Hatcher et al. 1994)**] that separate these terranes.

Development of the Appalachian orogen began in the late Precambrian with rifting of the Precambrian basement of ancestral North America, opening of an ocean basin, and formation of a passive margin (Figure 2.5.1-10). Sediments accumulated in local fault-bounded basins in the early phases of rifting, followed by deposition of a characteristic off-lapping, passive-margin sequence of marine carbonate and siliciclastic sediments.

The Grenville Front is the leading edge of a northeast-southwest-trending Precambrian collisional orogen that involved rocks of the pre-Appalachian basement of Laurentia (i.e., ancestral North America) (Figures 2.5.1-10 and 2.5.1-12). The following discussion is summarized from White et al. (2000). Like the younger Appalachian orogen, the Grenville orogen may have formed in part from exotic terranes that were assembled prior to 1160 Ma, then deformed and thrust westward over the pre-Grenville Laurentian margin between 1120 and 980 Ma. The Grenville orogen and Grenville front primarily are exposed in southeastern Canada, and can be traced in outcrop southwest to the latitude of Lake Ontario. Grenville-age rocks and structures continue on trend to the southeast into the United States, but are depositionally and structurally overlain by younger rocks, including terranes of the Appalachian orogen **(Bickford et al. 1986; Hauser 1993)**. Seismic reflection profiles indicate that the Grenville front and other prominent reflectors generally dip toward the east and extend to lower crustal depths **(White et al. 2000)**. Bollinger and Wheeler (1988) note that Iapetan normal faults likely decrease in size, abundance, and slip northwestward from the Grenville front.

The Penobscottian event is the earliest major orogeny recognized in the Appalachian belt and primarily is expressed in the northern Appalachians. Horton et al. (1989) stated that evidence for the Penobscottian orogeny has not been observed south of Virginia. The earliest Paleozoic deformation along or adjacent to the ancestral North American margin at the latitude of the VEGP site region occurred in the Middle Ordovician and is known as the Taconian event or orogeny (Figure 2.5.1-10). The onset of the Taconian event is marked regionally throughout much of the Appalachian belt by an unconformity in the passive-margin sequence and deposition of clastic sediments derived from an uplifted source area or areas to the east. The Taconic event at the latitude of the VEGP site region is interpreted by Horton et al. (1989) and Hatcher et al. (1994) to have resulted from the collision of one or more terranes with North America. Rocks of the eastern Blue Ridge and Inner Piedmont are interpreted to have originated east of the Laurentian passive margin in Middle Ordovician time and, thus, are candidates for Taconic collision(s).

Horton et al. (1989) included the eastern Blue Ridge at the latitude of the ESP study region in the Jefferson Terrane, a large body of sandstones, shales, basalt, and ultramafic rocks interpreted to be a metamorphosed accretionary wedge that accumulated above a subduction zone. Hatcher et al. (1994) suggested that the Hayesville thrust, which forms the western structural boundary of the eastern Blue Ridge and dips eastward beneath it, may be the "up-dip leading edge of an early Paleozoic subduction zone." If this interpretation is correct, then the Hayesville thrust fault may be a Taconic suture. The Carolina-Avalon Terrane also is interpreted by Horton et al. (1989) and Hatcher et al. (1994) to have been accreted during the Taconic orogeny. If this is correct, then the Towaliga fault between the Inner Piedmont and Carolina–-Avalon Terranes also may be a Taconic structure.

According to Horton et al. (1989), evidence for the middle Paleozoic Acadian orogeny is "neither pervasive nor widespread" south of New England. The Acadian event primarily is expressed at the latitude of the ESP study region by unconformities in foreland stratigraphic succession, plutonism, and activity of several major faults **(Hatcher et al. 1994)**, and possibly ductile folding elsewhere in the southern Appalachians **(Horton et al 1989)**. To date, geologists have not observed compelling evidence for a major accretion event at the latitude of the VEGP site region during the Acadian orogeny **(Horton et al. 1989; Hatcher et al. 1994)**.

The final and most significant collisional event in the formation of the Appalachian belt was the late Paleozoic Alleghanian orogeny, during which Gondwana collided with Laurentia, closing the intervening Paleozoic ocean basin (Figure 2.5.1-10). At the latitude of the VEGP site region, the Alleghanian collision telescoped the previously accreted Taconic terranes and drove them westward up and across the Laurentian basement, folding the passive margin sequence before them and creating the Valley and Ridge fold-and-thrust belt. The collisional process also thrust a fragment from the underlying Laurentian basement eastward over the passive margin sequence, forming the western Blue Ridge. Significant strike-slip faulting and lateral transport of terranes also are interpreted to have occurred during the Alleghanian orogeny **(Hatcher et al. 1994)**.

At the latitude of the VEGP site region, the Alleghanian suture between rocks of ancestral Africa and terranes accreted to North America is buried beneath the Coastal Plain sediments and not exposed. Hatcher et al. (1994) offered alternative interpretations that the Alleghanian suture (1) coincides with the western edge of the Mesozoic Dunbarton Basin, which implies that the suture was reactivated as a normal fault during Mesozoic rifting and opening of the modern Atlantic basin or (2) is located somewhere east of the Dunbarton Basin, and its location and geometry are not known precisely. Based on detailed petrologic and geochemical analyses of basement samples from deep boreholes at the SRS, Dennis et al. (2004) concluded that the basement rocks there correlate with meta-igneous rocks of the Carolina–Avalon Terrane, which implies that the Alleghanian suture must lie east of the VEGP site.

Despite uncertainties regarding the precise origin, emplacement, and boundaries of belts and terranes, there is good agreement among tectonic models regarding first-order structural features of the southern Appalachian orogenic belt. At the latitude of the VEGP site region, the ancestral North American basement of the Paleozoic passive margin underlies the Valley and Ridge, Blue Ridge, and Inner Piedmont provinces at depths of less than 6 to 9 mi (10 to 15 km), and possibly as shallow as 3.1 mi (5 km) or less beneath the Valley and Ridge (Figure 2.5.1-11). A basal decollement along the top of the North American basement is the root zone for Paleozoic thrust faults in the Valley and Ridge, Blue Ridge, and Inner Piedmont provinces. Although potential seismogenic sources may be present within the North American basement below the decollement **(Wheeler 1995)**, the locations, dimensions, and geometries of these deeper potential sources are not necessarily expressed in the exposed fold-thrust structures above the detachment.

Wheeler (1995) suggests that many earthquakes in the eastern part of the Piedmont province and beneath the Coastal Plain province may be associated spatially with buried normal faults related to rifting that occurred during the Mesozoic Era (Figure 2.5.1-10). Normal faults in this region that bound Triassic basins may be listric into the Paleozoic detachment faults [e.g., **(Dennis et al. (2004)**] or may penetrate through the crust as high-angle faults. However, no definitive correlation of seismicity with Mesozoic normal faults has been conclusively demonstrated.

2.5.1.1.4.2 Tectonic Stress in the Mid-Continent Region

Earth science teams (ESTs) that participated in the EPRI (1986) evaluation of intra-plate stress found that tectonic stress in the Central and Eastern United States (CEUS) region primarily is characterized by northeast-southwest-directed horizontal compression. In general, the ESTs 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 ESTs noted that the regional northeast-southwest trend of principal stress may vary in places 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.

During the 1980s, an international effort to collate and evaluate stress indicator data culminated in publication of a new World Stress Map **(Zoback and Zoback 1989; Zoback et al. 1989)**. Data for this map are ranked in terms of quality. 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 ESTs is statistically robust and is consistent with the theoretical trend of compressive forces acting on the North American plate from the mid-Atlantic ridge **(Zoback 1992)**.

The more recent assessments of lithospheric stress do not support inferences by some EPRI ESTs 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. Zoback and Zoback (1989) summarized a variety of data, 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 US are characterized by uniform 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 (1989) 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.

A detailed study by Moos and Zoback (1992) evaluated the orientations and magnitudes of the principal stresses at the Savannah River Site, north of the VEGP site, to a depth of about 4,000 ft (1,220 m) using data and observations from the New Production Reactor (NPR) borehole. They inferred that the maximum horizontal compressive stress is oriented N50°E to N70°E in the upper 2,100 ft (640 m) depth range, similar to regional trends in the eastern United States reported by Zoback (1992). Moos and Zoback (1992) estimated that the magnitude of the differential stress in the upper 2,100 ft (640 m) is close to the limit of the frictional strength on NW-striking reverse faults. In the depth range of about 3,000 ft to 3,700 ft (915 to 1130 m), Moos and Zoback (1992) found that the maximum horizontal compressive stress was directed about N33°E, more toward the north and counterclockwise of regional trends, and that the magnitude of the differential stress would likely favor strike-slip displacement rather than reverse faulting. In the 4,000 ft (1,220 m) depth range near the bottom of the hole, breakouts and other stress indicators suggest that the maximum horizontal compressive stress is directed toward N55°E, subparallel to regional trends.

The observations from the NPR borehole at the Savannah River Site thus appear to document a counterclockwise rotation of the maximum horizontal compressive stress restricted to a narrow depth range in the upper 4,000 ft (1,220 m) of the crust. Stress indicators from depth intervals above and below this excursion suggest that the NE-SW orientation of the maximum horizontal compressive stress is consistent with regional trends documented in the eastern United States **(Zoback 1992)**.

The significance of these results for tectonic stress and fault activity in the vicinity of the VEGP site are equivocal. The clockwise excursion in the orientation of the maximum horizontal compressive stress in the NPR borehole is limited to a narrow depth range, and may be similarly limited in horizontal extent. The specific cause of the stress rotation in the NPR borehole is not known. Moos and Zoback (1992) discuss several physical reasons why it is “not reasonable” to linearly extrapolate estimates of stress magnitudes at shallow depths to mid-crustal depths where moderate to large magnitudes typically nucleate. Thus, the data from the NPR borehole only provide information on the state of stress in the shallow crust directly adjacent to the borehole. The data suggest that while there may be local perturbations in the stress field near the borehole, the average state of crustal stress is probably characterized by horizontal NE-SW compression, similar to regional trends.

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 (1991) performed numerical modeling of stress in the continental US 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 (1991) emphasized 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 and Schubert 2002)**. The force is an integrated effect over oceanic lithosphere ranging in age from about 0 to about 100 Ma **(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 (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 10^{12} N/m (i.e., the total vertically integrated force acting on a column of lithosphere 3.28 ft [1 m] wide), which corresponds to average equivalent stresses of about 40 to 60 MPa distributed across a 30-mi-thick elastic plate. Richardson and Reding (1991) found that the fit of the model stress trajectories to data was improved by adding compressive stress (about 5 to 10 MPa) acting on the San Andreas fault and Caribbean plate boundary structures. The fit of the model stresses to data further indicates that shear stresses acting on these plate boundary structures must also be in the range of 5 to 10 MPa. Humphreys and Coblenz (in review) also found that the fit of numerical stress models for the North American plate was improved by imposing large compressive stresses on the San Andreas fault and Caribbean plate boundary structures.

Richardson and Reding (1991) noted that the general northeast-southwest orientation of principal stress in the CEUS also could be reproduced in numerical models that assume horizontal shear tractions acting on the base of the North American plate. Richardson and Reding (1991) did not favor this as a significant contributor to total stress in the mid-continent region, however, because their model would require an order-of-magnitude increase in the horizontal compressive stress from the eastern seaboard to the Great Plains. Using numerical models, Humphreys and Coblenz (in review) also evaluated the contribution of shear tractions on the base of the North American lithosphere to intra-continental stress, and concluded that (1) there is a viscous drag or resisting force acting on the cratonic root of North America as it moves relative to the asthenospheric mantle and that this drag supports part of the ridge-push force acting from the east and creates a stress shadow for the western US and (2) shear tractions on the base of North America from flow of the underlying asthenospheric mantle are a minor contribution, if any, to stress in the mid-continental lithosphere. Humphreys and Coblenz (in review) concluded that the dominant control on the northeast-southwest orientation of the maximum compressive principal stress in the CEUS is ridge-push force from the Atlantic basin.

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 Principal Regional Tectonic Structures

Principal tectonic structures and features in the southeastern United States and within the 200-mi VEGP site region can be divided into four categories based on their age of formation or reactivation, and are shown in Figures 2.5.1-12 and 2.5.1-13. These categories include structures that were most active during Paleozoic, Mesozoic, Tertiary, or Quaternary time. Most of the Paleozoic and Mesozoic structures are regional in scale and geologically and geophysically recognizable. The Mesozoic rift basins and bounding faults show a high degree of parallelism with the structural grain of the Paleozoic Appalachian orogenic belt, which generally reflects reactivation of pre-existing Paleozoic structures. Tertiary and Quaternary structures are generally more localized and may be related to reactivation of portions of older bedrock structures.

Regional Paleozoic Tectonic Structures

The VEGP site region encompasses portions of the Coastal Plain, Piedmont, Blue Ridge, Valley and Ridge, and Appalachian Plateau physiographic provinces (Figure 2.5.1-1). Rocks and structures within these provinces are associated with thrust sheets that formed during convergent Appalachian orogenic events of the Paleozoic Era. Tectonic structures of this affinity also exist beneath the sedimentary cover of the Coastal Plain province. These types of structures include the following: (1) sutures juxtaposing allochthonous (tectonically transported) rocks with 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 belts or terranes (Figure 2.5-14). The majority of these structures dip eastward and shallow into a low-angle, basal Appalachian decollement. The Appalachian orogenic crust is relatively thin across the Valley and Ridge province, Blue Ridge province, and western part of the Piedmont province and thickens eastward beneath the eastern part of the Piedmont province and the Coastal Plain province (Figure 2.5-11). Below the decollement are rocks that form the North American basement complex. The basement rocks contain northeast-striking, Late Precambrian to Cambrian normal faults that formed during the rifting that preceded the deposition of Paleozoic sediments (See Section 2.5.1.1.4.1).

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 (Figure 2.5.1-15), is occurring at depths ranging from 3 to 16 mi (5 to 25 km) **(Chapman and Krimgold 1994)**, which is generally below the Appalachian thrust sheets and basal decollement **(Bollinger and Wheeler 1988)**.

Paleozoic faults within 200 mi of the site are shown on Figure 2.5.1-13, and are described as follows:

- Augusta Fault: The Augusta fault zone is located near Augusta, Georgia, about 30 mi north of the VEGP site and separates amphibolite facies gneisses and schists of the Kiokee belt to the northwest from greenschist facies volcanic and volcaniclastic rocks of the Belair belt to the southeast **(Secor et al. 1986a; Maher 1987; Secor 1987)** (Figures 2.5.1-6 and 2.5.1-16). The Augusta fault trends east-northeast and dips moderately to the southeast. The fault contains two distinct deformation fabrics: a mylonite about 800 ft thick is overprinted by a brittle fabric. Kinematic analysis within the mylonite zone reveals a hanging-wall down-component during the movement history **(Maher 1987)**. Furthermore, the hanging wall consists of lower greenschist facies, while the footwall contains upper amphibolite facies. The relative positions of lower grade rocks placed structurally on higher grade rocks in combination with shear sense indicators indicates low-angle normal fault movement for the Augusta fault zone. This is a new view of the Augusta fault zone, which previously had been considered a ductile-to-brittle thrust fault or a strike-slip fault **(Snoke et al. 1980; Cook et al. 1981; Bobyarchick 1981)**. More recent interpretation has concluded that ductile faults with a normal sense component were an important aspect of late Alleghanian deformational history **(Hatcher et al. 1994)**. Recently, Maher et al. (1994) reported 40-Ar/39-Ar ages from samples along a traverse across the Modoc fault and Augusta fault zones. They concluded that a 274 Ma cooling age closely dates initiation of extensional movement on the Augusta fault zone. This cooling age indicates the time when the ductile fabric was generated and therefore when the fault moved. No seismicity is attributed to the Augusta fault.
- Modoc Fault Zone: The Modoc zone, located in South Carolina and Georgia about 40 mi north of the VEGP site, is a region of high ductile strain separating the Carolina Terrane (Carolina Slate and Charlotte belts) from the amphibolite facies migmatitic and gneissic rocks of the Kiokee belt **(Bramlett et al. 1982; Secor 1987)** (Figures 2.5.1-6 and 2.5.1-16). The northeast-trending Modoc zone dips steeply to the northwest and can be traced from central Georgia to central South Carolina based on geological and geophysical data. The Modoc fault zone contains fabrics characterized by brittle and ductile deformation produced

by ductile shear during an early phase of the Alleghanian orogeny **(Maher et al. 1992; Secor et al. 1986a, 1986b)** approximately 315 Ma **(Wyatt and Harris 2000)**. Mylonitic rocks are common within the zone, although the intensity of mylonitization varies widely **(Bramlett et al. 1982)**. The Modoc zone is overprinted by the Irmo antiform near Columbia, South Carolina. Extension of the Modoc fault zone further to the northeast is uncertain, but there are shear zones in North Carolina and Virginia that may be of the same deformational phase **(Bobyarchick 1981; Farrar 1985)**. Sacks and Dennis (1987) reported a normal-sense component in the Modoc zone on the northwest flank of the Kiokee belt. The approximately 315 Ma age of the mylonitic fabric on this fault zone indicates that this fault is very old and therefore neither active nor capable in terms of regulatory guidance. No seismicity is attributed to the Modoc fault zone.

- Eastern Piedmont Fault System: Hatcher et al. (1977) suggested that the Modoc shear zone, the Irmo shear zone, and the Augusta fault are part of the proposed Eastern Piedmont Fault System, an extensive series of faults and splays extending from Alabama to Virginia. Aeromagnetic, gravity, and seismic reflection data indicate that the Augusta fault zone continues in the crystalline basement beneath the Coastal Plain province sediments.
- Other Paleozoic Faults: Other Paleozoic faults within the greater site region include the Brevard fault, the Hayesville fault, the Towliga fault, the Central Piedmont suture, and the Eastern Piedmont fault system (Figure 2.5.1-14). No seismicity is attributed to these Paleozoic faults, and published literature does not indicate that any of these faults offset late Cenozoic deposits or exhibit geomorphic expression indicative of Quaternary deformation. In addition, Crone and Wheeler (2000) and Wheeler (2005) do not show any of these faults to be potentially active Quaternary faults. Therefore, these Paleozoic structures in the site region are not considered to be capable tectonic sources, as defined in Appendix A of RG 1.165.

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 seismic source model.

Regional Mesozoic Tectonic Structures

Wheeler (1995) suggested that many earthquakes in the eastern part of the Piedmont province and beneath the Coastal Plain province may be associated spatially with buried normal faults related to rifting that occurred during the Mesozoic Era. However, no definitive correlation of seismicity with Mesozoic normal faults has been conclusively demonstrated. Normal faults in this region that bound Triassic basins may be listric into the Paleozoic detachment faults [e.g., **(Dennis et al. 2004)**] or may penetrate through the crust as high-angle faults. Within regions of stable continental cratons, areas of extended crust potentially contain the largest earthquakes **(Johnston et al. 1994)**. Mesozoic basins have long been considered potential sources for earthquakes along the eastern seaboard **(Wentworth and Mergner-Keefer 1983)** and were considered by most EPRI teams in their definition of seismic sources **(EPRI 1986C)**. Mesozoic basins and faults in the site region include:

- Mesozoic Rift Basins - A broad zone of fault-bounded, elongate depositional basins associated with crustal extension and rifting formed during the opening of the Atlantic Ocean in early Mesozoic time. These rift basins are common features along the eastern coast of North America from Florida to Newfoundland (Figures 2.5.1-7 and 2.5.1-13). The VEGP site is located within the Dunbarton Basin. This basin is discussed in detail below (Figure 2.5.1-13).

- Dunbarton Basin - The Dunbarton Basin is a roughly east-northeast-trending Mesozoic rift basin located beneath the VEGP site and the SRS (Figure 2.5.1-16). The basin is approximately 31 mi long and 6 to 9 mi wide. Siple (1967) and Marine and Siple (1974) originally identified the general extent and shape of the Dunbarton Basin on the basis of Coastal Plain sediment cores and a limited amount of seismic data from the SRS, as well as aeromagnetic data from Petty et al. (1965). The Dunbarton Basin coincides with both gravity and magnetic lows and is bounded on the north by the Pen Branch fault **(Marine and Siple 1974; Chapman and DiStefano 1989; Stephenson and Stieve 1992; Cumbest et al. 1992, 1998, 2000; Domoracki 1994)**. The Pen Branch fault has had a long and varied history. The Pen Branch fault likely formed in the Paleozoic Era, and was reactivated as a normal fault during the Triassic Period. The Pen Branch fault was most recently rectivated as an oblique-reverse fault in the Cenozoic Era **(Cumbest et al. 1992, 1998, 2000)** and is discussed in greater detail in Section 2.5.1.2.4.1. It has been suggested that the Martin fault (discussed in Section 2.5.1.1.4.5) is the southeastern bounding fault of the Dunbarton Basin **(Snipes et al. 1993a)**, although Domoracki et al. (1999b) suggested that the Dunbarton Basin is instead a half-graben bounded only by the Pen Branch fault to the north.

<u>Regional Tertiary Tectonic Structures</u>

Within 200 mi of the VEGP site, only a few tectonic features, including arches, domes, and embayments, were active during the Tertiary Period. A series of topographic highs and lows in the crust (arches and embayments, respectively) oriented perpendicular to the hinge zone have exerted control over Coastal Plain sedimentation from late Cretaceous through Pleistocene time and are indicative of episodic, differential tectonic movement. The arches are broad anticlinal upwarps, whereas the embayments are broad, sediment-filled basement flexures.

The most prominent arches in the VEGP site region include the Cape Fear Arch on the South Carolina–North Carolina border and the Yamacraw Arch on the Georgia–South Carolina border. The Cape Fear Arch is bordered by the Salisbury embayment to the northeast and the Georgia embayment to the southeast. There is no evidence that these structures are active, and Crone and Wheeler (2000) classified the Cape Fear Arch as a Class C feature, based on lack of evidence for Quaternary faulting (Figure 2.5.1-17).

Regional Quaternary Tectonic Structures

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

Within a 200-mi radius of the VEGP site, Crone and Wheeler (2000) and Wheeler (2005) identified 11 potential Quaternary features (Table 2.5.1-1 and Figure 2.5.1-17). These include: the Charleston, Georgetown, and Bluffton paleoliquefaction features (Class A), the East Coast fault system (Class C), the Cooke fault (Class C), the Helena Banks fault zone (Class C), the Pen Branch fault (Class C), the Belair fault (Class C), the fall lines of Weems (1998) (Class C), the Cape Fear Arch (Class C), and the Eastern Tennessee seismic zone (Class C).

The Charleston features (including the East Coast fault system; the Cooke fault, the Helena Banks fault zone; and the Charleston, Georgetown, and Bluffton paleoliquefaction features) are discussed in detail in Section 2.5.1.1.4.4. The Pen Branch fault is discussed in detail Section 2.5.1.2.4.1, and the Eastern Tennessee Seismic Zone is discussed in detail in Section 2.5.1.1.4.6. The Belair fault zone and the fall lines of Weems (1998) are discussed in detail below:

- Belair Fault Zone - The Belair fault zone has been mapped for at least 15 mi (24 km) as a series of northeast-trending, southeast-dipping, oblique-slip faults near Augusta, Georgia, that generally parallel the structural grain of the Piedmont (Figures 2.5.1-16 and 2.5.1-17). The Belair fault juxtaposes Paleozoic phyllite over Late Cretaceous sands of the Coastal Plain province **(Prowell and O'Connor 1978)**. Mapping and structural analysis by Bramlett et al. (1982) indicate that the Belair fault is probably a tear fault or lateral ramp in the hanging wall of the Augusta fault zone. No geomorphic expression of the fault has been reported **(Crone and Wheeler 2000)**.
- Shallow trenches excavated across the Belair fault near Fort Gordon in Augusta, Georgia, were initially interpreted as revealing evidence for Holocene movement **(Prowell et al. 1975)**, but the apparent youthfulness of movement was probably the result of contaminated radiocarbon samples **(Prowell 2005)**. **Prowell and O'Connor (1978)** demonstrated that the Belair fault cuts beds of Late Cretaceous and Eocene age. Overlying, undeformed strata provide a minimum constraint on the last episode of faulting, which is constrained to sometime between post-late Eocene and pre-26,000 years ago **(Prowell 2005)**.

- There is no evidence of historic or recent seismicity associated with the Belair fault. Crone and Wheeler (2000) classified the Belair fault zone as a Class C feature, since the most recent faulting is not demonstrably of Quaternary age. Quaternary slip on the Belair fault zone is allowed, but not demonstrated, by the available data.
- Fall Lines of Weems (1998) - In his examination of longitudinal profiles of large streams flowing across the Piedmont and Blue Ridge provinces of North Carolina, Virginia and Tennessee, Weems (1998) identified numerous anomalously steep stream segments (Figure 2.5.1-17). Weems (1998) recognized that these “fall zones” are aligned from stream to stream along curvilinear paths, and that these paths are generally subparallel to the regional structural grain. Weems (1998) presented three hypotheses to explain these phenomena, including climatic factors, rock hardness, and neotectonics. Although some fall zones are spatially coincident with changes in rock hardness, Weems (1998) favored a neotectonic origin for the fall lines. Wheeler (2005) classified the fall lines of Weems (1998) as a Class C feature, since the fall zones are not demonstrably reproducible and tectonic faulting is not demonstrated by available data.
- The Cape Fear Arch - The Cape Fear Arch is previously discussed in this section (under Regional Tertiary Tectonic Structures). Crone and Wheeler (2000) classified the Cape Fear Arch as a Class C feature based on lack of evidence for Quaternary faulting.

Regional Geophysical Anomalies and Lineations

In addition to the tectonic features described above, a number of regional geophysical anomalies are located within 200 mi of the VEGP site. These include the East Coast Magnetic Anomaly, the Blake Spur Magnetic Anomaly, the New York-Alabama, Clingman and Ocoee lineaments, and the Grenville Front.

- East Coast Magnetic Anomaly - The East Coast Magnetic Anomaly (ECMA) is a broad, 200 to 300 nT magnetic high that is located approximately 30 to 120 mi (50 to 200 km) off the coast of North America, and which is continuously expressed for about 1,200 mi (1,900 km) from the latitude of Georgia to Nova Scotia **(Klitgord et al. 1988; Withjack et al. 1998)** (Figure 2.5.1-12). The ECMA is subparallel to the Atlantic coastline, and is spatially associated with the eastern limit of North American continental crust **(Klitgord et al. 1988)**. The ECMA has been variously interpreted to be a discrete, relatively magnetic body such as a dike or ridge, or an “edge effect” due to the juxtaposition of continental crust on the west with higher susceptibility oceanic crust on the east (see summary and additional references in Austin et al. 1990). In the vicinity of the ECMA, deep seismic reflection profiling in the Atlantic basin has imaged packages of east-dipping reflectors that underlie the sequence of Mesozoic-Tertiary passive-margin marine strata **(Sheridan et al. 1993)**. The rocks associated with the east-dipping reflectors are interpreted to be an eastward-thickening wedge of volcanic and volcaniclastic rocks that were deposited during the transition between rifting of the continental crust and opening of the Atlantic basin during the Mesozoic **(Withjack et al. 1998)**. Models of the magnetic data show that the presence of this volcanic "wedge" can account for the wavelength and amplitude of the ECMA **(Klitgord et al. 1988)**.

To summarize, the ECMA is a relict of the Mesozoic opening of the Atlantic basin, and probably arises from the presence of a west-tapering wedge of relatively magnetic volcanic rocks deposited along the eastern margin of the continental crust as the Atlantic basin was opening, rather than juxtaposition of rocks with differing magnetic susceptibilities across a fault. The ECMA is not directly associated with a fault or tectonic feature, and thus is not a potential seismic source.

- Blake Spur Magnetic Anomaly - The Blake Spur Magnetic Anomaly (BSMA) is an approximately 150 nT magnetic high located about 150 to 250 mi (250 to 400 km) east of the Atlantic coastline (Figure 2.5.1-12) (the following description is summarized from Klitgord et al. 1988). The BSMA is subparallel to the coastline, and can be traced for about 680 mi (1,100 km) from the latitude of Georgia northward to the Newark Basin. The BSMA is interpreted to have formed in middle Jurassic when the locus of sea-floor spreading in the young Atlantic basin abruptly jumped eastward to the ancestral western margin of Africa, which was the location of the BSMA at that time (see summary and additional references in Klitgord et al. 1988). The magnetic anomaly probably arises from a local variation in the thickness and/or susceptibility of the oceanic crust **(Austin et al. 1990)**. Since its formation, the BSMA has been a passive feature in the Atlantic crust. The BSMA is not directly related to a fault or other tectonic structure, and thus is not a potential seismic source.

- The New York-Alabama, Clingman and Ocoee Lineaments - King and Zietz (1978) identified a 1,000-mi- (1,600-km-) long lineament in aeromagnetic maps of the eastern United States that they referred to as the "New York-Alabama lineament" (NYAL) (Figure 2.5.1-12). The NYAL primarily is defined by a series of northeast-southwest-trending linear magnetic gradients in the Valley and Ridge province of the Appalachian fold belt that systematically intersect and truncate other magnetic anomalies. The NYAL also is present as complementary but less-well-defined lineament on regional gravity maps **(King and Zietz 1978)**.

  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 and Zietz 1981)**. The Ocoee lineament is a described as a splay that branches southwest from the Clingman lineament at about latitude 36°N (see summary in Johnston et al. 1985). The Clingman-Ocoee lineaments are sub-parallel to and located about 30 to 60 mi (50 to 100 km) east of the NYAL.

King and Zeitz (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, 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 (2000) interpreted the NYAL to be a right-lateral wrench fault that formed during an initial phase of late Proterozoic continental rifting that eventually led to the opening of the Iapetan 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 and Zietz 1981)**.

Johnston et al. (1985) 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 and Reinbold 1985)**]. Based on the orientations of nodal planes from focal mechanisms of small earthquakes, Johnston et al. (1985) noted that most events within the Ocoee block occurred by strike-slip displacement on north-south and east-west striking faults, and thus these geologists did not favor the interpretation of seismicity occurring on a single, through-going NE-SW-trending structure parallel to the Ocoee block boundaries.

The Ocoee block lies within a zone defined by Wheeler (1995, 1996) as the cratonward limit of normal faulting along the ancestral rifted margin of North America that occurred during the opening of the Iapetan Ocean in late Precambrian to Cambrian time. Synthesizing geologic and geophysical data, Wheeler (1995, 1996) mapped the northwest extent of the Iapetan faults in the subsurface below the Appalachian detachment, and proposed that earthquakes within the region defined by Johnston and Reinbold (1985) 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.

2.5.1.1.4.4 Charleston Tectonic Features

The August 31, 1886, Charleston, South Carolina, earthquake is one of the largest historical earthquakes in the eastern United States. The event produced Modified Mercalli Intensity (MMI) X shaking in the epicentral area and was felt strongly as far away as Chicago (MMI V) **(Johnston 1996)**. As a result of this earthquake, considerable effort has gone into identifying the source of the earthquake and recurrence history of large magnitude events in the region. In spite of this effort, the source of the 1886 earthquake has not been definitively attributed to any particular fault shown in Figure 2.5.1-18.

The 1886 Charleston earthquake produced no identifiable primary tectonic surface deformation; therefore, the source of the earthquake has been inferred based on the geology, geomorphology, and instrumental seismicity of the region (Figures 2.5.1-18 and 2.5.1-19).

Talwani (1982) infers that the 1886 event was produced by the north-northeast-striking Woodstock fault (inferred from seismicity) near its intersection with the northwest-striking Ashley River fault (also inferred from seismicity). Marple and Talwani (2000) have more recently suggested that a northeast-trending zone of river anomalies, referred to as the East Coast fault system, represents the causative fault for the 1886 Charleston event. The southern segment of the East Coast fault system coincides with a linear zone of micro-seismicity that defines the northeast-trending Woodstock fault of Talwani (1982) and the isoseismal zone from the 1886 earthquake.

Johnston (1996) estimated a moment magnitude of (**M)** 7.3±0.26 for the 1886 Charleston event. More recently, Bakun and Hopper (2004) estimated a smaller magnitude of **M** 6.9 with a 95 percent confidence level corresponding to a range of M 6.4 to 7.1. Both of these more recent estimates of maximum magnitude (Mmax) are similar to the upper-bound maximum range of Mmax values used in EPRI (1986) (body wave magnitude [m_b] 6.8 to 7.5). However, significant new information regarding the source geometry and earthquake recurrence of the Charleston seismic source warrants an update of the EPRI (1986) source models in the PSHA. The updated Charleston seismic source parameters are presented in Section 2.5.2.2.2.4.

Potential Charleston Source Faults

Since the EPRI (1986) source models were developed, a number of faults have been identified or described in the literature as possible sources related to the 1886 Charleston earthquake. These include paleoliquefaction features and numerous faults localized in the Charleston meizoseismal area approximately 85 mi from the site (Figure 2.5.1-18).

There is evidence, in the form of paleoliquefaction features in the South Carolina Coastal Plain, that the source of the 1886 Charleston earthquake has repeatedly generated vibratory ground motion. Paleoliquefaction evidence is lacking for prehistoric earthquakes elsewhere along much of the eastern seaboard [e.g., **(Amick 1990; Amcik et al. 1990c, 1990b)**]. At a minimum, the Charleston seismic source is defined as a seismogenic source according to RG 1.165. Whereas the 1886 Charleston earthquake almost certainly was produced by a capable tectonic source, that specific tectonic structure has yet to be identified. Various studies have proposed potential causative faults for the 1886 event; however, a positive linkage between a discrete structure and the Charleston earthquake has yet to be determined.

These potential causative features are shown in Figures 2.5.1-18 and 2.5.1-19 and described below:

- East Coast Fault System - The inferred East Coast fault system (ECFS, the southern section of which is also known as the “zone of river anomalies” or ZRA based on the alignment of river bends) is a northeast-trending, approximately 373-mi-long (600-km-long) fault system extending from west of Charleston, South Carolina, to southeastern Virginia **(Marple and Talwani 2000a)**. The ECFS comprises three, approximately 124-mi-long (200-km-long), right-stepping sections (southern, central, and northern; Figures 2.5.1-18 and 2.5.1-19). Evidence for the southern section is strongest, with evidence becoming successively weaker northward **(Wheeler 2005)**. Marple and Talwani (1993) identified a series of geomorphic anomalies (i.e., ZRA) located along and northeast of the Woodstock fault and attributed these to a buried fault much longer than the Woodstock fault. Marple and Talwani (1993, 2000) suggested that this structure, the ECFS, may have been the source of the 1886 Charleston earthquake. Marple and Talwani (2000) provided additional evidence for the existence of the southern section of the ECFS, including seismic reflection data, linear aeromagnetic anomalies, exposed Plio-Pleistocene faults, local breccias, and upwarped strata. Since most of the geomorphic anomalies associated with the southern section of the ECFS are in late Pleistocene sediments, Marple and Talwani (2000) speculated that the fault has been active in the past 130–10 ka, and perhaps remains active. Wildermuth and Talwani (2001) used gravity and topographic data to postulate the existence of a pull-apart basin between the southern and central sections of the ECFS, which would imply a component of right-lateral slip on the fault. Wheeler (2005) classified the ECFS as a Class C feature based on the lack of demonstrable evidence that the ECFS has or can generate strong ground motion and the lack of any demonstrable evidence for any sudden uplift anywhere along the proposed fault.

- Adams Run Fault - Weems and Lewis (2002) postulated the existence of the Adams Run fault (Figure 2.5.1-19) on the basis of microseismicity and borehole data. Their interpretation of borehole data suggests the presence of areas of uplift and subsidence separated by the inferred fault (Figure 2.5.1-19). However, review of these data shows that the pattern of uplift and subsidence does not appear to persist through time (i.e., successive stratigraphic layers) in the same locations and that the intervening structural lows between the proposed uplifts are highly suggestive of erosion along ancient river channels. In addition, there is no geomorphic evidence for the existence of the Adams Run fault, and analysis of microseismicity in the vicinity of the proposed Adams Run fault does not clearly define a discrete structure (Figure 2.5.1-20).

- Ashley River Fault - The Ashley River fault was identified by Talwani (1982) on the basis of a northwest-oriented, linear zone of seismicity located about 6 mi west of Woodstock, South Carolina, in the meizoseismal area of the 1886 Charleston earthquake (Figure 2.5.1-19). The postulated Ashley River fault, a southwest-side-up reverse fault, is thought to offset the north-northeast-striking Woodstock fault about 3 to 4 miles to the northwest near Summerville **(Talwani 1982; Talwani 2000a; Weems and Lewis 2002)**.

- Charleston Fault - Lennon (1986) proposed the Charleston fault on the basis of geologic map relations and subsurface borehole data (Figure 2.5.1-19). Weems and Lewis (2002) suggested that the Charleston fault is a major, high-angle reverse fault that has been active at least intermittently in Holocene to modern times. The Charleston has no clear geomorphic expression, nor is it clearly defined by microseismicity (Figure 2.5.1-20).

- Cooke Fault - Behrendt et al. (1981) and Hamilton et al. (1983) identified the Cooke fault based on seismic reflection profiles in the meizoseismal area of the 1886 Charleston earthquake (Figure 2.5.1-19). This east-northeast-striking, steeply northwest-dipping fault has a total length of about 6.2 mi (10 km) **(Behrendt et al. 1981; Hamilton et al. 1983)**. Marple and Talwani (1993, 2000) reinterpreted these data to suggest that the Cooke fault may be part of a longer, more northerly striking fault (i.e., the ZRA of Marple and Talwani [1993] and the ECFS of Marple and Talwani [2000]). Crone and Wheeler (2000) classified the Cooke fault as a Class C feature based on lack of evidence for faulting younger than Eocene.

- Helena Banks Fault Zone - The Helena Banks fault zone is clearly imaged on seismic reflection lines offshore of South Carolina **(Behrendt et al. 1983; Behrendt and Yuan 1987)** and was known to the six EPRI ESTs at the time of EPRI (1986) as a possible Cenozoic-active fault zone (Figure 2.5.1-19). Some ESTs recognized the offshore fault zone as a candidate tectonic feature for producing the 1886 event and included it in their Charleston seismic source zones. However, since 1986, three additional sources of information have become available:

  - In 2002, two magnitude m_b ≥3.5 earthquakes (m_b 3.5 and 4.4) occurred offshore of South Carolina in the vicinity of the Helena Banks fault zone in an area previously devoid of seismicity (Figure 2.5.1-19).

  - Bakun and Hopper (2004) reinterpreted intensity data from the 1886 Charleston earthquake and show that the calculated intensity center is located about 100 mi offshore from Charleston (although they ultimately concluded that the epicentral location most likely lies onshore in the cluster of seismicity in the Middleton Place–-Summerville area; Figure 2.5.1-19).

  - Crone and Wheeler (2000) described the Helena Banks fault zone as a potential Quaternary tectonic feature (although it was classified as a Class C feature that lacks sufficient evidence to demonstrate Quaternary activity). The occurrence of the 2002 earthquakes and the location of the Bakun and Hopper (2004) intensity center offshore suggest, at a low probability, that the fault zone could be considered a potentially active fault. If the Helena Banks fault zone is an active source, its length and orientation could possibly explain the distribution of paleoliquefaction features along the South Carolina coast.

- Sawmill Branch Fault - Talwani and Katuna (2004) postulated the existence of the Sawmill Branch fault on the basis of microseismicity and further speculated that this feature experienced surface rupture in the 1886 earthquake. According to Talwani and Katuna (2004), this approximately 3-mi-long (5-km-long), northwest-trending fault, which is a segment of the larger Ashley River fault, offsets the Woodstock fault in a left-lateral sense (Figure 2.5.1-19). Earthquake damage at three localities was used to infer that surface rupture occurred in 1886. However, field review of these features along the banks of the Ashley River (small, discontinuous cracks in a tomb that dates to 1671 AD and displacements [less than 10 cm] in the walls of colonial Fort Dorchester) are almost certainly the product of shaking effects as opposed to fault rupture. Moreover, assessment of microseismicity in the vicinity of the proposed Sawmill Branch fault does not clearly define a discrete structure distinct or separate from the larger Ashley River fault, which was defined based on seismicity (Figure 2.5.1-20).
- Summerville Fault - Weems et al. (1997) postulated the existence of the Summerville fault near Summerville, South Carolina, on the basis of previously located microseismicity (Figure 2.5.1-19). However, there is no geomorphic or borehole evidence for the existence of the Summerville fault, and analysis of microseismicity in the vicinity of the proposed Summerville fault does not clearly define a discrete structure (Figure 2.5.1-20).
- Woodstock Fault - The Woodstock fault, a postulated north-northeast-trending, dextral strike-slip fault, was identified by Talwani (1982) on the basis of a linear zone of seismicity located approximately 6 mi west of Woodstock, South Carolina, in the meizoseismal area of the 1886 Charleston earthquake (Figure 2.5.1-19). Madabhushi and Talwani (1990, 1993) used a revised velocity model to relocate Middleton Place–Summerville seismic zone earthquakes, and the results of this analysis were used to further refine the location of the postulated Woodstock fault. Talwani (1999, 2000) subdivided the Woodstock fault into two segments that are offset in a left-lateral sense across the northwest-trending Ashley River fault. Marple and Talwani include the Woodstock fault as part of their larger ZRA (1993) and ECFS (2000).

Charleston Area Seismic Zones

Three zones of increased seismic activity have been identified in the greater Charleston area. These include the Middleton Place–Summerville, Bowman, and Adams Run seismic zones. Each of these features is described in detail below, and the specifics of the seismicity catalog are discussed in Section 2.5.2.1.

- Middleton Place–Summerville Seismic Zone. The Middleton Place–Summerville seismic zone is an area of elevated microseismic activity located about 12 mi northwest of Charleston **(Tarr and Rhea 1983; Bollinger et al. 1991; Madabhushi and Talwani 1993; Talwani and Katuna 2004)** (Figure 2.5.1-19). Between 1980 and 1991, 58 events with Mb 0.8 to 3.3 were recorded in an 11-by-14-square-km area, with hypocentral depths ranging from about 1 to 7 mi (2 to 11 km) **(Madabhushi and Talwani 1993)**. The elevated seismic activity of the Middleton Place–Summerville seismic zone has been attributed to stress concentrations associated with the intersection of the Ashley River and Woodstock faults **(Talwani 1982; Madabhushi and Talwani 1993; Talwani and Katuna 2004; Gangopadhyay and Talwani 2005)**. Persistent foreshock activity was reported in the Middleton Place–Summerville seismic zone area **(Dutton 1889)**, and it has been speculated that the 1886 Charleston earthquake occurred within this zone [e.g., **(Talwani 1982; Tarr and Rhea 1983; Bakun and Hopper 2004)**].

- Bowman Seismic Zone. The Bowman seismic zone is located about 50 mi northwest of Charleston, South Carolina, outside of the meizoseismal area of the 1886 Charleston earthquake and about 60 mi east-northeast of the VEGP site (Figures 2.5.1-18 and 2.5.1-19). The Bowman seismic zone was identified on the basis of a series of $3<M_L<4$ earthquakes that occurred between 1971 and 1974 **(Tarr et al. 1981; Bollinger et al. 1991)**.

- Adams Run Seismic Zone. The Adams Run seismic zone, located within the meizoseismal area of the 1886 Charleston earthquake, was identified on the basis of four $M<2.5$ earthquakes, three of which occurred in a 2-day period in December 1977 **(Tarr and Rhea 1983)**. The Adams Run seismic zone is located about 75 mi east-southeast of the VEGP site (Figure 2.5.1-19). Bollinger et al. (1991) downplayed the significance of the Adams Run seismic zone, noting that, in spite of increased instrumentation, no additional events were detected after October 1979.

<u>Charleston Area Seismically Induced Liquefaction Features</u>

Liquefaction is the transformation of a saturated granular material from a solid to a liquefied state due to increased pore-water pressure. This transformation can result from applied shear stresses during coseismic shaking [e.g., **(Obermeier and Pond 1999)**]; thus, the presence of liquefaction features in the geologic record may be indicative of past earthquake activity in a region. Liquefaction features have been recognized throughout coastal South Carolina and have been attributed to both the 1886 Charleston and earlier moderate to large earthquakes in the region.

- 1886 Charleston Earthquake Liquefaction Features - Liquefaction features produced by the 1886 Charleston earthquake are most heavily concentrated in the meizoseismal area **(Dutton 1889; Seeber and Armbruster 1981; Amick 1990)**, but have been reported as far away as Columbia, Allendale, Georgetown **(Seeber and Armbruster 1981)** and Bluffton, South Carolina **(Talwani and Schaeffer 2001)** (Figure 2.5.1-18).

- Paleoliquefaction Features in Coastal South Carolina - Liquefaction features predating the 1886 Charleston earthquake are found throughout coastal South Carolina (Figures 2.5.1-18 and 2.5.1-19). The spatial distribution and ages of paleoliquefaction features in coastal South Carolina constrain possible locations and recurrence rates for large earthquakes [e.g., **(Obermeier et al. 1985, 1990; Amick 1990; Amcik et al. 1990c, 1990b)**]. Talwani and Schaeffer (2001) combined previously published data with their own studies of liquefaction features in the South Carolina coastal region to derive possible earthquake recurrence histories for the region. Talwani and Schaeffer's (2001) Scenario 1 allows for the possibility that some events in the paleoliquefaction record are smaller in magnitude (approximately M 6+), and that these more moderate events occurred to the northeast (Georgetown) and southwest (Bluffton) of Charleston. In Talwani and Schaeffer's (2001) Scenario 2, all earthquakes in the record are large events (approximately M 7+) located near Charleston. Talwani and Schaeffer (2001) estimated recurrence intervals of about 550 years and approximately 900 to 1,000 years from their two scenarios.

  Because there is very little surface expression of faults within the Charleston seismic zone, earthquake recurrence estimates are based largely on dates of paleoliquefaction events. The most recent summary of paleoliquefaction data **(Talwani and Schaeffer 2001)** suggests a mean recurrence time of 550 years for Charleston, which was used in the 2002 USGS model **(Frankel et al. 2002)**. This recurrence interval is less than the 650-year recurrence interval used in the USGS hazard model **(Frankel et al. 1996)** and is roughly an order of magnitude less than the seismicity-based recurrence estimates used in EPRI (1986). Refinements of the estimate of Charleston area earthquake recurrence are presented in detail in Section 2.5.2.2.2.4.

2.5.1.1.4.5 Savannah River Site Tectonic Features

A number of faults have been identified on the Savannah River Site (SRS), located in South Carolina directly across the Savannah River from the VEGP site (Figures 2.5.1-21, 2.5.1-22, and 2.5.1-23). Fault locations are based on a combination of seismic reflection and refraction studies, borehole studies, and groundwater investigations [e.g., **(Snipes et al. 1993a; Domoracki 1994; Stieve and Stephenson 1995; Cumbest et al. 1998, 2000)**].

The interpreted locations of the SRS faults have changed through time among different researchers and with the availability of additional data (Figures 2.5.1-21, 2.5.1-22, and 2.5.1-23). Because most SRS faults are defined in the subsurface primarily from interpretation of seismic reflection profiles, there is considerable uncertainty regarding the strike, extent, and continuity of some features. Mapping of these subsurface structures between limited data points on seismic profiles has evolved to where some faults defined by name in earlier studies are no longer identified in more recent compilations. For example, the Ellenton fault was initially mapped by Domoracki (1994) as northerly striking fault between the Pen Branch and Crackerneck faults (Figure 2.5.1-21). The Ellenton fault, however, does not appear in the most recent SRS fault maps **(Cumbest et al. 1998, 2000)** (Figure 2.5.1-23).

The most significant perturbations in subsurface basement topography are associated with the Pen Branch, ATTA, Crackerneck, Martin, and Tinker Creek faults (these are the so-called "first order SRS faults" of Cumbest et al. [2000]; Figure 2.5.1-23). Other faults that have been identified or postulated to exist on the SRS include the Ellenton, Lost Lake, Millett, Steel Creek, and Upper Three Runs faults (Figures 2.5.1-21 and 2.5.1-22). Four of the Savannah River faults are located within the VEGP site area (i.e., the Pen Branch, Steel Creek, Ellenton, and Upper Three Runs faults). Each SRS fault is discussed below, beginning with the four faults that lie within the VEGP site area:

- Pen Branch Fault - Because it extends under the VEGP site, the Pen Branch fault is discussed in detail in Section 2.5.1.2.4.
- Steel Creek Fault - The Steel Creek fault is located in the southwest portion of the SRS, about 2.5 mi (4 km) from the VEGP site **(Stieve and Stephenson 1995; Cumbest et al. 2000)** (Figure 2.5.1-21). This northeast-trending fault is located within the Dunbarton Basin and, along with the Pen Branch fault, forms a horst **(Stieve and Stephenson 1995)**. The Steel Creek fault extends upward into Cretaceous units, but the uppermost extent of faulting remains unresolved **(Stieve and Stephenson 1995)**.
- Ellenton Fault - The Ellenton fault is located in the southeastern portion of the SRS, about 5 mi from the VEGP site (Figure 2.5.1-21). The Ellenton fault strikes north-northwest and is near vertical to steeply east-dipping **(Stieve and Stephenson 1995; Cumbest et al. 1998)**. The Ellenton fault appears to have an east-side-down sense of slip, but data quality are poor **(Stieve and Stephenson 1995)**, and the latest mapping does not show this fault **(Cumbest et al. 1998, 2000)**.
- Upper Three Runs Fault - The Upper Three Runs fault is located in the northwest portion of the SRS, about 5 mi from the VEGP site **(Cumbest et al. 2000)** (Figure 2.5.1-21). The northeast-trending Upper Three Runs fault is restricted to crystalline basement; seismic reflection profiling revealed no evidence for this fault offsetting Coastal Plain sediments **(Chapman and DiStefano 1989; Stieve and Stephenson 1995)**. The Upper Three Runs fault has been interpreted as an older (initially Paleozoic) fault that soles into the Augusta fault at depth, possibly reactivated as a Mesozoic normal fault (**Cumbest and Price 1989b; Domoracki 1994; Stieve and Stephenson 1995)**.

- ATTA Fault - The ATTA fault is located in the northeast portion of the SRS, about 16 mi from the VEGP site (Figure 2.5.1-23). The near-vertical ATTA fault strikes approximately N36°E. Based on geometrical analysis of seismic reflection data, the maximum east-side-up vertical separation of basement rocks by the ATTA fault is about 82 ft (25 m) **(Cumbest et al. 2000)**. Upward penetration of the ATTA fault is uncertain because of the lack of good reflectors overlying the fault in the shallow section **(Stieve and Stephenson 1995)**.

- Crackerneck Fault - The Crackerneck fault is located in the northwestern portion of the SRS, about 10 mi from the VEGP site (Figure 2.5.1-23). The Crackerneck fault strikes N22°E and dips steeply to the east **(Cumbest et al. 1998)**. Based on geometrical analysis of seismic reflection data, the maximum vertical separation of basement rocks by the Crackerneck fault is about 98 ft (30 m) **(Cumbest et al. 2000)**. Offset decreases upward out of the Barnwell Group within the Coastal Plain section to 23 ft (7 m) at the top of the Upper Eocene Dry Branch formation (approximately 38.8 Ma) **(Cumbest et al. 2000)**.

- Martin Fault - The Martin fault is located just south of the SRS, about 9 mi from the VEGP site **(Cumbest et al. 2000)** (Figures 2.5.1-22 and 2.5.1-23). There is little subsurface control constraining the location and extent of the Martin fault, but aeromagnetic data anomalies associated with this fault trend N55°E for a distance of about 25 mi **(Cumbest et al. 2000)**. The dip direction of the Martin fault is unknown **(Cumbest et al. 2000)**. Based on data from two boreholes, Snipes et al. (1993a) estimated about 60 to 100 ft of vertical separation of the basement surface associated with the Martin fault. It has been suggested that the Martin fault is the southeastern bounding fault of the Dunbarton Basin **(Snipes et al. 1993a)**, although Domoracki et al. (1999b) suggested that the Dunbarton Basin is instead a half-graben bounded only by the Pen Branch fault to the north.

- Tinker Creek Fault - The Tinker Creek fault is located in the northern portion of the SRS, about 12 mi from the VEGP site **(Cumbest et al. 2000)** (Figure 2.5.1-23). The Tinker Creek fault strikes approximately N36°E and dips steeply to the southeast. Based on geometrical analysis of seismic reflection data, the vertical separation of basement rocks by the Tinker Creek fault increases to the northeast to a maximum of about 79 ft (24 m) **(Cumbest et al. 2000)**. Cumbest et al. (1998) suggested that the Tinker Creek fault may be of regional importance, but the southeastern extent of the fault remains unresolved.

- Lost Lake Fault -. The Lost Lake fault is located in the northwestern portion of the SRS, about 12 mi from the VEGP site **(Cumbest et al. 1989)** (Figure 2.5.1-22). The Lost Lake fault has been mapped based on its apparent control on contaminant flow paths. Seismic reflection and borehole data constraining the location, geometry, sense of slip, and recency of movement on the Lost Lake fault are lacking **(Cumbest et al. 1998)**.

- Millett Fault - A USGS Open File Report by Faye and Prowell (1982) postulated the existence of the Millett and Statesboro faults. Based on the interpretation of (1) lithic fragments in cuttings from two water wells, (2) groundwater data, and (3) changes in Savannah River sinuosity, Faye and Prowell (1982) proposed a 40-mi-long, northeast-striking Millett fault that vertically separates the buried Triassic/Cretaceous contact by 600 ft in a southeast-side-up sense.

  A detailed multidisciplinary study was undertaken by Georgia Power Company (GPC) to investigate the postulated Millett and Statesboro faults **(Bechtel 1982)**. This study included:

  - Geologic mapping
  - Analysis of aerial and landsat imagery
  - Core drilling along two transects
  - Petrographic, X-ray, and heavy minerals analysis of core samples
  - Downhole geophysical surveys
  - Seismic reflection profiling
  - Water level monitoring
  - Groundwater modeling
  - Analysis of surface water flow

  The Bechtel (1982) study concludes that there is no evidence for a capable fault or any fault as young as the undeformed Blue Bluff Marl of mid-Eocene age (40 Ma). This study also demonstrated that the original interpretation of a 600-ft (183-m) vertical separation of the base of the Coastal Plain sediments was incorrect. Core VSC-4, located 200 ft (61 m) from the original well (AL66) where Triassic rock was interpreted in the cuttings, demonstrated the presence of Cretaceous Coastal Plain sediments 400 ft (122 m) below where Triassic rock had originally been interpreted by Faye and Prowell (1982).

  Upon review of the Bechtel (1982) study, the NRC staff concluded that the Millett fault is not a capable fault as defined in Appendix A to 10 CFR Part 100 in the vicinity of the VEGP site. **(Knight, J., 1993)**.

2.5.1.1.4.6 Seismic Sources Defined by Regional Seismicity

Within 200 mi of the VEGP site, there are four areas of concentrated seismicity. Three of these (the Middleton-Place Summerville, Bowman, and Adams Run seismic zones) are located in the

Charleston, South Carolina area and are discussed in Section 2.5.1.1.4.4. The fourth area of concentrated seismicity in the site region is the Eastern Tennessee seismic zone, a northeast-trending concentration of small-to-moderate earthquakes (Figure 2.5.1-15).

Eastern Tennessee Seismic Zone

The Eastern Tennessee seismic zone (ETSZ) is a pronounced seismic source in the central and southeastern United States (Figure 2.5.1-15). Most of the seismicity associated with the Eastern Tennessee seismic zone is located more than 200 mi from the VEGP site, but diffuse seismicity associated with the southeastern margin of the Eastern Tennessee seismic zone is located just within 200 mi of the VEGP site. The zone, located in the Valley and Ridge province of eastern Tennessee, is about 185 mi (300 km) long and 30 mi (50 km) wide and has not produced a damaging earthquake in historical time **(Powell et al. 1994)**. However, this zone is the second most active seismic area in the United States east of the Rocky Mountains, after the New Madrid seismic zone **(Bollinger et al. 1991)**, and produced the second highest release of seismic strain energy in the CEUS during the 1980s, when normalized by crustal area **(Powell et al. 1994)**.

Earthquakes in the Eastern Tennessee seismic zone are occurring at depths from 3 to 16 mi (5 to 26 km) within Precambrian crystalline basement rocks buried beneath the exposed thrust sheets of Paleozoic rocks. None of the Eastern Tennessee seismic zone earthquakes has exceeded a moment magnitude (M) of 4.6 **(Chapman et al. 1997)**. The mean focal depth within the seismic zone is 9 mi (15 km), which is well below the Appalachian basal decollement's maximum depth of 3 mi (5 km) **(Powell et al. 1994)**. The lack of seismicity in the shallow Appalachian thrust sheets implies that the seismogenic structures in the Eastern Tennessee zone are unrelated to the surface geology of the Appalachian orogen **(Johnston et al. 1985)**. The majority of earthquake focal mechanisms show right-lateral slip on northerly-trending planes or left-lateral slip on easterly-trending planes **(Chapman et al. 1997)**. A smaller number of focal plane solutions show right-lateral motion on northeasterly trending planes that parallel the overall trend of seismicity **(Chapman et al. 2002a)**. Statistical analyses of focal mechanisms and epicenter locations suggest that seismicity is occurring on a series of northeast-trending en-echelon basement faults intersected by several east-west-trending faults **(Chapman et al. 1997)**. Potential structures most likely responsible for the seismicity in Eastern Tennessee are reactivated Cambrian or Precambrian normal faults formed during rifting that led to the Iapetan Ocean and presently located beneath the Appalachian thrust sheets **(Bollinger and Wheeler 1988; Wheeler 1995)**.

Earthquakes within the Eastern Tennessee seismic zone cannot be attributed to known faults **(Powell et al. 1994)**, and no capable tectonic sources have been identified within the seismic zone. However, the seismicity is spatially associated with major geophysical lineaments. The western margin of the Eastern Tennessee seismic zone is sharply defined and is coincident with the prominent gradient in the magnetic field defined by the New York-Alabama magnetic lineament **(Chapman et al. 2002a)**. Most seismicity lies between the New York–Alabama lineament on the west and the Clingman and Ococee lineaments on the east **(Johnston et al. 1985)**.

In spite of the observations of small to moderate earthquakes in the Eastern Tennessee seismic zone, no geological evidence has demonstrated the occurrence of prehistoric earthquakes larger than any historical shocks within the seismic zone **(Chapman et al. 2002a; Wheeler 2005)**. Some researchers have suggested secondary evidence for possible large, prehistoric earthquakes in the region [e.g., **(Hatcher et al. 1996)**], but none of this work is conclusive. As a result, Wheeler (2005) classifies the Eastern Tennessee seismic zone as a Class C feature for lack of geological evidence of large earthquakes.

The EPRI source model **(EPRI 1986C)** includes various source geometries and parameters to represent the seismicity of the Eastern Tennessee seismic zone. Subsequent hazard studies have used Mmax values within the range of maximum magnitudes used by the six EPRI models. Collectively, upper-bound maximum values of Mmax used by the EPRI teams ranged from mb 6.6 to 7.4. Using three different methods specific to the Eastern Tennessee seismic source, Bollinger (1992) estimated a Mmax of mb 6.45. Chapman and Krimgold (1994) used a Mmax of mb 7.25 for the Eastern Tennessee zone and most other sources in their seismic hazard analysis of Virginia. Both of these more recent estimates of Mmax are similar to the range of Mmax values used in EPRI (1986). Therefore, it is concluded that no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model. Additional discussion of the significance of the ETSZ on the Vogtle ESP seismic hazard is provided in Section 2.5.2.2.2.5.

<u>Selected Seismogenic and Capable Tectonic Sources Beyond the Site Region</u>

Because of the potential for distant, large earthquakes in the CEUS contributing to the long period ground motion hazard at the VEGP site, a discussion of three additional seismic sources located beyond 200 mi from the site is warranted. These sources are the Central Virginia, New Madrid, and Giles County seismic zones.

Central Virginia Seismic Zone

The Central Virginia seismic zone is an area of persistent, low-level seismicity in the Piedmont province, located more than 350 mi from the VEGP site (Figure 2.5.1-15). The zone extends about 75 mi in a north-south direction and about 90 mi in an east-west direction from Richmond to Lynchburg **(Bollinger and Sibol 1985)**. The largest historical earthquake to occur in the Central Virginia seismic zone was the body-wave magnitude (m_b) 5.0 Goochland County event on December 23, 1875 **(Bollinger and Sibol 1985)**. The maximum intensity estimated for this event was MMI VII in the epicentral region.

Seismicity in the Central Virginia seismic zone ranges in depth from about 2 to 8 mi (4 to 13 km) **(Wheeler and Johnston 1992)**. Coruh et al. (1988) suggested 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 (4 to 13 km) **(Wheeler and Johnston 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 paleoliquefaction sites have been identified within the seismic zone **(Crone and Wheeler 2000; Obermeier and McNulty 1998)**. The paleoliquefaction sites reflect prehistoric occurrences of seismicity within the Central Virginia seismic zone and do not indicate the presence of a capable tectonic source.

The 1986 EPRI source model includes various source geometries and parameters to capture the seismicity of the Central Virginia seismic zone **(EPRI 1986C)**. Subsequent hazard studies have used 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 ESTs range from m_b 6.6 to 7.2 (discussed in Section 2.5.2.2). More recently, Bollinger (1992) has estimated a Mmax of m_b 6.4 for the Central Virginia seismic source. Chapman and Krimgold (1994) have used a Mmax of m_b 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 EPRI (1986). Similarly, the distribution and rate of seismicity in the Central Virginia seismic source have not changed since the 1986 EPRI study (discussed in Section 2.5.2.2.8). Thus, there is no change to the source geometry or rate of seismicity. 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.

New Madrid Seismic Zone

The New Madrid seismic zone extends from southeastern Missouri to southwestern Tennessee and is located more than 400 mi west of the VEGP site Figure 2.5.1-15). The New Madrid seismic zone lies within the Reelfoot rift and is defined by post-Eocene to Quaternary faulting and historical seismicity. Given the significant distance between the site and the seismic zone, the New Madrid seismic zone did not contribute to 99 percent of the hazard at the VEGP site in

EPRI (1986). However, it is described in this section because several recent studies provide significant new information regarding magnitude and recurrence interval for the seismic zone.

The New Madrid seismic zone is approximately 125 mi (220 km) long and 25 mi (40 km) wide. Research conducted since 1986 has identified three distinct fault segments embedded within the seismic zone. These three fault segments include a southern northeast-trending dextral slip fault, a middle northwest-trending reverse fault, and a northern northeast-trending dextral strike-slip fault **(Wheeler and Crone 2001)**. In the current east-northeast to west-southwest directed regional stress field, Precambrian and Late Cretaceous age extensional structures of the Reelfoot rift appear to have been reactivated as right-lateral strike-slip and reverse faults.

The New Madrid seismic zone produced three historical, large magnitude earthquakes between December 1811 and February 1812 **(Hough et al. 2000)**. The December 16, 1811, earthquake is associated with strike-slip fault displacement along the southern portion of the New Madrid seismic zone. Johnston (1996) estimated a magnitude of **M** 8.1±0.31 for the December 16, 1811, event. However, Hough et al. (2000) re-evaluated the isoseismal data for the region and concluded that the December 16 event had a magnitude of **M** 7.2 to 7.3. Bakun and Hopper (2004) similarly concluded this event had a magnitude of **M** 7.2.

The February 7, 1812, New Madrid earthquake is associated with reverse fault displacement along the middle part of the New Madrid seismic zone **(Johnston and Schweig 1996)**. This earthquake most likely occurred along the northwest-trending Reelfoot fault that extends approximately 43 mi from northwestern Tennessee to southeastern Missouri. The Reelfoot fault is a northwest-trending, southwest-vergent reverse fault. The Reelfoot fault forms a topographic scarp developed as a result of fault-propagation folding **(Van Arsdale et al. 1995; Kelson et al. 1996; Van Arsdale 2000)**. Johnston (1996) estimated a magnitude of **M** 8.0±0.33 for the February 7, 1812, event. However, Hough et al. (2000) re-evaluated the isoseismal data for the region and concluded that the February 7 event had a magnitude of **M** 7.4 to 7.5. More recently, Bakun and Hopper (2004) estimated a similar magnitude of **M** 7.4.

The January 23, 1812, earthquake is associated with strike-slip fault displacement on the East Prairie fault along the northern portion of the New Madrid seismic zone. Johnston (1996) estimated a magnitude of **M** 7.8±0.33 for the January 23, 1812, event. Hough et al. (2000), however, re-evaluated the isoseismal data for the region and concluded that the January 23 event had a magnitude of **M** 7.1. More recently, Bakun and Hopper (2004) estimated a similar magnitude of **M** 7.1.

Because there is very little surface expression of faults within the New Madrid seismic zone, earthquake recurrence estimates are based largely on dates of paleoliquefaction and offset geological features. The most recent summaries of paleoseismologic data **(Tuttle et al. 2002,**

2005; Guccione 2005) suggest a mean recurrence time of 500 years, which was used in the 2002 USGS model **(Frankel et al. 2002)**. This recurrence interval is half of the 1,000-year recurrence interval used in the 1996 USGS hazard model **(Frankel et al. 1996)**, and an order of magnitude less than the seismicity-based recurrence estimates used in EPRI (1986).

The upper-bound maximum values of Mmax used in EPRI (1986) range from m_b 7.2 to 7.9. Since the EPRI study, estimates of Mmax have generally been within the range of maximum magnitudes used by the six EPRI models. The most significant update of source parameters in the New Madrid seismic zone since the 1986 EPRI study is the reduction of the recurrence interval to 500 years.

Giles County Seismic Zone

The Giles County seismic zone is located in Giles County, southwestern Virginia, near the border with West Virginia and more than 250 mi from the VEGP site (Figure 2.5.1-15). The largest known earthquake to occur in Virginia and the second largest earthquake in the entire southeastern United States is the 1897 M 5.9 **(Johnston et al. 1994)** Giles County event, which likely produced an MMI VIII in the epicentral area.

Earthquakes in the Giles County seismic zone occur within Precambrian crystalline basement rocks beneath the Appalachian thrust sheets at depths from 3 to 16 mi (5 to 25 km) **(Bollinger and Wheeler 1988)**. Earthquake foci define a 25-mi-long (40-km-long), northeasterly striking, tabular zone that dips steeply to the southeast beneath the Valley and Ridge thrust sheets **(Bollinger and Wheeler 1988; Chapman and Krimgold 1994)**. The lack of seismicity in the shallow Appalachian thrust sheets, estimated to be about 2 to 3.5 mi (4 to 6 km) thick, implies that the seismogenic structures in the Giles County seismic zone, similar to those inferred for the Eastern Tennessee seismic zone, are unrelated to the surface geology of the Appalachian orogen **(Bollinger and Wheeler 1988)**. The spatial distribution of earthquake hypocenters, together with considerations of the regional tectonic evolution of eastern North America, suggests that the earthquake activity is related to contractional reactivation of late Precambrian or Cambrian normal faults that initially formed during rifting associated with opening of the Iapetan Ocean **(Bollinger and Wheeler 1988; Bollinger et al. 1991)**.

No capable tectonic sources have been identified within the Giles County seismic zone, nor does the seismic zone have recognizable geomorphic expression **(Wheeler 2005)**. Thus, in spite of the occurrence of small to moderate earthquakes, no geological evidence has demonstrated the occurrence of prehistoric earthquakes larger than any historical shocks within the zone **(Wheeler 2005)**. As a result, Wheeler (2005) classifies the Giles County seismic zone as a Class C feature for lack of geological evidence of large earthquakes.

A zone of small Late Pliocene to Early Quaternary age faults has been identified within the Giles County seismic zone, near Pembroke, Virginia **(Crone and Wheeler 2000)**. The Pembroke zone is a set of extensional faults exposed in terrace deposits overlying limestone bedrock along the New River. Crone and Wheeler (2000) rated these faults as Class B features because it has not yet been determined whether these faults are tectonic or the result of solution collapse in underlying limestone units. The shallow Pembroke faults do not appear to be related to the seismicity within the Giles County seismic zone, which is occurring beneath the Appalachian basal decollement in the North American basement.

The EPRI source model includes various source geometries and parameters to represent the seismicity of the Giles County seismic zone **(EPRI 1986C)**. Subsequent hazard studies have used Mmax values that were within the range of maximum magnitudes used by the six EPRI models. Collectively, upper-bound maximum values of Mmax used by the EPRI teams ranged from m_b 6.6 to 7.2 (discussed in Section 2.5.2.2). More recently, Bollinger (1992) estimated an Mmax of m_b 6.3 for the Giles County seismic source using three different methods. Chapman and Krimgold (1994) used a Mmax of m_b 7.25 for the Giles County zone and most other sources in their seismic hazard analysis of Virginia. Both of these more recent estimates of Mmax are similar to the range of Mmax values used in EPRI (1986). Therefore, no new information has been developed since 1986 that would require a significant revision to the EPRI seismic source model.

2.5.1.1.5 Regional Gravity and Magnetic Data

Regional maps of the gravity and magnetic fields in North America were published by the Geological Society of America in 1987 as part of the Society's Decade of North American Geology (DNAG) project. The maps present the potential field data at 1:5,000,000-scale and thus are useful for identifying and assessing regional gravity and magnetic anomalies with wavelengths on the order of about 6 mi (10 km) or greater. Gravity and magnetic data also have been incorporated in the E-5 DNAG crustal transect, which traverses the Appalachian orogen from eastern Tennessee to the offshore Atlantic basin **(Hatcher et al. 1994)**, and encompasses the VEGP site region. At a local scale, Cumbest et al. (1992) developed models of gravity and magnetic data to evaluate the geometry and structure of the Mesozoic Dunbarton basin beneath the SRS northeast of the VEGP site. These models in particular provide important insights for the interpretation of potential field data in the VEGP site region.

2.5.1.1.5.1 Regional Gravity Data

The 1987 DNAG gravity map and the gravity profile along the E-5 DNAG crustal transect document low gravity values beneath the Valley and Ridge and Blue Ridge provinces relative to the Cumberland Plateau province to the west (Figure 2.5.1-11). The approximately 40 to

60 mgal eastward decrease in gravity along the western margin of the Valley and Ridge province is likely due to eastward thickening of the relatively less dense carbonate and siliciclastic rocks above the Grenville metamorphic basement in the Valley and Ridge province relative to the Cumberland Plateau.

Bouguer gravity values increase by about 80 mgal across an approximately 62 mi (100 km) distance from the eastern Blue Ridge to the Inner Piedmont Terrane **(Committee for Gravity Anomaly Map of North America 1987; Hatcher et al. 1994)**. As documented by the DNAG gravity map, this gradient is present across the Piedmont physiographic province along much of the length of the Appalachian belt. At the latitude of Virginia, north of the ESP study region, this gradient has been interpreted to reflect the eastward thinning of the North American continental crust and associated positive relief on the Moho with proximity to the Atlantic margin. Inspection of the crustal structure interpreted by Hatcher et al. (1994) along the E-5 transect (Figure 2.5.1-11) indicates that the eastward increase in gravity across the Inner Piedmont at the latitude of the ESP study region also is associated with obducted metavolcanic rocks of the Carolina–Avalon Terrane that have been overthrust onto the Grenvillian basement. A gravity model by Iverson and Smithson (1983) along the southern Appalachian COCORP seismic reflection profile suggests that the gradient probably arises from both eastward thinning of continental crust and the obduction of the Inner Piedmont and Carolina Terranes, which have higher average densities than the underlying Grenvillian crust.

The gravity profile along the DNAG E-5 crustal transect indicates that the gravity field east of the Inner Piedmont is relatively uniform (Figure 2.5.1-11), and the 1987 DNAG gravity map shows that the Coastal Plain is characterized by relatively low amplitude anomalies with wavelengths on the order of about 12 to 25 mi (20 to 40 km) superimposed on this uniform field. Detailed modeling of the gravity field in this region by Cumbest et al. (1992) indicates that the most prominent anomalies are associated with monzogranite plutons, which are relatively less dense than the intruded country rock and thus give rise to local gravity lows, and mafic intrusions, which are relatively more dense and give rise to local gravity highs. From modeling of gravity data, Cumbest et al. (1992) found that the predicted anomaly associated with the Mesozoic Dunbarton Basin is a subordinate feature of the gravity field compared to the anomalies associated with the plutons and mafic intrusions.

The relationship between gravity anomalies in the VEGP site region and bedrock geology inferred by Dennis et al. (2004) to underlie the Coastal Plain sediments from borehole and other subsurface data is illustrated in Figure 2.5.1-24. The extremes in the local gravity field are highs associated with Triassic-Jurassic mafic intrusive complexes southeast of the VEGP site and lows associated with granitic plutons mapped to the north-northeast and east-northeast of the site. The lateral extent of the gravity lows associated with the Graniteville and Springfield plutons suggest that these bodies may be larger than inferred by Dennis et al. (2004). The

Dunbarton Basin is spatially associated with an approximately 5-mi-wide (8-km-wide), northeast-southwest-trending gravity low northwest of the mafic intrusive complexes and associated gravity high southeast of the VEGP site. A northwest-southeast profile of the gravity data through the VEGP site (Figure 2.5.1-25) illustrates in detail that the gravity low associated with the Dunbarton Basin is a very modest second-order feature superimposed on the 25-mi-long (40-km-long) west-to-east increase in gravity between the granitic plutons and mafic intrusive complexes, consistent with the findings of Cumbest et al. (1992).

The gravity profile also shows that the Belair fault, which separates the Kiokee belt on the northwest from the Belair belt to the southeast, is adjacent to an approximately 35 mgal gravity gradient (Figure 2.5.1-25). This eastward decrease in gravity shows that the high-grade metamorphic rocks of the Kiokee belt are generally denser than the relatively lower-grade rocks of the Belair belt. However, the magnitude of the gradient also could be affected by the presence of the relatively lower density granitic plutons to the east and southeast of the Belair fault.

To summarize, gravity data published since the mid-1980s document that long-wavelength anomalies along the E-5 DNAG crustal transect through the VEGP site region are characteristic of large parts of the Appalachian belt and reflect first-order features of the various provinces and accreted Paleozoic terranes and west-to-east thinning of the ancestral North American (Grenvillian) continental crust during the Mesozoic. The dominant short-wavelength characteristics of the gravity field in the vicinity of the VEGP site are gravity highs and lows associated with Mesozoic mafic and Paleozoic granitic intrusions, respectively. Detailed gravity modeling by Cumbest et al. (1992) shows that the gravity low associated with the Triassic Dunbarton Basin is a subordinate feature in the regional field. The gravity data acquisition and modeling studies performed to date do not show any evidence for Cenozoic tectonic activity or specific Cenozoic structures. There are no large, unexplained anomalies in the gravity data.

2.5.1.1.5.2 Regional Magnetic Data

Data compiled for the DNAG magnetic map reveal numerous northeast-southwest-trending magnetic anomalies that are generally parallel to the structural grain of the Paleozoic Appalachian orogenic belt **(Committee for Magnetic Anomaly Map of North America 1987)**. For example, a magnetic profile along the DNAG E-5 crustal transect **(Hatcher et al. 1994)** reveals an approximately 800 nT southeastward decrease in magnetic intensity between the Cumberland Plateau and western Valley and Ridge provinces, and the western Blue Ridge province (Figure 2.5.1-11). The DNAG magnetic map indicates that this gradient is present to the southwest and northeast along the western Appalachian belt in adjacent parts of Alabama and Kentucky and is spatially associated with the contact between Precambrian metamorphic basement and overlying Paleozoic accreted terranes. In general, the western Valley and Ridge province and eastern Blue Ridge along the E-5 crustal transect are characterized by relatively low magnetic intensities, and the western Blue Ridge and Inner Piedmont are relative magnetic highs, probably indicating a greater abundance of mafic rocks in the accreted Taconic units east of the Hayesville thrust.

The accreted Carolina–Avalon Terrane is characterized by short-wavelength, high-amplitude anomalies (approximately 200 nT over distances of about 6.2 mi [10 km]). Detailed modeling of magnetic data from the SRS northeast of the VEGP site indicates that these anomalies may be associated with mafic intrusions that are vertically elongated and have east-dipping boundaries **(Cumbest et al. 1992)**. Felsic plutons in this region, which are inferred to exist from borehole data and gravity modeling, have modest susceptibility contrasts with the country rock they intrude and thus do not generate high-amplitude magnetic anomalies **(Cumbest et al. 1992)**. Similarly, Mesozoic basin sediments are inferred to have relatively low susceptibility contrasts with the pre-intrusive basement rock, and modeling by Cumbest et al. (1992) suggests that the anomaly associated with the sediments and margins of the Dunbarton Basin is a second-order feature of the magnetic field relative to the amplitudes of the anomalies produced by the intrusive mafic rocks. The Towaliga fault along the western margin of the Carolina–Avalon Terrane is associated with alternating low and high short wavelength magnetic anomalies **(Hatcher et al. 1994)** characteristic of those produced by a susceptibility contrast across a dipping structural contact.

Comparison of aeromagnetic data from the VEGP site region **(Daniels 2005)** with bedrock geology inferred by Dennis et al. (2004) to underlie the Coastal Plain sediments from borehole and other subsurface data illustrates the relationships described by Cumbest et al. (1992). The Dunbarton Basin is associated with a northeast-southwest-trending magnetic low and is bounded on the south by pronounced magnetic highs associated with Triassic–Jurassic mafic intrusive complexes (Figure 2.5.1-26). A northwest-southeast-trending profile of the magnetic intensities that passes through the VEGP site shows that the magnetic low associated with the Dunbarton Basin is similar to a magnetic low approximately 12 mi to the southeast that is not associated with a known Triassic basin (Figure 2.5.1-27); also, the figure shows that the high magnetic anomalies associated with the mafic intrusive complexes extend northward into the basin (Figure 2.5.1-26). These relations are consistent with the conclusion of Cumbest et al. (1992) that the magnetic signature of the Dunbarton Basin is very modest relative to that of the mafic intrusive complexes.

The magnetic map and profile of the greater VEGP site region **(Daniels 2005)** also include the Kiokee and Belair belts of the Carolina arc terrane northwest of the VEGP site. Both of these belts are characterized by closely spaced, short-wavelength anomalies with amplitudes of about 100 to 200 gammas (Figure 2.5.1-11). In general, the magnetic intensities in the Belair belt are slightly higher than those of the Kiokee belt. Daniels (1974) noted that the bulk composition of the Kiokee belt is probably more felsic than that of the Belair belt and thus has a lower magnetic susceptibility.

To summarize, magnetic data published since the mid-1980s provide additional characterization of the magnetic field in the VEGP site region. Detailed modeling of magnetic data from the SRS provides insights into the origins of magnetic anomalies that extend southwest into the vicinity of the VEGP site. The first-order magnetic anomalies in the VEGP site region are associated primarily with northeast-southwest-trending Paleozoic and Mesozoic intrusive rock bodies. The magnetic data do not show evidence for any Cenozoic structures in the site region and do not have sufficient resolution to identify or map discrete faults, such as border faults along the Dunbarton Basin. No large, unexplained anomalies are found in the magnetic data.

2.5.1.2 Site Area Geology

This section describes the geology and structural geology of the site area (within a 5-mi radius of the VEGP site).

2.5.1.2.1 Site Area Physiography and Geomorphology

The site area lies in the Upper Coastal Plain of the Coastal Plain Physiographic Province and is bordered by the Savannah River to the east (Figures 2.5.1-1 and 2.5.1-28). The surrounding topography consists of gently rolling hills with a principally dendritic drainage pattern. Surficial soils are typically well drained. All major streams are tributary to the Savannah River.

The site area lies at the northern extent of a broad westward migrating meander in the Savannah River where the sinuosity decreases from about 1.8 to 1.3 **(Geomatrix 1993)**. Incision of the river has formed steep bluffs and topographic relief of nearly 150 ft from the river surface to the plant site. The river level adjacent to the plant site is at an elevation of approximately 80 ft msl, with a gradient of less than 1 ft/mi **(Geomatrix 1993)**. The flood plain is a broad alluvial surface that is 4 to 10 ft above the channel. The youngest alluvium lies along the western side of the river, while older terraces are preserved on the east side of the flood plain (Figures 2.5.1-29 and 2.5.1-31). Stream valleys are predominantly symmetrical, with slopes ranging from 0.2 to 0.6 percent. The surface topography ranges from an elevation of about 200 to nearly 300 ft msl across the site area (Figures 2.5.1-30 and 2.5.1-32).

Several surface depressions were noted during the initial site investigation and were extensively studied. The topography and surface drainage within the site has been modified during and after construction of the existing VEGP units, making evaluation of these previously mapped features impractical. In the Coastal Plain, surface depressions can be categorized as eolian features known as Carolina bays, whereas others may result from solution of underlying calcareous sediment **(Siple 1967; USACE 1952; Smith 1931)**.

Carolina bays, which are shallow, elliptical depressions with associated sand rims, are common throughout the Atlantic Coastal Plain and are most numerous in North and South Carolina. They are surficial features that have no effect on the subsurface sediments. Distinquishing features of Carolina bays are the elliptical shape, preferential orientation of the long axis at S50°E, and sand rims along the east and southeast flanks **(Johnson 1942)**. A discussion of various hypotheses for the timing and mode of origin of these bays is provided in Section 2.5.1.1.1.

Surface depressions that do not meet the criteria of Carolina bays are typically irregularly shaped, localized features resulting most likely from the dissolution of calcareous stratum at depth. Initial site studies conducted for the existing VEGP Units 1 and 2 concluded that these features resulted principally from dissolution of a limestone unit and that lower-lying carbonate-bearing units were not involved. Section 2.5.3.8.2.1 contains a discussion of the significance of these features.

2.5.1.2.2 Site Area Geologic History

The Upper Coastal Plain is essentially a flat-lying section of unconsolidated fluvial and marine sediments overlying a basement complex of Paleozoic crystalline metamorphic and igneous rock as well as Triassic–Jurassic basin sediments. Evolution of the basement complex and the effect on the Coastal Plain section is regional in nature and is discussed in Sections 2.5.1.1.2 and 2.5.1.1.4.

The Paleozoic rocks and the Triassic sediments were beveled by erosion, forming the base for Coastal Plain sediment deposition. The erosional surface dips southeast approximately 50 ft/mi **(Fallaw and Price 1995)**. The Coastal Plain section consists of stratified sand, clay, limestone, and gravel that dip gently seaward. The oldest Coastal Plain sediments beneath the site area are Late Cretaceous and consist of predominantly siliciclastics deposited in an upper deltaic, fluvial setting that continued throughout the Cretaceous. Paleocene sedimentation continued, with a strong fluvial influence giving way to more marginal marine to shallow shelf deposition well into the Middle Eocene, marked by deposition of mixed clastic-carbonate sediments. Upper Eocene sedimentation occurred in more marginal and inner-tidal settings. Miocene (or younger) high energy fluvial deposits are present at higher topographic locations and in some areas incised deeply into the underlying Eocene section. A thin veneer of late Miocene to early Pliocene eolian sands overlies some of the higher topographic areas. The youngest sediments consist of Quaternary alluvium present within the stream and river valleys.

2.5.1.2.3 Site Area Stratigraphy

The site area stratigraphy is based on site-specific data as well as regional geologic studies and includes the following sources of information:

- Regional geologic maps, studies, and related data
- Site area studies performed for VEGP
- Borehole data, including core and geophysical logs acquired during the ESP investigation (Figure 2.5.1-33)
- Surface geophysical surveys, including seismic reflection and refraction (Figures 2.5.1-34, 2.5.1-35, 2.5.1-36, and 2.5.1-37).

Numerous geologic studies have been conducted in the surrounding area since initial studies were conducted for VEGP Units 1 and 2. Most of these studies were focused in the vicinity of the SRS. Many of these studies focused on correlating both geologic and hydrogeologic formations present in South Carolina and Georgia, resulting in an updated stratigraphic nomenclature. The most current stratigraphic nomenclature from Huddleston and Summerour (1996) and Falls and Prowell (2001) is used below. A correlation chart showing current USGS, Georgia Geological Survey, South Carolina, and SRS and VEGP Units 1 and 2 FSAR nomenclature is provided as Figure 2.5.1-8. A site stratigraphic column based on data from borehole B-1003 is shown on Figure 2.5.1-38.

2.5.1.2.3.1 Basement Rock

Regionally, the basement surface has been leveled by erosion and dips to the southeast approximately 50 ft/mi **(Fallaw and Price 1995)**. Basement rock lithology within the site area consists of Paleozoic crystalline rock as well as Triassic–Jurassic sedimentary rock of the Dunbarton Basin. Basement rock lithology has been determined directly from core data from boring B-1003 and inferred from seismic reflection and refraction surveys performed as part of the ESP investigation. These data are corroborated regionally with other core data and geophysical surveys, as discussed in Section 2.5.1.1.3.5.

Boring B-1003 was drilled within the VEGP site area to acquire detailed stratigraphic, lithologic, geophysical (including natural gamma, electrical resistivity, compressional velocity, and shear wave velocity) and depth-to-basement data. Data from B-1003 identifies Triassic–Jurassic basement rock at a depth of 1,049 ft (-826 ft msl). Data from four seismic reflection and refraction lines described in Section 2.5.1.2.4.2, as well as borehole and seismic reflection data from other regional studies including the SRS **(Cumbest et al. 1992; Snipes et al. 1993a)**, determine the northern boundary of the Dunbarton Basin to strike northeast-southwest across the site area, defining the approximate boundary between the Triassic–Jurassic sedimentary rock underlying the southeastern portion and Paleozoic crystalline rock underlying the northwestern portion of the site area (Figure 2.5.1-39).

Although no borehole data confirm the lithology of the Paleozoic crystalline rock with the site area, data from regional studies, as well as regional gravity and magnetic surveys, suggest a complex of metavolcanics **(Cumbest et al. 1992; Snipes et al. 1993a)**. The Triassic–-Jurassic sedimentary rocks of the Dunbarton Basin consist of mudstones, sandstones, and conglomerates of varying degrees of lithification, as determined from borehole B-1003.

2.5.1.2.3.2 Site Area Coastal Plain Stratigraphy

The Paleozoic and Triassic basement complex is unconformably overlain by poorly consolidated to unconsolidated Coastal Plain sediments that dip and thicken to the southeast. These sediments range in age from Upper Cretaceous to Miocene except where the Miocene Hawthorn Formation has been removed by excavation and are approximately 1,049 ft thick in the site area, based on boring B-1003 that was drilled as part of the ESP investigation.

The stratigraphy defined for the site area adopts the nomenclature of Huddleston and Summerour (1996), as shown on Figures 2.5.1-8, and 2.5.1-38, and was based primarily on lithology except where carbonate fossils provided more definitive stratigraphic correlation. Cretaceous sediments that had been assigned to the Tuscaloosa Formation were assigned to the Cape Fear, Pio Nono, Gaillard/Black Creek and Steel Creek formations. Tertiary sediments were further subdivided based on both lithology and carbonate fossils where present. The youngest sediments of Quaternary age consisted of alluvial deposits within stream and river valleys.

More recent investigations in the vicinity of the site area have included detailed lithological and paleontological studies to correlate stratigraphic units between Georgia and South Carolina **(Falls and Prowell 2001)**. Over the last two decades, detailed work that focused on the SRS (Figure 2.5.1-8), has resulted in a more detailed lithostratigraphic column. Cross-well correlation using lithologic data from both the core and down-hole geophysical logs provides a means to correlate many of the same geologic units present at the site area with those mapped on a more regional scale.

The following sub-sections describe each geologic unit, from oldest to youngest, and are based primarily on lithologic descriptions of the core log from boring B-1003 drilled and logged as part of the ESP site investigation (Appendix 2.5A) (Figure 2.5.1-38). The most recent stratigraphic column published by the USGS **(Falls and Prowell 2001)** is cited where those studies provide confirmatory information that is directly applicable to the site area. In addition, geologic studies and correlation with SRS stratigraphic units are cited where they provide confirmatory information directly applicable to the site area.

Cretaceous Stratigraphy

Upper Cretaceous age sediments unconformably overlie both crystalline and Triassic–Jurassic basement rock in the site area. The initial site stratigraphy for the VEGP Units 1 and 2 assigned all Cretaceous sediments to the Tuscaloosa Formation. More recent investigations have identified four geologic formations within the Cretaceous section. The following discussions rely primarily on core log data from boring B-1003 (Appendix 2.5A). Contacts, as interpreted from geophysical well logs might vary from the depths identified on the boring logs. The following sections describe these units from oldest to youngest.

Cape Fear Formation

In boring B-1003, the base of the Cape Fear Formation was determined to be at a 1,049 ft depth (-826 ft msl) and the top at a depth of 858 ft (-635 msl). This results in a thickness of 191 ft. The base of the Cape Fear Formation in Boring B-1003 consists of sandy silt overlain by well sorted quatz gravel with subrounded to angular grains. This is overlain by generally well sorted gray sandy silt with layers of subrounded gravels, pebbles and silty sand. The silty sand is overlain by sandy silt that in turn is overlain by a clayey sand granding up into a sandy elastic clay. The top portion of the Cape Fear formation is logged as a gray to dark gray, well sorted, subrounded to subangular fine to coarse clays sand with a weak cementation zone near the top and arkosic layers.

In nearby boreholes, the Cape Fear Formation consists of poorly sorted, silty to clayey quartz, occasionally arkosic, sands with interbedded clays. Grains tend to be subangular to angular. The sands are medium to coarse, with occasional pebble zones. Lithification ranges from moderate to high due to the presence of cristobalite as a clay matrix, which can also yield a greenish blue hue to the sediments. Numerous stacked fining-upward sequences, lack of marine fossils, and the presence of terrestrial microflora and root clasts suggest deposition within a fluvial dominated delta plain **(Falls and Prowell 2001)**.

Pio Nono Formation

In boring B-1003 the base of the Pio Nono Formation is logged at a depth of 858 ft (-635 msl) and the top is logged at a depth of 798 ft (-575 msl). The base consists of a light gray, well sorted clayey sand. This is overlain primarily by well sorted sand with silt that grades upward from a gray-tan fine to medium grained unit into a white-gray fine to coarse unit with some gravel. This formation consists primarily of moderately to well sorted quartz sands with little silt and clayey sands. Traces of manganese staining and mica are present. Grains tend to be rounded to subangular, medium to coarse, with some gravel lags. The sands are typically non-lithified, with a few slightly cemented zones. The top of the formation is logged as a tan-white poorly sorted fine to medium grained sand with silt. A fining-up sequence was logged between depths of 808 and 803 ft.

Both fining and coarsening upward sequences are noted in nearby boreholes. The top of the formation may be marked by a thick bed of oxidized clay. In borehole B-1003, this clay was logged as marking the base of the overlying Galliard/Black Creek formations. The lack of marine fossils, presence of oxidized zones, and gravel lags suggest a deltaic environment.

Upper Gaillard Formation/Black Creek Formation

The Upper Gaillard/Black Creek Formation consists of thick alternating sequences of moderately to well sorted silty, clayey sands and silty clay beds. The sands tend to be medium to coarse grained, rounded to subrounded, and contain trace amounts of mica and glauconite. Fining-upward sequences are present, as well as gravel lag deposits. The more clayey beds tend to be dark and oxidized, with trace amounts of lignite and root clasts. The presence of marine fauna and glauconite suggests a marine influence, while oxidation in the clay beds and the presence of root clasts suggest a more lagoonal setting, suggesting a prograding delta sequence.

The base of the Upper Gaillard/Black Creek Formation in boring B-1003 was noted at a 798-ft depth (-575 ft msl) and the top at a 587-ft depth (-364 ft msl), resulting in an overall thickness of 211 ft. As noted above, the base is logged as a 12-ft layer of black to dark gray clay. The clay is overlain by poorly graded silty sands to sandy silts. A 24-ft layer of gray sandy silt with clayey sand layers occurs between depths of 603 ft to 627 ft. The top of the Gaillard/Black Creek Formation consists of well sorted gray fine to coarse sand with silt.

Steel Creek Formation

The basal contact of the Steel Creek Formation with the underlying upper Gaillard/Black Creek Formation is at a depth of 587 ft (-364 ft msl). The top of the Steel Creek Formation is noted at a depth of 477 ft (-254 ft msl), resulting in a total thickness of 110 ft.

Sediments of the Steel Creek Formation are predominantly sands and silty sands. These sands are well to poorly sorted, with trace amounts of mica, kaolin, and lignite. Sand grains are subrounded to subangular. Multiple fining-upward sequences, the presence of lignite, and the oxidation of clays indicate a more fluvial, delta plain depositional setting. The top of the Steel Creek Formation in boring B-1003 is marked by a transistion from gray poorly sorted fine to coarse sand with a kaolinitc clay matrix to a clayey sand of the ovelying Black Mingo Formation.

Tertiary Stratigraphy

Tertiary sediments ranging from Paleocene to Miocene age unconformably overlie the Cretaceous section in the site area. The site stratigraphy defined for the VEGP UFSAR divided the Tertiary section into the Ellenton Formation, Huber Formation, Lisbon Formation, Barnwell Group, and Hawthorne Formation. Further subdivisions were made for the Lisbon and Barnwell Group because these units are more easily mapped due to exposure within incised valleys as well as to more available borehole data units due to the relative shallow depth of these units. The Tertiary section also contains considerably more calcareous sediments, thus providing more biostratigraphic constraint.

More recent investigations, including detailed palynotologic and paleontologic studies, have refined the Tertiary stratigraphy in the vicinity of the site area **(Fallaw and Price 1995; Falls and Prowell 2001)**. Huddelston and Summerour (1996) divide the Tertiary units, from oldest to youngest, into the Black Mingo, Snapp and Congaree formations, the Bennock Millpond/Still Branch Sand, the Lisbon Formation and the Barnwell Group. As with the Cretaceous section, core log data from boring B-1003 are used to describe the site stratigraphy. The following sections describe these units from oldest to youngest.

Black Mingo Formation

In boring B-1003, the base of the Black Mingo Formation was noted at a depth of 477 ft (-254 ft msl) and the top at a depth of 438 ft (-215 ft msl), giving an overall thickness of 39 ft. The base of the unit is marked by the occurrence of a gray, poorly sorted fine to coarse sand and is overlain by a 15-ft thick layer of gray sandy clay on some coarse subangular quartz sand and lignite fragments. This, in turn is overlain by light gray well sorted fine to medium clayey sand and sand with silt. The top of the Black Mingo Formation is logged as a 7-ft thick layer of dark gray sandy clay.

Snapp Formation

The base of the Snapp Formation in boring B-1003 is logged at a depth of 438 ft (-215 ft msl) and the top at a depth of 331 ft (-108 ft msl), resulting in an overall thickness of 107 ft. The base is marked by well sorted sand with quartz gravel that appears to be a channel lag deposit. This grades up into a tan-gray well sorted silty fine to medium grained sand that is overlain by 18 ft of gray-red silty clay containing some 4- to 6-inch layers of white and gray fine to coars silty quartz sand. The clay is overlain by well sorted clayey sand that is overlain by a sandy to silty gray to red-brown clay. This 12-ft clay layer is overlain by a silty to clayey sand sequence overlain by about 5 ft of gray-reddish brown clay. The top of the Snapp Formation is a light grayish white poorly sorted medium to coarse grained clayey sand.

Congaree Formation

The base of the Congaree Formation is logged at a depth of 331 ft (-108 ft msl) and the top at a depth of 216 ft (7 ft msl), resulting in a thickness of 115 ft. Although the core was not recovered in one 5-ft run, the base of the Congaree Formation appears to be marked by a 14-ft thick dark grayish black clay. This clay is overlain by black, well sorted silty sand overlain by a 1-ft thick black clay layer. This clay layer is overlain by gray, well sorted clayey sand that grades upward into a light gray well sorted silty sand overlain by a sandy silt. The top of the Congaree Formation consists of gray, well sorted silty sands.

Still Branch Sand

The base of the Still Branch Sand is noted at a depth of 216 ft (7 ft msl) and the top at a depth of 149 ft (74 ft msl), resulting in a thickness of 67 ft. The base of the Still Branch is marked by the occurrence of dark greenish gray sandy silt containing 1- to 3-inch thick sand layers. This silt is overlain by light gray well sorted medium to coarse sand that is overlain by dark gray, well sorted clayey sand. This is overlain by dark, greenish gray calcareous silty sand overlain by calcareous clayey sand. An overlying silty to clayey sand sequence is overlain by a gray well sorted fine to medium sand at the top of the unit.

Lisbon Formation

The Lisbon Formation includes members that have been extensively mapped in the upper Coastal Plain of Georgia and South Carolina. These include the Blue Bluff Marl, McBean Limestone, and, in South Carolina, the Tinker Formation. These units commonly interfinger. In general, the Lisbon Formation is more fossiliferous and ranges from calcareous sands to coquina, while the Tinker Formation in South Carolina is the clastic equivalent consisting predominantly of well sorted quartz sands. The Blue Bluff Marl tends to be more micritic, with shell fragments suspended in a micrite matrix with occasional shell-rich zones. Lithologies, fossil assemblage, and the interfingered nature suggest shallow shelf to neritic environment. In the VEGP site area, the Blue Bluff Marl is noted as the dominant facies and is exposed in the western bluffs along the site boundary with the Savannah River. The Lisbon's Blue Bluff Marl also contains a carbonate limited to the VEGP region called the McBean Limestone (Figures 2.5.1-28 and 2.5.1-38). This unit was extensively studied and mapped as the foundation bearing unit for VEGP Units 1 and 2 and is discussed in Section 2.5.1.2.6.

In boring B-1003, the base of the Lisbon Formation is noted at a depth of 149 ft (74 ft msl) and the top at a depth of 86 ft (137 ft msl), with an overall thickness of 63 ft. The base of the Lisbon Formation is marked by the occurrence of a 12 ft-thick greenish gray, sandy non-plastic silt with fossil fragments and 1- to 3-in thick layers of fossiliferous limestone. This is overlain by a 23-ft thick layer of greenish gray higly plastic sandy silt with fossil fragments and 1- to 3-inch layers of hard fossiliferous limestone. This in turn is overlain by greenish gray well sorted, strongly cemented sand with 1- to 2-inch layers of hard fossiliferous limestone. This is overlain by a light greenish gray fossiliferous limestone. The top of the Lisbon Formation is marked by the occurrence of dark green-gray non-plastic calcareous sandy silt (Blue Bluff Marl). The marl contains 2- to 4-in-thick layers of hard fossiliferous limestone.

Barnwell Group

The Barnwell Group overlying the Blue Bluff Marl member includes four subdivisions, from olderst to youngest, the Clinchfield Formation and Utley Limestone member, Dry Branch Formation, Tobacco Road Sand, and Hawthorne Formation. Due to the location and surface elevation of boring B-1003, the portion of the Barnwell Unit penetrated by boring B-1003 includes only the Clinchfield Formation and lower portion of the Dry Branch Formation. However, most of these units are exposed along the bluffs of the Savannah River or within stream valleys, or lie on topographically higher areas surrounding the site area (Figures 2.5.1-28 and 2.5.1-31).

Clinchfield Formation

Boring B-1003 notes the base of the Clinchfield Formation (Utley Limestone member) at a depth of 86 ft (137 ft msl) and the top at a depth of 48 ft (175 ft msl), with an overall thickness of 38 ft. The thickness of this unit was noted as variable within the excavation exposures, which is consistent with active solutioning of carbonate material.

The Clinchfield Formation consists predominantly of calcareous sands and biomoldic limestones. Some silty and clayey sands are also present, with varying amounts of carbonate material and silicified zones.

Exposures of the Clinchfield Formation along the Savannah River and within excavations for VEGP noted varying degrees of weathering and evidence of solution cavities, indicating that the process of carbonate removal is ongoing. This process could be a primary contributing factor to the development of surface depressions noted in the site area.

Dry Branch Formation

In boring B-1003, the Dry Branch Formation overlying the Clinchfield Formation contains more clayey, laminated sands and silty sands. The base was logged at a depth of 48 ft (175 ft msl) and this unit occurs at the ground surface. The Dry Branch Formation consists primarily of silty, clayey quartz sands. Varying amounts of carbonate material are sometimes present, often in the form of bioherms. The sands are generally moderately to well sorted and subrounded to subangular. Lignite and manganese staining is often present, with a notable lack of glauconite.

Portions of the Dry Branch Formation become significantly more clayey, with finely laminated beds reaching thicknesses of several to tens of feet. The lithology, absence of glauconite, and inclusion of bioherms indicate a back barrier depositional setting.

Tobacco Road Sand

The Tobacco Road Sand consists of moderately to poorly sorted sands and clayey sands with varying amounts of kaolin. Sands tend to be subrounded to rounded, with coarse rounded pebbly zones present as a basal lag in some areas. Where this unit is exposed, the sediments are oxidized and, in many cases, Ophiomorpha burrows are present, as well as cross-beds and convoluted bedding indicative of an open bay, tidal flat setting.

Boring B-1003 did not penetrate this unit due to the surface elevation surrounding the boring location; however, the unit is exposed in stream valleys and road cuts surrounding the site area. The thickness of the Tobacco Road Sand varies due to incision by the overlying Hawthorne Formation but can be in excess of 50 ft. Where present at the VEGP site, the top of the Tobacco Road Sand generally occurs at the ground surface.

Hawthorne Formation

The Hawthorne Formation consists of poorly sorted sands and clayey sands. Sands range from fine to cobble size and are well rounded. Clay is present in the form of laminae to cobble-size clasts. This unit was not identified in any of the borings drilled as part of the ESP subsurface investigation program. It was likely removed during excavation for the existing units and, therefore, is no longer present in the developed portions of the VEGP site. However, it is present in higher elevations of the site area. Obvious incision and cross-cutting channels are noted in exposures, with channel sequences often indicated by coarse channel lags indicative of a high energy fluvial setting. The age of the Hawthorne Formation is problematic due to the lack of fossils. However, Falls and Prowell (2001) indicate a Miocene age for this formation.

Pinehurst Formation

The Pinehurst Formation was not mapped by the USGS, nor is it considered to be a significant stratigraphic unit regionally. The Pinehurst Formation is encountered sporadically within the site area and at the VEGP site. Where preserved, the Pinehurst Formation is less than a few meters thick, and therefore does not appear on geologic maps of the VEGP site. The unit is typically clean, well sorted fine sand. Although bedding is often absent, cross-bedding and remnant dune morphology have been noted in a few exposures, indicative of an eolian deposit **(Prowell 1996)**.

Although the age of the these sediments is not definitive, the unconformable position above the Hawthorne and lack of these sediments overlying Late Pliocene marine sediments downdip of the site area place the relative age as lower Pliocene **(Prowell 1996)**.

Quaternary Stratigraphy

Alluvium exists within the surrounding stream and river valleys and forms terraces that can be locally delineated and mapped. As noted on Figures 2.5.1-29 and 2.5.1-31, in the vicinity of the site area, a modern alluvial flood plain and several alluvial terraces are present on the east side of the Savannah River. The higher terraces show distinctive oxidation and weathering, and the relative position above the Holocene flood plain indicates a Pleistocene age **(Prowell 1996)**.

2.5.1.2.4 Site Area Structural Geology

In the site vicinity, the basement rock beneath the Coastal Plain consists of Paleozoic crystalline rock as well as Triassic–Jurassic sedimentary rock of the Dunbarton Basin. The VEGP site lies near the buried northwest margin of the approximately 9-mi-wide (15-km-wide) Dunbarton Basin, which formed during Mesozoic rifting and opening of the Atlantic Ocean. Deep boreholes within and adjacent to the SRS that penetrate basement indicate that the Paleozoic crystalline rock northwest of the Dunbarton basement has been overprinted with a foliation that dips about 40 to 60 degrees. Based on regional correlations, the foliation strikes northeast and dips to the southeast **(Dennis et al. 2004)**.

The upper surface of the basement has been leveled by erosion and dips to the southeast between 48 ft/mi **(Snipes et al. 1993a)** and 37 ft/mi **(Wyatt 2000)**. In the site area, the regional basement surface is unconformably overlain by loosely consolidated, fluvial, deltaic, and shallow marine Coastal Plain sediments. The depth to the Triassic–Jurassic basement rock beneath the site is 1,049 ft (-826 ft msl), based on borehole B-1003.

Within the 5-mi site area radius, a total of four basement-involved faults have been identified, namely the Pen Branch, Ellenton, Steel Creek, and Upper Three Runs faults. The Ellenton fault does not appear in the most recent SRS fault maps **(Cumbest et al. 1998, 2000)** and, if it exists, is not considered a capable structure (Figure 2.5.1-23). The Upper Three Runs fault is restricted to basement rocks, with no evidence that it offsets Coastal Plain sediments **(Chapman and DiStefano 1989; Stieve and Stephenson 1995)**. Similarly, the Steel Creek fault is not considered a capable tectonic source. The Pen Branch fault is thought to have been the northern bounding (normal) fault of the Mesozoic Dunbarton extensional basin, subsequently reactivated as a reverse or reverse-dextral slip fault, as documented by post-extension, reverse offsets of Late Cretaceous and younger horizons **(Snipes et al. 1993a; Cumbest et al. 1998, 2000)**. The Pen Branch fault is discussed in detail below.

Only one fold potentially of tectonic origin has been identified in the site area. Bechtel site drawings AX6DD377 ("Top of the bearing horizon") and AX6DD378 ("Bottom of the bearing horizon") show an apparent monoclinal flexure of the Blue Bluff Marl, with a hingeline that trends approximately northeast-southwest (Figure 2.5.1-39). Because of its spatial association with the Pen Branch fault, it is likely that this feature is the result of reverse or reverse-oblique slip on the Pen Branch fault. In previous studies [e.g., **(Bechtel 1989)**], this monocline has been referred to as a 3-degree dip reversal and interpreted to be of sedimentalogical origin.

2.5.1.2.4.1 Pen Branch Fault

The Pen Branch fault is neither exposed nor expressed at the surface at the SRS, but its location at the SRS is constrained by dense subsurface well control, as well as seismic-reflection geophysical data [e.g., **(Snipes et al. 1993a; Stieve and Stephenson 1995; Henry 1995; Cumbest et al. 1998, 2000)**]. The Pen Branch exceeds 25 mi in length and is interpreted to comprise several subparallel segments that strike N46-66°E and dip 60-75°SE **(Cumbest et al. 2000)** and project southwestward beneath the VEGP site. Crone and Wheeler (2000) assigned the Pen Branch fault to Class C because of the lack of evidence for post-Eocene slip. Due to its proximity to the VEGP site, however, the Pen Branch fault is examined in detail in this SSAR.

The Pen Branch fault was first discovered in the subsurface of the SRS in 1989 from the interpretation of earlier seismic reflection surveys and other geologic studies **(Marine and Siple 1974; Chapman and DiStefano 1989; Snipes et al. 1993a; Stieve et al. 1994)**. A brief history of the Pen Branch fault and issues concerning the VEGP site and the NRC are as follows:

- January 8, 1989: Draft report on newly discovered Pen Branch fault issued by David Snipes (Clemson University), Wallace Fallaw (Furman University), and Van Price, Jr. (SRS). This report was provided to the NRC, which received it on January 12, 1989. The authors presented evidence of late Eocene movement, but emphasized "compelling evidence for absence of recent movement" and that the fault should not be assumed to be capable. The nearest seismic line data were located about 7 mi east of the VEGP site. This draft report also projected the fault toward the VEGP site based on an interpreted offset of Utley Limestone outcrop located near the existing VEGP intake structure and the dip reversal in the Blue Bluff Marl (Figure 2.5.1-34).

- January 18, 1989: The NRC formally requested the Georgia Power Company to assess capability and impact of the proposed Pen Branch fault on the VEGP site. Southern Company assembled a field review team to review field conditions within and nearby the VEGP site.

- January 29, 1989: Bechtel submitted a response to the NRC.

- February 1989: The NRC issued Supplement No. 8 to the VEGP Units 1 and 2 SER. The NRC concluded that the Pen Branch fault is not capable and that there was no evidence that suggested Tertiary offset on the Pen Branch fault within 6 mi northeast of the VEGP site.
- September 1989: Bechtel issued the Pen Branch fault report **(Bechtel 1989)** that summarized work performed for the January 29, 1989, response to the NRC.
- 1991: A high-resolution shallow seismic reflection survey focused on the uppermost 300 ft of Coastal Plain strata at the SRS conducted by Berkman (1991) was designed to investigate the capability of the Pen Branch fault (Figure 2.5.1-34). Deformation associated with the Pen Branch fault was observed in the Cretaceous Cape Fear Formation but no higher in the stratigraphic section, confirming the non-capable staus of the Pen Branch fault.
- October 1993: Snipes et al. (1993a) paper on the Pen Branch fault was published in *Southeastern Geology*. This paper concluded that the Pen Branch fault must lie more than 2 mi upstream from the VEGP site, based on SRS borehole PBF-6 (which encountered sheared Triassic basin sediments) (Figure 2.5.1-34).
- 1993: Geomatrix (1993) Savannah River fluvial terrace study concluded that the Pen Branch fault has no geomorphic expression, no tectonic deformation is observed within a resolution of approximately 7 to 10 ft, and the Pen Branch fault is not capable.
- 1994: The Confirmatory Drilling Project was designed to investigate the capability of the Pen Branch fault at the SRS **(Stieve et al. 1994)**. This report combined previous data with 18 borings to conclude that deformation associated with the Pen Branch fault likely pre-dates the Williamsburg Unconformity (about 50 Ma) and that the Pen Branch fault is therefore not capable.
- 1995: As part of a groundwater contamination study in Burke County, Georgia, Henry (1995) collected and interpreted a total of 70 mi of high-resolution seismic reflection data from the Savannah River between the Richmond/Burke county line and the Burke/Screven county line. In addition, a medium-resolution seismic survey was conducted in the Savannah River between Hancock Landing and the VEGP boat ramp (Figure 2.5.1-34). Henry (1995) concluded that the Pen Branch fault appears as a high-angle, southeast-side-up reverse fault located ~1,000 ft downstream from Hancock Landing. Henry (1995) interpreted the Pen Branch fault as extending upward through the Paleocene Ellenton Formation and into strata dated as possible Eocene that lie below the unconformity at the base of Savannah River alluvium.
- 1998: As part of an investigation of tritium in the Gordon and other aquifers in Burke County, Georgia, Summerour et al. (1998) reported seismic refletion data collected and interpreted

by Waddell et al. (1995). This land-based seismic reflection survey was located on an unimproved road ~0.5 mi west of River Road (Figure 2.5.1-34). Numerous, minor faults were interpreted to cut reflectors within the Coastal Plain section. The basement reflector, however, is not clearly faulted and, therefore, the interpretation that the Pen Branch fault is imaged in this profile **(Summerour et al. 1998)** is questionable. Based in part on the Waddell et al. (1995) seismic reflection data, Summerour et al. (1998) reported that the Gordon Aquifer is not affected by Pen Branch fault.

- 1998: Cumbest et al. (1998) integrated more than 60 basement borings and 100 mi of seismic reflection profiling to refine the location of the Pen Branch fault at the SRS. Based on their review of existing data, Cumbest et al. (1998) concluded that no faults on the SRS, including the Pen Branch fault, are capable.
- 2000: Based on geometrical analysis of seismic reflection data, the maximum vertical separation of the contact between basement rocks and overlying Coastal Plain sediments by the Pen Branch fault (segment 4) is estimated to be about 92 ft **(Cumbest et al. 2000)**. The offset decreases upward within the Coastal Plain section to 30 ft at the top of the Upper Cretaceous/Lower Paleocene Pee Dee/Ellenton formation (approximately 66.4 Ma) **(Cumbest et al. 2000)**.

2.5.1.2.4.2 Site Subsurface Investigation of the Pen Branch Fault

The Pen Branch fault, which juxtaposes Paleozoic crystalline rock against Triassic (Dunbarton) Basin sedimentary rock at the SRS, has been interpreted to project southwestward into Georgia near the VEGP site. Past interpretations have projected the fault and basin boundary at VEGP site **(Snipes et al. 1989)**, north of the VEGP site **(Snipes et al. 1993a)**, and south of VEGP **(Cumbest et al. 1998; 2000)**. These and all other available data on the location of the fault were compiled and assessed. The study concluded that the Pen Branch fault is located in proximity to the VEGP site.

The seismic acquisition program was designed to image the subsurface structure and characterize the basement lithology and velocities beneath the VEGP site as input to the development of the Safe Shutdown Earthquake (SSE). An additional, specific goal of this study was to image the northeast-striking, non-capable Pen Branch fault, which has been imaged and mapped on the SRS northeast of the VEGP site [e.g., **(Cumbest et al. 2000**)], and determine its precise location, strike, and dip beneath the study region.

Seismic reflection and refraction data were collected on the VEGP site in January and February 2006. The seismic data were acquired by Bay Geophysical under contract to Southern Nuclear Operating Company (SNC); details of the acquisition and preliminary processing of the data are fully documented in the final technical report (Appendix 2.5B). The survey included four seismic reflection and three seismic refraction lines (Figures 2.5.1-35 and Figure 2.5.1-36, respectively). The seismic array was designed to: (1) image the Pen Branch fault, with the assumption that it continues on strike to the southwest from the SRS into the VEGP site area and (2) assess the depth and character of the basement rocks beneath the Atlantic Coastal Plain deposits.

As noted in the report by Bay Geophysical (Appendix 2.5B), vibrations from the existing VEGP generated coherent noise that significantly compromised the quality of the seismic data for lines 1, 2, and 3. The noise masked the first arrivals along most or substantial parts of these three lines, which made it impossible to apply refraction static corrections to the reflection data. Consequently, these lines have anomalies in the reflector geometries that arise from the lack of a proper static solution rather than real earth structure, making detailed geologic interpretation of these lines problematic. Similarly, masking of refractor first breaks by noise from the existing VEGP on lines 1 and 3 made it impossible to confidently and accurately pick the first arrivals for use in 2-D P-wave velocity inversions (Appendix 2.5B). Based on a field assessment of the quality of the refraction data for lines 1 and 3 by Dr. Cumbest, SNC decided not to collect refraction data along line 2, as initially planned.

Reflection and refraction data from line 4 were not affected by noise problems from the existing VEGP. Consequently, line 4 has greater detail than the other three lines in the seismic survey (Figure 2.5.1-37).

<u>Strike of the Pen Branch Fault</u>

The seismic reflection data acquired for this ESP study clearly document that the Pen Branch fault strikes northeast and dips southeast beneath the VEGP site. When the intersections of the Pen Branch fault with the top of basement interpreted from reflection data are plotted on a map, it is apparent that the strike of the fault through the VEGP site is not uniform. The fault-basement intersections on lines 1, 2, and 3 fall along a common trend of about N34°E. In contrast, the fault-basement intersections on lines 4 and 1 define a more westerly trend of about N45°E (Figures 2.5.1-39 and 2.5.1-42). Although it is possible that uncertainty in picking the fault-basement intersection on the eastern three lines (especially lines 1 and 2) may account for some of the difference in strike across the plant site, the different strike east and west of line 2 is likely real for the following reasons:

- The fault and basement offset are best imaged and most confidently interpreted on lines 1 and 4. Thus, the N45°E strike of the fault determined between these two lines is likely accurate.
- The trends of the structure contours on the monoclinally folded Blue Bluff Marl beneath the VEGP site range from about N25°E to N40°E and are similar to the N34°E strike of the fault measured between lines 1, 2, and 3 (Figure 2.5.1-39). Most kinematic models of fault-

related folding assume that the axes of fault-propagation folds and monoclines are parallel to the strike of an underlying thrust or reverse fault [e.g., **(Suppe and Medwedeff 1990; McConnell 1994)**]. If these assumptions are correct, then the strike of the fault beneath the monocline is closer to N32°E, as inferred from seismic lines 1, 2, and 3, than to N45°E.

- Cumbest et al. (2000) have documented that the strike of the Pen Branch fault is not constant beneath the SRS northeast of the VEGP site. They found that the fault consists of several discrete reaches or segments variously separated by small offsets and changes in strike. The range in strike of the fault northeast of Savannah River is about N46°E to N66°E **(Cumbest et al. 2000)**.

The change in strike of the Pen Branch fault across the VEGP site may be part of a regional trend. On the SRS northeast of VEGP, the Pen Branch fault includes a distinct reach that strikes about N58°E between seismic lines SRL-7 and PBF-2A and another distinct reach that strike about N53°E between seismic line PBF-2A and point in the Savannah River channel where the fault was imaged by Henry (1995), representing a counterclockwise rotation in strike from northeast to southwest. Southwest of the Savannah River, the fault strike rotates counterclockwise again to about N34°E beneath the VEGP site. The strike of the fault rotates clockwise to about N45°E between seismic lines 1 and 4. The overall trend is a bend or left jog in the strike of the fault. It is possible that the changes in strike between individual reaches of the fault also include small offsets of the fault plane, as inferred for the fault on the SRS to the north **(Cumbest et al. 2000)**.

Dip of the Pen Branch Fault

The dip of the Pen Branch fault is estimated primarily from its expression in the version of seismic line 4 displayed at a constant velocity of 12,000 ft/s (Figure 2.5.1-37). Based on measurements of P-wave velocities for Triassic basin rocks in boreholes throughout the region **(Chapman and DiStefano 1989)**, 12,000 ft/s probably best characterizes the average velocity of the Triassic rocks in the hanging wall of the Pen Branch fault along line 4. The dip of the Pen Branch fault reflector in the version of line 4 displayed at a velocity of 12,000 ft/s is about 40° (Figure 2.5.1-37).

The geometry of the fault in line 4 reflects an apparent dip because the seismic line is not perpendicular to the fault. The general trend of the section of line 4 corresponding to the well-imaged fault plane (i.e., between shotpoints 225 and 310) intersects the N45°E strike of the fault at an angle of about 55°. Using the apparent dip relation [**(Marshak and Mitra 1988)**, equation 3-7] to account for the obliquity of the seismic line relative to fault strike, a true dip of about 45° to the southeast is derived for the fault plane imaged on line 4 in Figure 2.5.1-40.

Fault-Fold Relationships

The plan projection of the intersection of the Pen Branch fault with the top of basement on lines in the seismic reflection array is located beneath or slightly to the southeast of the antiformal hinge at the top of the monocline in the Blue Bluff Marl (Figure 2.5.1-39). This relationship is shown more directly by two geologic cross sections that pass through borehole B-1003 (Figures 2.5.1-40 and 2.5.1-41).

The northwest-southeast cross section (Figure 2.5.1-40) is oriented perpendicular to the local strike of the Pen Branch fault and displays the fault and its relationship to the monoclinal fold in the Blue Bluff Marl with a minimum of geometric distortion. The plan projection of the offset of the basement surface is located about 100 ft northwest of the upper axial hinge of the monocline (Figure 2.5.1-39). The projection of the fault beyond its termination in the Cretaceous Coastal Plain deposits approximately intersects the synformal hinge at the base of the monocline in the marl (Figure 2.5.1-40). The east-west cross section (Figure 2.5.1-41) is oriented to pass through the locations of boreholes B-1002, B-1003, and B-1004; because the section is oblique rather than perpendicular to the fault and monocline, the geometry of the structures is slightly distorted. In particular, the apparent dip of the fault in the east-west section (35°; Figure 2.5.1-41) is lower than the true dip (about 45°) in the northwest-southeast section (Figure 2.5.1-40). Although the base of the monocline in the marl is poorly constrained by available borehole data at the west end of the east-west cross section, the Pen Branch fault appears to project toward the base of the fold (Figure 2.5.1-40).

2.5.1.2.4.3 Evaluation of Quaternary River Terrace Overlying Pen Branch Fault

The seismic reflection profiles and deep borehole (B-1003) performed at the VEGP site as part of the ESP study helped refine the location of the Pen Branch fault in Georgia and in the westernmost portion of the SRS in South Carolina. These new data combined with reflection profiles in the Savannah River **(Henry 1995)** and earlier SRS studies were integrated to develop a more accurate representation of the fault's location and geometry beneath the VEGP site and Quaternary terraces flanking the Savannah River on the SRS (Figure 2.5.1-34). Previous geomorphic study of the fluvial terraces by Geomatrix (1993) concluded that the Pen Branch fault is not a capable tectonic source and that there is no observable deformation within a resolution of 7 to 10 ft, in the overlying Ellenton Terrace (Qte) surface estimated to be 350 ka to 1 Ma. The 10-ft contour interval of the USGS 7.5-minute topographic maps limited the resolution of this previous study.

Given the higher degree of confidence in the location of the Pen Branch fault beneath the Savannah River fluvial terraces, which represent the only significant Quaternary deposits and surfaces that straddle the Pen Branch fault, a focused study was undertaken to survey and interpret remnants of the Ellenton Terrace (Qte) surface located approximately 4 miles east-northeast of the VEGP site (Figure 2.5.1-9 and Figure 2.5.1-43). The purpose of this effort was to improve the resolution of the terrace surface elevation and independently assess the presence or absence of Quaternary tectonic deformation on the Pen Branch fault, which has been classified by Crone and Wheeler (2000) as a potential Quaternary fault having insufficient geologic evidence to demonstrate Quaternary slip or deformation (Class C in Table 2.5.1-1).

The scope of this investigation included a review of previous studies, as well as geomorphic mapping, analysis of aerial photographs, surveying the portion of the Qte terrace surface at the SRS that overlies the Pen Branch fault, and the construction and analysis of longitudinal terrace profiles.

Geomorphic mapping and field reconnaissance

Prior to surveying the study area to acquire elevation data, a geomorphic map of the Qte terrace surface at the SRS in the vicinity of the Pen Branch fault was prepared in order to establish geologic context, to ground-truth and refine the mapping of Geomatrix (1993), and to investigate the degree of erosion and/or anthropogenic disturbance of the terrace surface. The primary focus of the geomorphic mapping was to define the portions of the Qte surface that appear best preserved (minimal deflation and modification) (Figure 2.5.1-44). Preparation of this map included inspection of aerial photographs dating from 1943 to 2004, field reconnaissance, and hand-auger soil borings. A review of multiple sets of aerial photography reveal the presence of several closed depressions (of variable size), marshy areas, and tributary drainages in the study area (Figure 2.5.1-44). The aerial photographs also show that that the study area was farmland at least as early as 1943. Today the study area is crossed by an SRS power line right-of-way and several dirt roads, and timber has been sporadically logged from portions of the study area over the past few decades.

Survey data acquisition

A total of approximately 2,600 elevation data points were collected by a Georgia Power Company survey team using a combination of differential Global Positioning System (DGPS) and total station survey techniques. Survey data were collected in UTM zone 17N coordinates, using NAD 27 geographic datum and NAVD 88 elevation datum.

The preponderance of survey data was collected from, and adjacent to, dirt roads and the power line right-of-way (Figures 2.5.1-43 and 2.5.1-44). The Qte terrace surface in the vicinity of the Pen Branch fault on the SRS is largely covered by trees and dense undergrowth. This dense vegetative cover hinders the acquisition of survey data. Fortunately, a power line right-of-way oriented approximately normal to the local strike of the Pen Branch fault and approximately parallel to the long-axis of the Qte terrace deposits extends through the study area. As much as possible, survey data were collected away from obviously disturbed, eroded, and/or modified areas as identified by geologic reconnaissance and aerial photograph interpretation.

Terrace surface longitudinal profiles

A longitudinal profile of the Qte terrace surface was constructed by projecting elevation data onto a profile line oriented approximately N35°W (approximately normal to the local strike of the Pen Branch fault and approximately parallel to the long-axis of the Qte terrace and paleo-Savannah River valley in the study area) (Figures 2.5.1-43 and 2.5.1-44). Data points interpreted as representing the best-preserved remnants of the Qte terrace surface (as determined from field reconnaissance and aerial photograph inspection) are shown on the longitudinal profile in red (Figure 2.5.1-45). Data points interpreted as representing more modified or eroded portions of Qte terrace surface and those points collected away from the Qte terrace surface are shown on the longitudinal profile in gray.

Sources of error/uncertainty

There are three main sources of uncertainty and error that contribute to the overall uncertainty in the original elevation and variability of the Qte terrace surface. Each of these is discussed below:

Geologic context – The largest contributor to the overall uncertainty in characterizing the Qte terrace surface elevation is due to ambiguities regarding geologic and geomorphic context. The Qte terrace surface was initially deposited as a planar feature with some inherent variability. Since the river abandoned this surface 350 ka to 1 Ma, however, the Qte surface has been modified by geologic and anthropogenic processes. Geologic processes of deposition, erosion, and settlement resulting from dissolution of the underlying carbonate sands have all contributed to the modification the original surface. Significant deflation of the surface has occurred as a result of dissolution collapse as evidenced from the abundant closed depressions across the Qte and younger Qtb terrace surfaces (Figures 2.5.1-29 and 2.5.1-45). Incision of tributary drainages into the terrace have locally removed the deposits and surface as well as the deflated adjacent portions of the surface. Deposition of alluvial and colluvial material onto the eastern margin of the Qte terrace near the mouth of Fourmile Branch and along the base of the southwest-facing slopes has locally increased the ground surface elevation. Anthropogenic processes that have modified the original terrace surface, to a much lessor extent, include agricultural land use practices (logging and farming) and other human activities related to development of the SRS. A primary focus of this study was to define those portions of the Qte terrace surface that best preserved remnants of the original terrace (least modified) and to resolve the magnitude of the scatter in elevation data. Any remaining perturbations, if any, in the overall Qte terrace surface that cannot be explained by erosion, settlement, deposition, and/or anthropogenic modification can be considered to be of possible tectonic origin.

Survey error – The uncertainty due to both systematic and random errors associated with the collection and processing of survey data is estimated to be about 3 cm (about 1.2 inches) in the horizontal dimension and about 5 cm (about 2 inches) in the vertical dimension. The contribution of survey error to overall uncertainty is considered to be negligible.

Profile construction error – The projection of elevation data onto profile line A-A' (Figure 2.5.1-45) introduces a minimal amount of error into the analysis. The magnitude of the positional error resulting from profile construction is minimized by minimizing the distance over which points are projected, and by constructing the profile approximately parallel to the long-axis of the Qte terrace surface in the study area and approximately normal to the local strike of the Pen Branch fault. The amount of error introduced into this analysis from the construction of the longitudinal profile is minimal.

<u>Results</u>

The geomorphic map presented in Figure 2.5.1-44 shows the best-preserved remnants of the Qte terrace surface in the study area (red shaded areas). The influence of dissolution collapse-related depressions on the terrace surface is most clearly seen in the vicinity of depressions D1 and D2, as short-wavelength variations in the topographic surface. In addition, portions of the Qte terrace surface have been sites for the local deposition of alluvium and colluvium. These two areas are located at the northwestern-most extent of the survey and southeast of depession D2 (Figure 2.5.1-45).

Taken together, the overall uncertainty in the elevation of the best-preserved remnants of the Qte terrace surface is estimated to be about 3 ft. As shown in Figure 2.5.1-45, the elevation data for the terrace remnant range between elevations of 153 and 156 ft.

Longitudinal profile A-A' (Figure 2.5.1-45) indicates about 25 ft of variability in the present topography of the Qte terrace deposit in the study area. Most of this variability is the result of erosion and deflation of the terrace surface.

A longitudinal profile of the Qte fluvial terrace surface in the study area provides evidence demonstrating the absence of discernible tectonic deformation due to the underlying Pen Branch fault within the limit of resolution of the terrace elevation data (Figure 2.5.1-45). The results of this study demonstrate a lack of tectonic deformation in the 350 ka to 1 Ma year old fluvial terrace surface within a resolution of about 3 ft. This observation is consistent with previous studies at both the VEGP site and the SRS that have concluded the Pen Branch fault is not a capable tectonic source.

2.5.1.2.5 Site Area Geologic Hazard Evaluation

No geologic hazards have been identified within the VEGP site area. Surface depressions associated with dissolution of carbonate bearing stratum of the Utley member of the Clinchfield Formation are discussed in Sections 2.5.1.2.3.2 and 2.5.3.8.2.1 and do not affect the foundation-bearing layer (Blue Bluff Marl). However, structures founded above the Blue Bluff Marl will require subsurface exploration to identify low-bearing-strength layers associated with dissolution processes noted in the site area.

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 layer will be evaluated and mapped as the excavation is completed.

Heave monitor installation will be required prior to excavation, and settlement monitoring will be required during and post construction. Heave measurements will be used to quantify recompression during reloading to measure actual net settlements.

2.5.1.2.6.2 Zones of Alteration, Weathering, and Structural Weakness

The Blue Bluff Marl will form the foundation-bearing layer and consists of unweathered, slightly lithified, micritic limestone. Some desiccation is expected; however, visual examination of the exposure will be required, and long-term exposure may require a thin application of shotcrete for protection. Any noted desiccation, weathered zones, joints, or fractures will be mapped and evaluated.

2.5.1.2.6.3 Deformational Zones

No deformational zones within the Blue Bluff Marl were reported from the detailed excavation mapping for VEGP. However, proximity of the Pen Branch fault to the VEGP site (Figures 2.5.1-34 and 2.5.1-42) may have produced deformational features during development of the anticlinal structure described in Section 2.5.1.2.4.1. However, these features, if present, are not expected to compromise the foundation-bearing capabilities. Excavation mapping will be required during construction, and any noted deformational zones will be evaluated.

2.5.1.2.6.4 Prior Earthquake Effects

Extensive studies of outcrops and alluvial terrace and flood plain deposits have not indicated any evidence for post-Miocene earthquake activity, as discussed in Section 2.5.1.2.4.

2.5.1.2.6.5 Effects of Human Activities

No mining operations (other than borrow of surficial soils), excessive extraction or injection of groundwater, or impoundment of water has occurred within the site area that can affect geologic conditions.

2.5.1.2.7 Site Groundwater Conditions

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

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

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.

Source: Crone and Wheeler 2000; Wheeler 2005